# aquifers of west texas

## Report 356

edited by
Robert E. Mace
William F. Mullican III
Edward S. Angle

## Texas Water Development Board

P.O. Box 13231, Capitol Station
Austin, Texas 78711-3231

**December 2001**

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# Texas Water Development Board

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**Craig D. Pedersen, Executive Administrator**

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# Preface

On behalf of the Texas Water Development Board, I want to welcome you to Alpine for the Aquifers of West Texas Conference. The Texas Water Development Board, along with our co-sponsors, Sul Ross State University and the University of Texas at El Paso, Center for Environmental Resource Management, hope that you will benefit greatly from the presentations made at the conference and during the field trip, as well as from the technical papers compiled in this document, Report 356, Aquifers of West Texas.

As a West Texas landowner, I keenly recognize the critical value of the precious water resources in our beautiful, productive, but arid lands. One of the most positive aspects of Senate Bill 1, passed by the 1997 Texas Legislature, was the regional water planning process. This process significantly improved our understanding of our water resources and their availability to meet future needs. There is still, however, so much more to be known about the hydrogeology of our West Texas aquifers. The Aquifers of West Texas Conference is, in large part, a compilation of much of the hydrogeologic information available regarding our groundwater resources. I believe that the valuable hydrogeologic information included in Report 356, along with the technical interaction and exchange of ideas to occur throughout this conference, will have a positive impact on our understanding of the water resources of Texas for many years to come.

Increasing the hydrogeologic science about our West Texas aquifers and enhancing public understanding about these underground water resources are vital to the future policy decision made about use of this water.

On behalf of the citizens of West Texas, we thank you for your participation in this most important effort.

Kathleen Hartnett White
Board Member
Texas Water Development Board

# Note from the Editors:

The start of regional water planning, the pressures of drought, and the challenges facing growing urban centers have catalyzed interest in the water resources of West Texas. Therefore, when one of our board members, Ms. Kathleen Hartnett White, approached staff in late 2000 with the concept of holding a conference focusing on the science of aquifers in West Texas, we thought it was a great idea. Many water-resource issues in West Texas are controversial, and reliable scientific information is needed to help address many of the issues. Therefore, the purpose of the conference was to (1) review what is known about the aquifers of West Texas and (2) identify what needs to be done to better understand the aquifers.

When we organized the conference, we first identified the topics we wanted addressed and then identified potential speakers to invite to discuss each of the topics. After preparing an outline of the topics, we realized we had the skeleton of a good book about the aquifers of West Texas. Therefore, we asked speakers to also prepare a chapter to include in the volume you are now holding. This volume is meant to be a stand-alone document as well as a proceedings of the conference held in Alpine December 4th through 6th, 2001, including a field trip.

Orchestrating the conference and this document was a great task, and we are thankful for the assistance of many people. First, we thank our co-conveners for the conference, Sul Ross State University (David Rohr and Kevin Urbanczyk, Department of Earth and Physical Sciences) and the Center for Environmental Resource Management at The University of Texas at El Paso (Ed Hamlyn). We are also thankful to Barbara Kauffman (Rio Grande Council of Governments), Janet Adams (Jeff Davis County Underground Water Conservation District), Kate Hoskins (Culberson County Groundwater Conservation District), and Carla Daws (TWDB) for publicizing the conference.

We thank the authors for sharing their time and knowledge in preparing these papers. We are particularly thankful to Ian Jones, Sanjeev Kalaswad, and Zhuping Sheng for producing their papers with short notice. We thank the groundwater conservation districts for participating in the conference. We are grateful to Lana Dieterich of the Bureau of Economic Geology for her review of the document and Mike McCathern and Zelphia Bloodworth for final formatting and production work. We also thank our executive administrator, Craig Pedersen, and our deputy executive administrator, Dr. Tommy Knowles, for their reviews and support. We also thank our board member, Ms. Kathleen Hartnett White, for the initial idea and support throughout this effort.

Robert E. Mace
William F. Mullican III
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# Chapter 1

# Aquifers of West Texas: An Overview

Robert E. Mace[1]

## Introduction

Water is essential to the social, economic, and environmental well-being of Texas, especially in the arid areas of the west part of the state, where water is scarce and highly valued. Drought and increasing demand, primarily because of a growing urban population, have heightened concerns over water resources in the area. The cities of Ciudad Juarez, Mexico, and El Paso, USA, are quickly depleting their fresh groundwater resources. Juarez, it is estimated, will pump the last of its fresh groundwater beneath the city by 2005, and El Paso by 2020 (Washington and Perez, 2001). International and State boundaries complicate water policy in the area. Pumping in Juarez has drawn groundwater flow from Texas into Mexico, and pumping in Juarez and El Paso has drawn groundwater flow from New Mexico into Texas.

As local water resources and options decline, large urban areas such as El Paso and Ciudad Juarez are looking elsewhere for water. El Paso is considering desalination of local, poorer quality water as another possible water source. El Paso is also looking north and east for other groundwater resources (Brown and Caldwell, 2001a,b,c; FWTPG, 2001; HBC, 2001). Continuing and increasing urban demands can affect other elements of the local economy as well, such as agriculture and ranching, water supplies to smaller communities, and flow to springs that may harbor endangered and threatened species or have aesthetic and recreational value.

Although West Texas has not been graced by many major aquifers, it is home to many minor aquifers of varying water quality, yields, and geology. El Paso and its surrounding urban area currently rely on groundwater for about half of its water supply, and Ciudad Juarez relies entirely on groundwater from the Hueco Bolson aquifer. All of the towns and rural areas in West Texas rely entirely on groundwater, and total groundwater usage in the area has ranged from 320,000 to 720,000 acre-ft/yr over the past 20 yr. A better understanding of these aquifers is important for us to know how to best manage the scarce water resources that do exist in West Texas. The purpose of this paper is to present a general overview of the aquifers of West Texas and recent scientific and planning activities concerning these aquifers. Additional chapters in this report discuss the aquifers and some of the issues in far greater detail.

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[1] Texas Water Development Board

# Location, Physiography, and Climate

We focus on the part of West Texas that includes Brewster, Culberson, El Paso, Hudspeth, Jeff Davis, Loving, Pecos, Reeves, Ward, and Winkler Counties (fig. 1-1). All of the counties in the area have fewer than 16,000 people, with the exception of El Paso County, which had about 680,000 people in 1997 (table 1-1). The population in the area has grown by nearly 500,000 people since 1950, with 98 percent of that growth occurring in El Paso County (table 1-1). In 1997, Brewster, Culberson, Pecos, Reeves, and Winkler Counties received more than 90 percent of their water from aquifers (table 1-1).

West Texas is primarily located in the Basin and Range Physiographic Province (Thornbury, 1965), which is characterized by asymmetric ridges or mountains and broad intervening basins (Bates and Jackson, 1984). Basins in the area have land-surface elevations of about 3,000 ft, with mountain ranges rising several thousand feet higher. Mountain ranges in the area include the Guadalupe Mountains (8,751 ft), Eagle Mountains (7,484 ft), the Quitman Mountains (5,200 ft), the Carrizo Mountains (5,200 ft), the Sierra Blanca Peaks (6,800 ft), the Davis Mountains (8,206 ft), and the Chisos Mountains (7,825 ft). The Diablo Plateau lies in the north–to-central part of Hudspeth County.

The Rio Grande and Pecos River are the major rivers than cut through the West Texas area (fig. 1-1). Upstream of El Paso, flow in the Rio Grande is primarily controlled by releases from Caballo Reservoir, located downstream of Elephant Butte. Downstream of El Paso, flow in the river consists of treated municipal wastewater from El Paso, untreated municipal wastewater from Ciudad Juarez, irrigation return-flow, and occasional floodwater and runoff. So much water leaks from the river into the ground that the river is often dry between southern Hudspeth County and Presidio, where Rio Conchos joins the Rio Grande.

The Pecos River is a major tributary to the Rio Grande that originates in New Mexico. The river is impounded in Red Bluff Lake in Loving County and is used for irrigation in Pecos, Reeves, and Ward Counties.

Most of the study area is in the mountain and subtropical arid climate regions of Texas (Larkin and Bomar, 1983) and lies in the north part of the Chihuahuan Desert (Schmidt, 1979). The Guadalupe, Davis, and Chisos Mountains of the Trans-Pecos region of Texas are in the mountain climate region, characterized by cooler temperatures, lower relative humidity, and moderate amounts of irregular rainfall. The rest of the study area is primarily in the subtropical arid climate region, influenced by the flow of air from the Gulf of Mexico that is disturbed by intermittent seasonal intrusions of continental air.

West Texas is the most arid region of the state, and, because of its low rainfall and high evaporation, is in drought during all or part of most years (Bomar, 1995). Average annual precipitation ranges from 8 inches in the El Paso area to more than 18 inches in the Davis Mountains (fig. 1-2a). In general, mountainous areas receive more rainfall than the surrounding valleys. Average annual gross lake-surface evaporation rates range from less

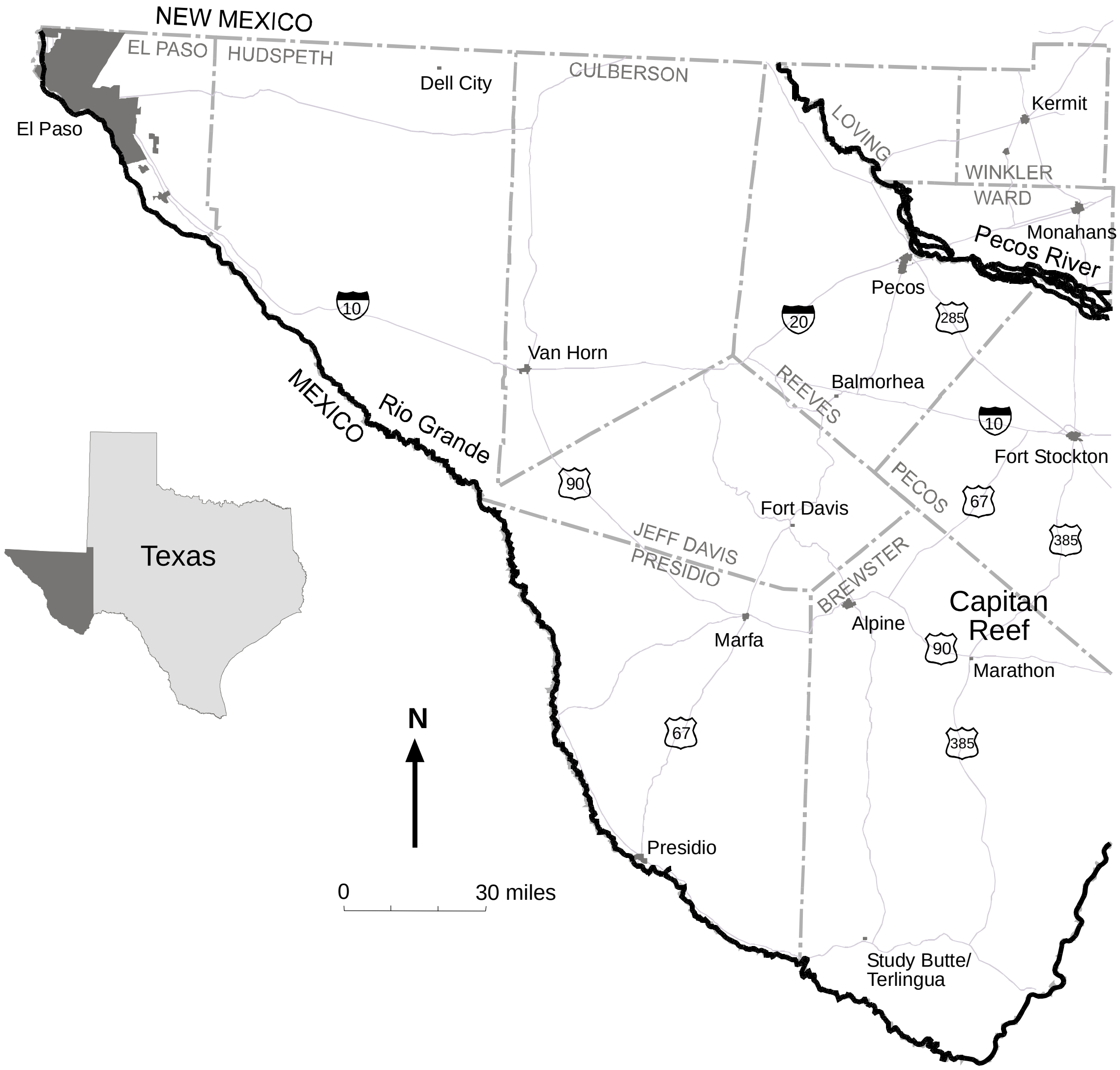


Figure 1-1: Location of the study area in West Texas.

than 80 inches in Loving and Winkler Counties to over 95 inches in southern Presidio County near the Rio Grande (fig. 1-2b).

# Aquifers of West Texas

The West Texas area includes all or parts of 12 aquifers recognized by the State (fig. 1-3). Three major aquifers, the Hueco-Mesilla Bolson, the Cenozoic Pecos Alluvium, and the Edwards-Trinity (Plateau), are found in the area. Nine minor aquifers are also located in the area, including the Bone Spring-Victorio Peak, Capitan Reef, Dockum, Igneous, Marathon, Rustler, and West Texas Bolsons aquifers. The Texas Water Development Board (TWDB) assigns a major and minor status to the state's aquifers on the basis of the quantity of water supplied by each aquifer (Ashworth and Hopkins, 1995). In addition to

Table 1-1: Population and groundwater use in Far West Texas counties for selected years.

| County | Population 1950 | Population 1980 | Population 1990 | Population 1997 | Groundwater use (acre-ft) 1980 | Groundwater use (acre-ft) 1990 | Groundwater use (acre-ft) 1997 | %GW |
|---|---|---|---|---|---|---|---|---|
| Brewster | 7,309 | 7,573 | 8,681 | 9,279 | 3,126 | 2,551 | 3,664 | 93.0 |
| Culberson | 1,825 | 3,315 | 3,407 | 3,299 | 76,119 | 12,580 | 9,773 | 99.9 |
| El Paso | 194,968 | 479,899 | 591,610 | 683,657 | 99,923 | 118,330 | 97,734 | 36.6 |
| Hudspeth | 4,298 | 2,728 | 2,915 | 3,397 | 141,649 | 51,526 | 132,327 | 66.6 |
| Jeff Davis | 2,090 | 1,647 | 1,946 | 2,028 | 26,872 | 3,767 | 898 | 85.7 |
| Loving | 227 | 91 | 107 | 95 | 64 | 44 | 70 | 10.5 |
| Pecos | 9,939 | 14,618 | 14,675 | 15,883 | 111,250 | 67,552 | 82,865 | 96.6 |
| Presidio | 7,354 | 5,188 | 6,637 | 7,484 | 14,200 | 7,027 | 4,977 | 19.0 |
| Reeves | 11,745 | 15,801 | 15,852 | 15,329 | 120,524 | 40,117 | 106,136 | 91.5 |
| Ward | 13,346 | 13,976 | 13,115 | 12,797 | 33,311 | 10,670 | 10,821 | 55.8 |
| Winkler | 10,064 | 9,944 | 8,626 | 8,335 | 8,356 | 3,171 | 3,647 | 99.9 |
| **Total:** | 263,165 | 554,780 | 667,571 | 761,583 | 635,394 | 317,335 | 452,912 | 62.3 |

**% GW** = percent of total water use in 1997 that was met with groundwater.
Population and groundwater use for all of Brewster and Pecos Counties are included.

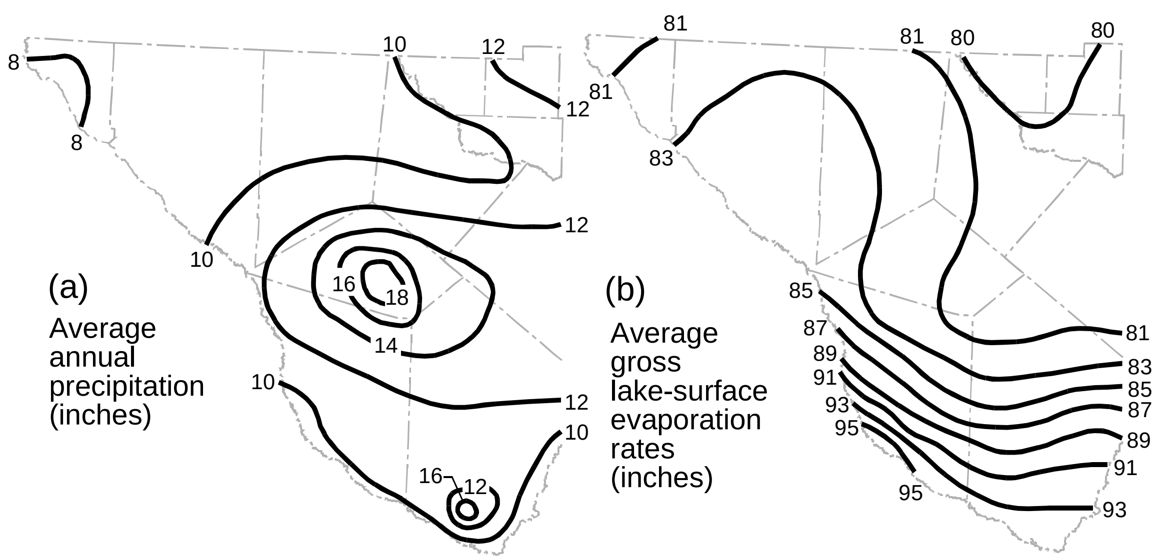


Figure 1-2: Amount of (a) average annual precipitation and (b) average gross lake-surface evaporation in the Far West Texas area (after Larkin and Bomar, 1983).

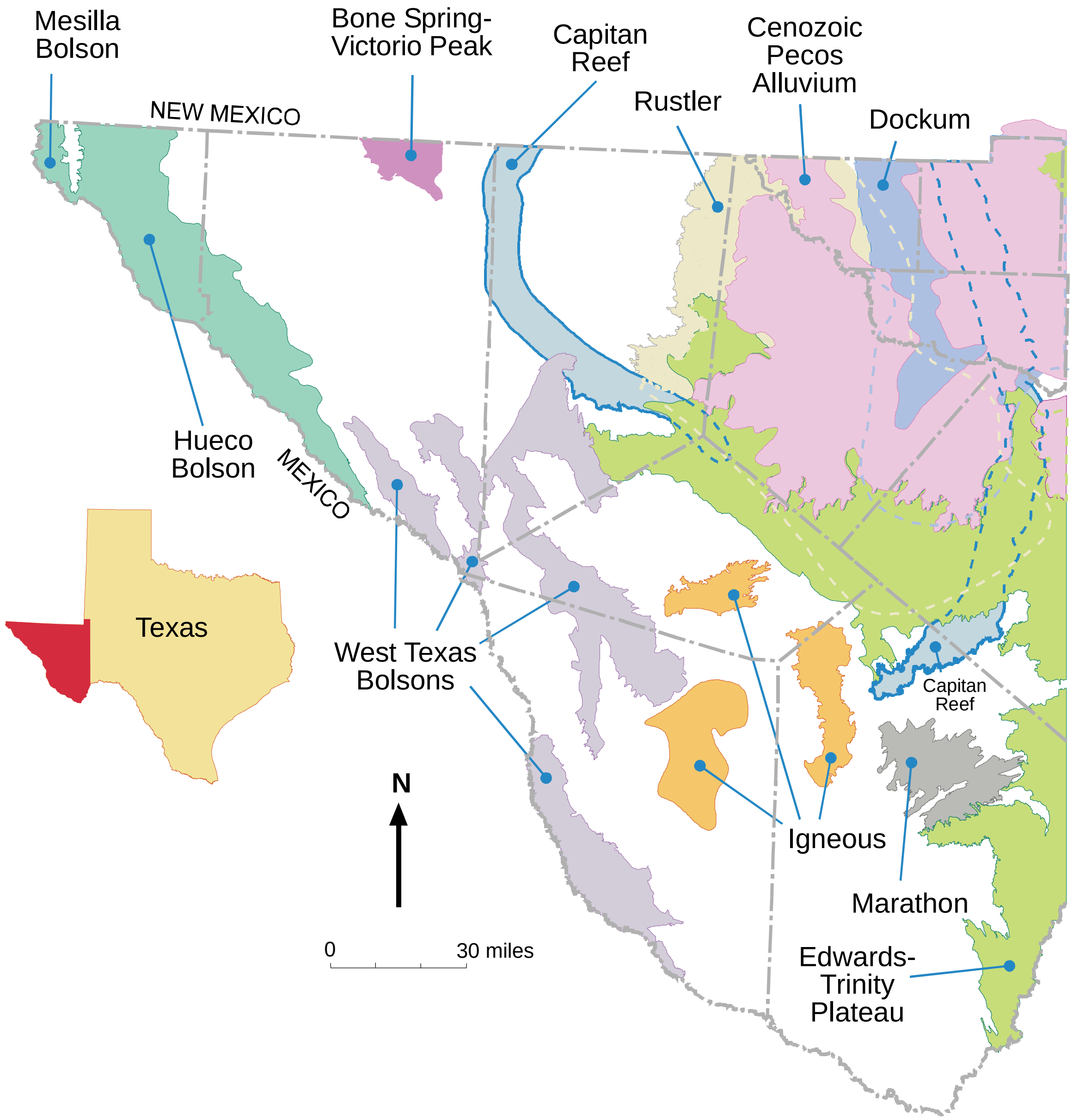


Figure 1-3: Location of recognized aquifers in Far West Texas.

the aquifers recognized by the TWDB, there are several other geologic formations that locally produce water.

Several of the aquifers (Hueco-Mesilla Bolson, the Cenozoic Pecos Alluvium, the Edwards-Trinity [Plateau], Bone Spring-Victorio Peak) have had a number of scientific studies done on them, especially the Hueco-Mesilla Bolson aquifer. However, several others (Capitan Reef, Dockum, Igneous, Marathon, Rustler, and West Texas Bolsons) have had few to almost no groundwater studies done on them.

Although we show the aquifers as separate entities, many are hydraulically connected to each other. For example, some of the West Texas Bolsons are connected to the Igneous and Capitan Reef aquifers (Angle, this volume; Brown and Caldwell, 2001c; Chastain-Howley, this volume). Brown and Caldwell (2001c) showed that much of the water produced from wells in the Ryan Flat Bolson aquifer at Antelope Valley Farm is sourced from igneous rocks underlying the bolson deposits. The Cenozoic Pecos Alluvium, Edwards-Trinity (Plateau), Dockum, Rustler, and Capitan Reef aquifers also hydraulically intermingle with each other in different areas. Fault systems and other flow paths can allow groundwater to move across areas without recognized aquifers (Sharp, this volume). Flow systems in West Texas can be very complex.

Much of the general information presented next is from Aquifers of Texas (Ashworth and Hopkins, 1995), the Far West Texas Regional Water Plan (FWTPG, 2001), water-use information from TWDB surveys and estimates, and selected publications.

## Hueco-Mesilla Bolson aquifer

The Hueco-Mesilla Bolson aquifer consists of two bolsons: the Hueco and the Mesilla Bolsons (fig. 1-3). The Hueco Bolson is in El Paso and Hudspeth Counties, Texas, extends into Mexico south of the Rio Grande, and extends north of El Paso County, Texas, into New Mexico (fig. 1-3). A small part of the Mesilla Bolson extends into El Paso County (fig. 1-3), with most of the aquifer in New Mexico to the north. The Hueco Bolson is about 9,000 ft thick and consists of silt, sand, and gravel in the upper part and silt and clay in the lower part. The Mesilla Bolson is about 2,000 ft thick and consists of clay, silt, sand, and gravel. Pumping by El Paso and Ciudad Juarez have caused large water-level declines, changing groundwater flow directions, flow rates, and water quality and causing a minor amount of land subsidence. Pumping from the aquifer in Texas over the past 20 yr has ranged from about 96,000 to about 138,000 acre-ft/yr (table 1-2). The Hueco and Mesilla Bolsons are discussed in more detail by Sheng and others in chapter 6 and by Hawley and others in chapter 7, respectively.

## Edwards-Trinity (Plateau) aquifer

The Edwards-Trinity (Plateau) aquifer, in the east part of the area in Brewster, Culberson, Jeff Davis, Pecos, Reeves, and Winkler Counties (fig. 1-3), extends eastward to the Hill Country of Texas. The Edwards-Trinity (Plateau) aquifer consists of rocks of the Comanche Peak, Edwards, and Georgetown Formations and the Trinity Group. The Trinity Group consists primarily of sands (Antlers and Maxim sands) and limestones. The Comanche Peak, Edwards, and Georgetown Formations consist primarily of limestones and dolomites. Pumping from the aquifer in the counties in the study area over the past 20 yr has ranged from about 52,000 to about 95,000 acre-ft/yr (table 1-2). The Edwards-Trinity (Plateau) aquifer is discussed in more detail by Anaya in chapter 8.

Table 1-2: Groundwater use for the different aquifers in the Far West Texas area (acre-ft).

| | Year | | | | | | |
|---|---|---|---|---|---|---|---|
| **Aquifer** | **1980<br>1990<br>1997** | **1984<br>1991** | **1985<br>1992** | **1986<br>1993** | **1987<br>1994** | **1988<br>1995** | **1989<br>1996** |
| Bone Spring-Victorio Peak | 132,891 | 100,667 | 91,757 | 42,803 | 46,316 | 52,749 | 92,364 |
| | 48,091 | 49,719 | 38,452 | 113,041 | 173,046 | 137,625 | 128,964 |
| | 129,592 | | | | | | |
| Capitan Reef | 15,264 | 952 | 844 | 165 | 809 | 791 | 811 |
| | 690 | 559 | 438 | 145 | 2,832 | 2,257 | 2,118 |
| | 2,129 | | | | | | |
| Cenozoic Pecos Alluvium | 196,423 | 120,469 | 99,691 | 91,153 | 72,175 | 75,479 | 103,674 |
| | 68,414 | 66,193 | 62,452 | 386,754 | 149,972 | 157,070 | 147,845 |
| | 147,711 | | | | | | |
| Dockum | 5,336 | 5,488 | 5,178 | 3,866 | 3,657 | 4,003 | 3,857 |
| | 3,547 | 3,871 | 5,226 | 5,900 | 5,598 | 4,582 | 4,445 |
| | 4,411 | | | | | | |
| Edwards-Trinity | 74,085 | 91,346 | 77,047 | 64,836 | 58,845 | 58,572 | 64,257 |
| | 59,265 | 56,432 | 55,844 | 95,457 | 52,024 | 58,926 | 55,009 |
| | 55,821 | | | | | | |
| Hueco-Mesilla Bolson | 103,952 | 114,176 | 113,301 | 114,274 | 125,636 | 124,353 | 138,203 |
| | 121,518 | 111,200 | 108,631 | 104,540 | 97,257 | 96,556 | 103,505 |
| | 95,633 | | | | | | |
| Igneous | 6,826 | 3,953 | 4,010 | 3,915 | 3,154 | 3,863 | 3,435 |
| | 3,338 | 3,582 | 3,618 | 3,620 | 4,291 | 4,266 | 4,167 |
| | 4,821 | | | | | | |
| Marathon | 119 | 85 | 78 | 80 | 80 | 90 | 92 |
| | 98 | 100 | 93 | 120 | 117 | 121 | 126 |
| | 130 | | | | | | |
| Rustler | 286 | 351 | 252 | 235 | 220 | 175 | 207 |
| | 173 | 168 | 183 | 598 | 1,405 | 1,593 | 1,491 |
| | 1,532 | | | | | | |
| West Texas Bolsons | 91,033 | 25,363 | 28,093 | 22,566 | 21,263 | 24,569 | 20,418 |
| | 17,538 | 12,985 | 17,019 | 10,568 | 10,957 | 11,330 | 11,625 |
| | 12,839 | | | | | | |
| Other aquifers | 12,063 | 5,347 | 5,117 | 3,841 | 4,868 | 4,690 | 4,440 |
| | 3,374 | 3,323 | 2,722 | 1,937 | 1,202 | 1,028 | 913 |
| | 890 | | | | | | |

## Cenozoic Pecos Alluvium aquifer

The Cenozoic Pecos Alluvium aquifer, located in Jeff Davis, Loving, Pecos, Reeves, Ward, and Winkler Counties in West Texas (fig. 1-3), extends to the east in Texas and to the north into New Mexico. The aquifer consists of sands, gravels, and clays of ancient river deposits that can be up to 1,500 ft thick. The aquifer is connected to the Dockum and Edwards-Trinity (Plateau) aquifers where they exist underneath the alluvium. Water quality is naturally highly variable and has also been locally impacted by past activities of the petroleum industry. Water levels have declined more than 200 ft in south-central Reeves and northwest Pecos Counties but have remained somewhat steady since the 1970's, with a decrease in irrigation. Lowered water levels have decreased base flow to the Pecos River and, in some cases, now cause the river to lose water to the aquifer. Pumping from the aquifer over the past 20 yr has ranged from about 62,000 to about 138,000 acre-ft/yr (table 1-2). Reeves County has been the largest user of groundwater from the aquifer, using 67 percent of the total water pumped in 1997. The Cenozoic Pecos Alluvium aquifer is discussed in more detail by Jones in chapter 9.

## Bone Spring-Victorio Peak aquifer

The Bone-Spring Victorio Peak aquifer, located in Hudspeth County (fig. 1-3), extends northward into the Crow Flats area of New Mexico. The aquifer consists of about 2,000 ft of limestone beds of the Bone Spring and Victorio Peak Formations, with water occurring in fractures and solution cavities. The aquifer is primarily used for irrigation, although Dell City relies on the aquifer for municipal supply. Water levels have historically declined in the aquifer but have remained relatively steady since the late 1970's. Pumping from the aquifer in Texas over the past 20 yr has ranged from about 38,000 to about 170,000 acre-ft/yr (table 1-2). The Bone-Spring Victorio Peak aquifer is discussed in more detail by Ashworth in chapter 10.

## Capitan Reef aquifer

The Capitan Reef aquifer consists of two strips located in Culberson, Hudspeth, Jeff Davis, Pecos, Reeves, Ward, and Winkler Counties (fig. 1-3) and extends northward into New Mexico. The aquifer is an ancient reef consisting of 2,360 ft of dolomite and limestone, and, in Texas, generally has poor water quality except in the exposed areas of the aquifer. Most of the water pumped from the aquifer is in Ward and Winkler Counties for water-flooding operations in oil-producing areas. A small amount of water is used for irrigation in Pecos and Culberson Counties. Carlsbad, New Mexico, relies on the aquifer for municipal use. Pumping from the aquifer in Texas over the past 20 yr has ranged from about 150 to about 15,000 acre-ft/yr (table 1-2). Recent pumping has been about 2,100 acre-ft/yr. The Capitan Reef aquifer is discussed in more detail by Uliana in chapter 11.

## Dockum aquifer

The Dockum aquifer, located in Loving, Pecos, Reeves, Ward, and Winkler Counties in West Texas (fig. 1-3), extends to the east and northeast beneath the Ogallala and Edwards-Trinity (Plateau) aquifers and to the north into New Mexico. The Dockum aquifer consists of up to 700 ft of sand and conglomerate, with layers of silt and shale of the Dockum Group. Water quality is variable and is used for water-flooding operations in oil-producing areas of the southern High Plains. Pumping from the aquifer in the counties in the study area over the past 20 yr has ranged from about 3,900 to about 5,900 acre-ft/yr (table 1-2). The Dockum aquifer is discussed in more detail by Bradley and Kalaswad in chapter 12.

## Igneous aquifer

The Igneous aquifer is currently represented on TWDB maps in three separate pieces in Brewster, Jeff Davis, and Presidio Counties near Alpine, Fort Davis, and Marfa, respectively (fig. 1-3). Recent work by the Far West Texas Planning Group, included in part in chapter 13 and summarized by LBG-Guyton (2001), suggests that the aquifer has a much greater extent coincident with the general occurrence of igneous (or volcanic) rocks in the area. Groundwater in the Igneous aquifer occurs primarily in the fractures of tuffs and other volcanic rocks in the aquifer, with thicknesses of about 900 to 1,000 ft. Alpine, Fort Davis, and Marfa rely on the aquifer as a source of municipal water. Pumping from the aquifer over the past 20 yr has ranged from about 3,100 to about 6,800 acre-ft/yr (table 1-2). The Igneous aquifer is discussed in more detail by Chastain-Howley in chapter 13.

## Marathon aquifer

The Marathon aquifer is located in North-Central Brewster County in the vicinity of Marathon (fig. 1-3). Groundwater occurs in fractures and solution cavities at depths between 350 and 900 ft. Many shallow wells in the area produce from alluvial deposits that overlie the Marathon aquifer. Pumping from the aquifer over the past 20 yr has ranged from about 90 to about 130 acre-ft/yr (table 1-2). The Marathon aquifer is discussed in more detail by Smith in chapter 14.

## Rustler aquifer

The Rustler aquifer is located in Culberson, Jeff Davis, Loving, Pecos, Reeves, and Ward Counties (fig. 1-3). Groundwater occurs in the partly dissolved dolomite, limestone, and gypsum beds of the Rustler Formation. The poor-quality water is used primarily for irrigation, livestock, and for waterflooding operations in oil-producing areas. Pumping from the aquifer in the counties in the study area over the past 20 yr has ranged from about 170 to about 1,600 acre-ft/yr(table 1-2). The Rustler aquifer is discussed in more detail by Boghici and Van Broekhoven in chapter 15.

## West Texas Bolsons aquifers

The West Texas Bolsons aquifers, located in Culberson, Hudspeth, Jeff Davis, and Presidio Counties (fig. 1-3), are part of the Red Light Draw, Eagle Flat, Green River Valley, and Presidio-Redford Bolsons, as well as the Salt Basin. The Salt Basin is divided into the Wild Horse, Michigan, Lobo, and Ryan Flats. Composition of the bolson aquifers depends on the rock types of the nearby eroded mountains and ranges from coarse-grained volcanic rocks and limestones to fine-grained silt and clay lake deposits. Groundwater from the bolson aquifers is used for irrigation and municipal supply in parts of Culberson, Hudspeth, Jeff Davis, and Presidio Counties. Presidio, Sierra Blanca, Valentine, and Van Horn rely on the bolson aquifers for municipal water. Pumping from the aquifer over the past 20 yr has ranged from about 10,000 to about 91,000 acre-ft/yr (table 1-2). Pumping in recent years has been about 12,000 acre-ft/yr. The West Texas Bolsons aquifers are discussed by Angle in chapter 16 and Darling and Hibbs in chapter 17.

## Other aquifers

Large areas of West Texas do not have a TWDB-recognized major or minor aquifer beneath them (see white areas in fig. 1-3). This does not mean, however, that there are no groundwater resources in these areas. The Diablo Plateau area in northern Hudspeth County has the potential to produce large amounts of water (see chapter 18 by Mullican and Mace), and the Igneous aquifer probably has a greater extent than previously realized (see chapter 13 of this report). Other areas may have small, local aquifers that can supply water for limited purposes. According to TWDB information, about 900 to as much as 12,000 acre-ft/yr has been pumped from other aquifers in the area (table 1-2). Further study and evaluation of these areas will increase our knowledge of water resources in these areas.

# Springs

The many springs and seeps in the West Texas area have played an important part in the area’s history. Native Americans relied on the springs as sources of water, as did later settlers. The path of the Old Spanish Trail through the area was largely determined by the occurrence of springs in and along mountains (Brune, 1981).

Springs in the area currently provide water to ranches and small communities, such as the village of Kent in southeastern Culberson County. Farmers use the flow from Balmorhea Springs to irrigate crops in Reeves County. A number of springs are valued for aesthetic and recreational uses, such as the pool at Balmorhea Springs and the hot springs in Big Bend. The springs are sources of water to wild game and habitats of threatened and endangered species. A number of springs have stopped flowing because of lowered water tables or drought (Brune, 1981), including Kokernot Springs in Alpine, Davis Spring in Fort Davis, and, recently, Phantom Lake Springs near Balmorhea. Several species that relied on spring flow are now extinct, and others are in danger in the West Texas area (Garrett and Edwards, this volume).

Comanche Springs in Fort Stockton went dry and resulted in a landmark legal case between the Pecos County Water Control District No. 1 and Clayton Williams (Brown, 2001). The springs, first noted in 1684 (Brune, 1975), supplied water to irrigate 6,000 acres and stopped flowing after Mr. Williams installed and started pumping from a well field upgradient of the springs during the drought of the 1950's. The courts decided that the rule of capture applied and that the water district had no recourse.

# Groundwater Conservation Districts

Groundwater in Texas is governed by the common-law rule of capture unless there is a groundwater conservation district. Rule of capture allows a landowner to produce as much groundwater as the landowner chooses, absent malice or willful waste, without liability to neighbors who may claim that pumping has depleted their wells. The Legislature enabled the regulation of groundwater by creating groundwater conservation districts, the first of which was created in 1949 (High Plains Underground Water Conservation District No. 1). Groundwater conservation districts have broad regulatory authority and are recognized by the Legislature as the State’s preferred method of managing groundwater resources.

West Texas is home to four confirmed groundwater conservation districts: the Hudspeth County Underground Water Conservation District No. 1, the Culberson County Groundwater Conservation District, the Jeff Davis County Underground Water Conservation District, and the Presidio County Underground Water Conservation District (fig. 1-4a). The 2001 Legislature created two additional districts in the area: the Brewster County Groundwater Conservation District and the Middle Pecos Groundwater Conservation District (fig. 1-4a). As of fall 2001, these two districts were awaiting confirmation elections.

Although not a groundwater conservation district, the El Paso Water Utilities–Public Service Board (EPWU) is an important manager of groundwater resources in the area. EPWU manages and operates the water and wastewater system for El Paso and operates 105 wells in the Hueco-Mesilla Bolson aquifer.

# Regional Water Planning

Through Senate Bill 1, the 1997 Legislature enacted comprehensive water management to plan for drought and meet increasing demands as population grows (Hubert, 1999). Senate Bill 1 is a “bottom up” water-planning process that allows individuals representing different interest groups to serve as members of Regional Water Planning Groups. The interest groups include the public and representatives of counties, municipalities, industries, agriculture, environmental, small business, steam-electric power generating utilities, river authorities, water districts, water utilities, and others

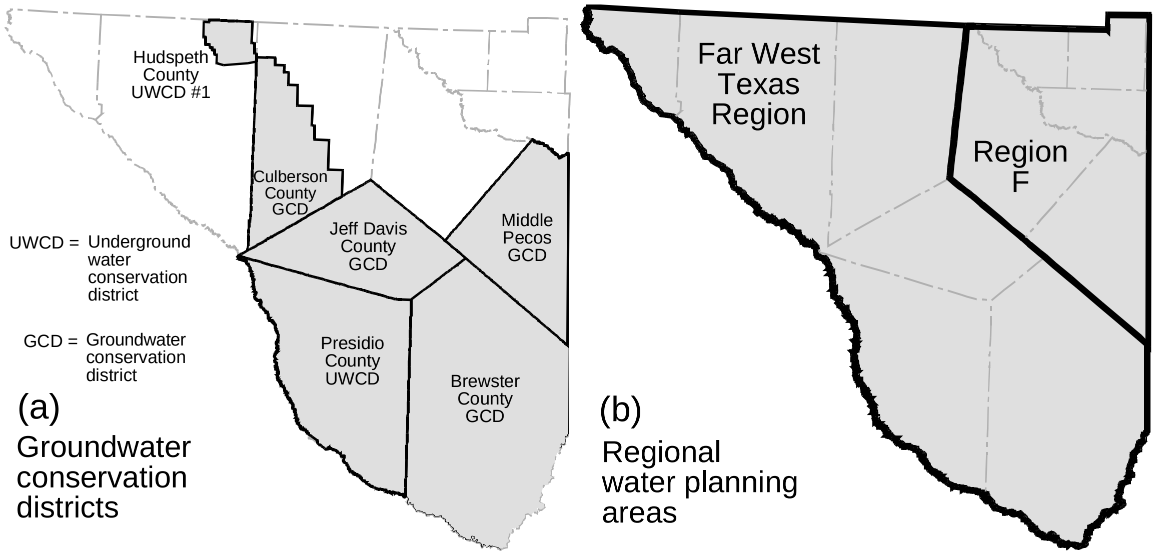


Figure 1-4: Location of (a) confirmed and created groundwater conservation districts and (b) regional water planning areas in the Far West Texas area. Note that both regions extend farther to the east.

selected by the Planning Groups. A total of 16 Planning Groups cover the state, which are charged with preparing regional water plans for their respective planning areas. These plans will show, for each planning area, how to conserve water, meet future water needs, and respond to future droughts.

Each Planning Group submitted its plan in January 2001. The TWDB is assembling the individual plans into a comprehensive State Water Plan for delivery on January 5, 2002. After January 5, 2002, the TWDB will provide financial assistance only to those projects that are consistent with the regional water plans, and the Texas Natural Resource Conservation Commission will issue water right permits only for purposes consistent with the plan. These water plans will be updated every 5 yr.

The West Texas area includes all of the Far West Texas Region and part of Region F (fig. 1-4b). The regional water plan for the Far West Texas Region shows that the region has some real challenges in meeting its future water needs, especially during a drought of record. The Planning Group showed that freshwater resources in the part of the Hueco-Mesilla Bolson aquifer available to El Paso will be greatly depleted by 2030. Furthermore, the Rio Grande will not be available for use during severe droughts.

The Far West Texas Planning Group recommended a number of strategies to meet future needs for water, including:

- conservation of surface water used for irrigation,
- purchase of irrigation rights,
- reuse of treated wastewater,

- desalination of brackish groundwater, and
- purchase and use of groundwater from outside El Paso County.

Expanded use of groundwater is intended as an emergency supply of water during times of drought.

Even with these strategies to supply more water, however, the region will be unable to meet all needs for water after 2030. Municipalities in El Paso County are projected to have water needs of over 200,000 acre-ft/yr in 2050.

Water plans for these regions can be found on the TWDB Web page (www.twdb.state.tx.us). In early 2002, a new version of the State Water Plan, which includes a statewide summary of the regional water plans, will be available from the TWDB.

# Groundwater Availability Modeling

Texas is developing new, state-of-the-art computer models of groundwater resources. In 1999, the Legislature provided initial funding for development of groundwater availability models for the major aquifers. And in 2001, the Legislature directed the TWDB to develop groundwater availability models for the minor aquifers.

There are several ongoing modeling projects in West Texas. The U.S. Geological Survey (USGS) expects to release a report in late 2001 on a model it developed of the Hueco Bolson aquifer. Sheng and others (2001) reported on how the USGS model has been used to evaluate management of the bolson aquifer in the El Paso area. TWDB is working on a model of the Edwards-Trinity Plateau aquifer (see Anaya, this volume) and expects to be completed by the end of 2002. A model of the Cenozoic Pecos Alluvium aquifer will also be done by the TWDB and its contractors by the end of 2004. The Beldon Foundation is funding work on a model of the bolson aquifer in Wild Horse Valley (CCGCD, 2001; Finch and Armour, 2001). Models of these aquifers will be useful tools for assessing the possible impacts of increased pumping on water levels and spring flows.

Several scientific models of some of the minor aquifers have been developed (e.g., Bone-Spring/Victorio Peak: Mayer and Sharp, 1998; Red Light Draw and Eagle Flat: Darling and others, 1994, and Hibbs, 1996; Wildhorse Flat: Nielson and Sharp, 1985; Diablo Plateau: Mullican and Senger, 1990, 1992). The challenge for future modeling of these minor aquifers will be availability of enough information on the aquifers and an adequate understanding of the flow.

Final reports, models, and aquifer information will be posted on the TWDB GAM Web page (www.twdb.state.tx.us/gam).

# Summary

Although the west part of Texas has been blessed with many aquifers, it faces many challenges with its desert climate and growing metropolis. Because of the arid climate, recharge to many of the aquifers is minimal. As a result of minimal recharge, large volumes of pumping cause water levels to decline. The resulting water-level declines cause reductions in the volume of fresh water, flow to springs, and water quality. El Paso is particularly susceptible to drought: the Rio Grande will not offer any water in a severe drought, and fresh groundwater resources in the Hueco-Mesilla Bolson are expected to be depleted by 2020.

Groundwater conservation districts, regional water planning groups, and groundwater availability modeling are helping to further our understanding of the aquifers and the options for meeting future water needs. However, additional study is needed, particularly on the less-studied minor aquifers in the area.

# Acknowledgments

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# Chapter 2

# Geologic History of West Texas

Kevin Urbanczyk[1], David Rohr[1], and John C. White[1]

## Introduction

The region embraced within the Trans-Pecos region of Texas encompasses many snapshots of North American geologic history. Precambrian crystalline metamorphic rocks are exposed in the Franklin Mountains, Van Horn Mountains, and Sierra Diablo Mountains. Xenoliths of these rocks recovered from volcanic rocks in the Davis Mountains, Bofecillos Mountains, and Chisos Mountains provide strong evidence that almost all of Trans-Pecos Texas is underlain by Precambrian rocks similar to those that crop out at the surface. Cambrian to Pennsylvanian rocks crop out in the Franklin Mountains, Marathon Basin, Solitario, and at Persimmon Gap. These rocks represent a transgressive, then regressive, marine sequence that was caught between the North American continent and another unidentified continent during the Pennsylvanian and intensely deformed and thrust onto North America forming the Marathon-Ouachita Mountains. The foreland basin of these mountains became the Permian Basin, and the carbonate rocks associated with this intracratonic sea now crop out in the Guadalupe, Glass, Apache, Van Horn, and Sierra Diablo mountain ranges. A depositional hiatus from the Triassic to Mid-Cretaceous was followed by the deposition of Mid- to Late-Cretaceous limestone that covers much of central and west Texas and frequently hosts important aquifers. From the Late Cretaceous to the Early Tertiary, these rocks were locally deformed during the Laramide Orogeny, which can be seen in the Del Norte-Santiago Mountains, Mariscal Mountain, the Terlingua-Fresno Monocline, and in the Chihuahua Tectonic Belt. Laramide compression was followed by a long period of large-scale ignimbritic volcanism in Trans-Pecos Texas. As compression continued to wane, ignimbritic volcanism yielded to smaller-scale effusive volcanism that was coupled with extensional tectonics, resulting in Basin and Range structures and related mountain ranges in the Trans-Pecos. Between these ranges, which include the Franklin, Hueco, Guadalupe, Delaware, Sierra Diablo, Sierra Vieja, and Van Horn mountains, large basins formed that filled with thick sequences of gravel and sand eroded from the adjacent mountains. It is in this setting that we presently reside.

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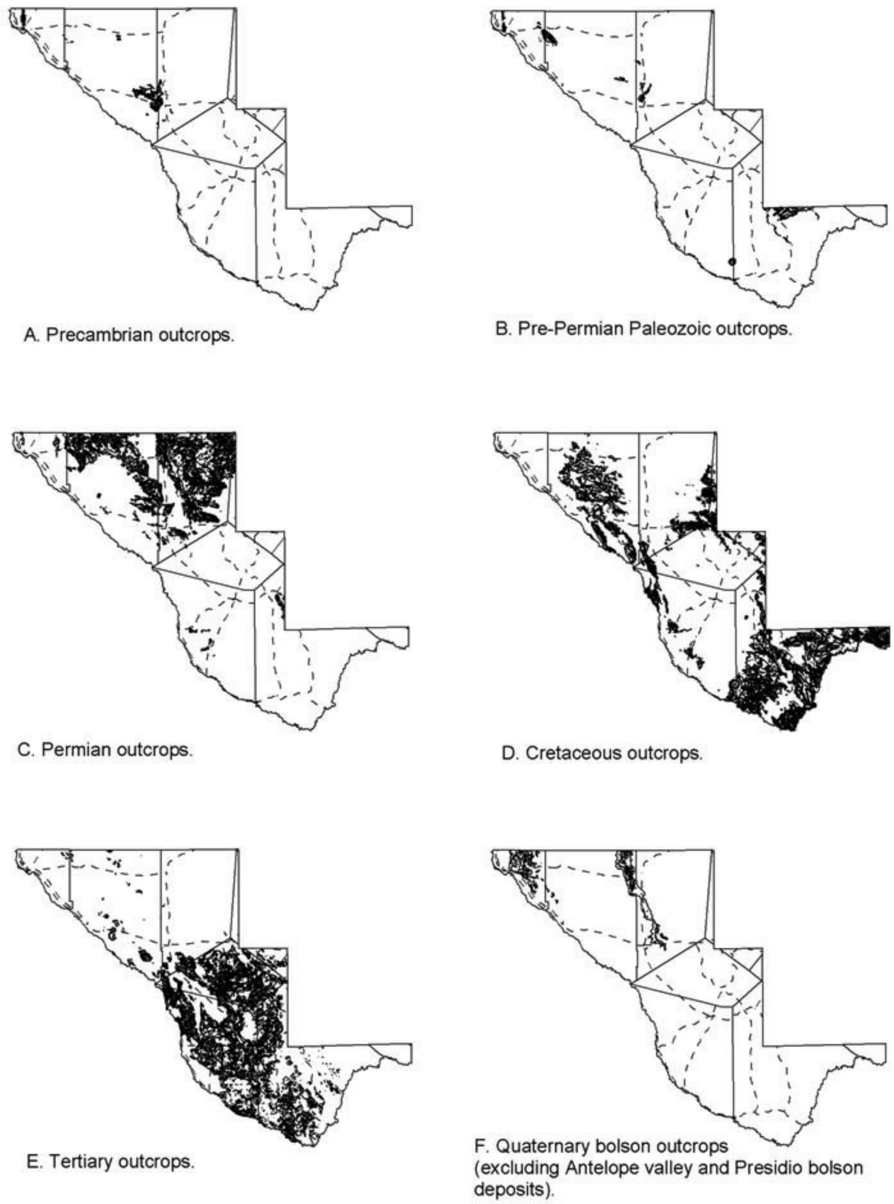


Figure 2-1: Geologic maps of time intervals discussed in the text.

# Precambrian Geology

Precambrian metamorphic rocks crop out only near Van Horn and El Paso (fig. 2-1). The rocks have been interpreted to represent igneous and sedimentary rocks associated with an island arc system that were metamorphosed and accreted to the margin of North America during collision with an unknown continent. These rocks are coeval with the Llano terrane of Central Texas, which also formed during the Grenville orogeny, at approximately 1 billion years ago.

# Paleozoic Geology

Although Paleozoic strata underlie most of Trans-Pecos Texas, they crop out only in a few regions (fig. 2-1). These can be broadly divided into the Paleozoic shelfal facies in the El Paso and Van Horn areas, the basinal (or geosynclinal) Marathon facies, and the Permian intracratonic basinal facies in the Guadalupe Mountains southward through the Sierra Diablo and Apache Mountains, the Glass Mountains, and the Shafter-Pinto Canyon area.

## Paleozoic Facies and Tectonics

What is now the Trans-Pecos was situated along the south margin of the North American continent (Laurentia) during early Paleozoic time. The El Paso and Van Horn areas were the site of deposition of limestone, dolomite, and sandstone in a shallow, tropical sea located near the edge of the continental shelf (Stoudt, 1996). The strata in these areas were slightly deformed during Mesozoic and Cenozoic uplift, resulting in the beds being tilted from horizontal. The Franklin Mountains are a good example of a block-faulted mountain range.

Lower Paleozoic sandstone and limestone unconformably overlie Precambrian granite and metamorphic rocks in the El Paso and Van Horn areas. In the Sierra Diablo, late Paleozoic uplift has resulted in the older Paleozoic rocks being stripped off, and upper Paleozoic rocks lie unconformably over Precambrian.

In contrast, the Paleozoic rocks in the Marathon Basin have very different lithologies and have been folded and faulted to a much greater degree. These rocks, exposed in the Marathon Basin and in the Solitario to the southwest, are part of the Ouachita Orogenic Belt. The strata were originally deposited as shales, cherts, and turbidite sandstones in a deep oceanic setting south of the North American continent and represent one of the longest intervals of essentially continuous deposition known (King, 1978). At the close of the Paleozoic, the south margin of North America collided with South America (then part of Gondwanaland), and the rocks were deformed and thrust northward over the North American platform margin. Folding and thrust faults (fig. 2-2) result in some wells penetrating distinctive units such as the Caballos Novaculite (Folk and McBride, 1978) several times in one borehole.

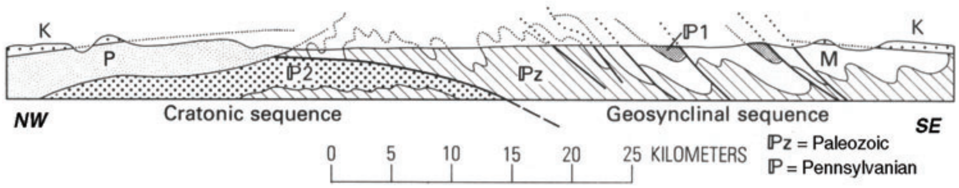


Figure 2-2: Cross section of the Marathon Basin showing the complex structure. Deformation near the rocks during the Ouachita Orogeny end of the Paleozoic caused extensive folding and faulting when geosynclinal facies were thrusted over cratonic (modified from King, 1981).

## The Permian Basin

Up to 3,000 m of Permian rocks underlies much of West Texas and southeastern New Mexico. Because of prolific oil production, the stratigraphy of the West Texas or Permian Basin has been studied in great detail (Hill, 1996). Subsurface information in the forms of well logs and seismic profiles is available for most of the area. The Permian Basin can be thought of as a small but deep inland sea that developed in a sag of the North American continental crust after collision of the continent with Gondwanaland.

The earliest oil drilling in the Permian Basin revealed that the basin actually consists of three smaller basins: the Midland, Delaware, and Marfa Basins. During the Permian Period, these basins were enclosed marine basins accumulating organic-rich sediments. The edges of the two basins in Trans-Pecos Texas (the Delaware and Marfa Basins) were partly surrounded by massive limestone reefs. Parts of the reef can now been seen at the surface in the Guadalupe Mountains, including the peak of El Capitan. The margins of the basins are characterized by rapid lithologic facies changes. The basinal facies are bedded shales and sandstones, the reef is massive limestone, and the back reef is commonly dolomite. A small amount of Permian gypsum, the Castile Formation, is exposed near the Texas-New Mexico border. Permian strata were uplifted to their present elevation during the Cenozoic, and have been subjected to relatively minor deformation, mostly normal faulting (fig. 2-3).

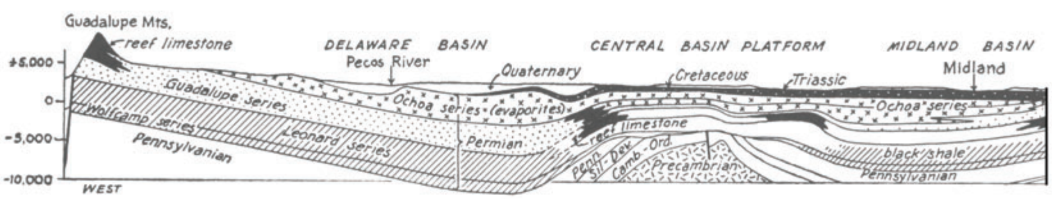


Figure 2-3: Cross section of the Permian Basin. Most of the structural features were present during the Permian and controlled the sedimentary facies (modified from King, 1959).

# Mesozoic Geology

The paucity of Triassic and Jurassic rocks in Trans-Pecos Texas either at the surface or in the subsurface strongly suggests that this region was subaerially exposed during these periods and experienced no active tectonism (McCormick and others, 1996). In adjacent northern Chihuahua, however, rifting and subsidence of the Chihuahua Trough related to the opening of the Gulf of Mexico and Atlantic Ocean began during the Middle Jurassic. Triassic rocks are lacking throughout the Trans-Pecos area and adjacent Chihuahua. Jurassic rocks have been mapped at the surface only in the Malone Mountains, near Sierra Blanca (Albritton and Smith, 1965) and in the subsurface in the Chihuahua Trough (Henry and Price, 1985). Widespread clastic and carbonate sedimentation did not begin in the Trans-Pecos until the Middle Cretaceous with the deposition of the Comanchean series rocks, which represent primarily carbonate sedimentation associated with a widespread, intracontinental sea that inundated much of North America from Texas to Alberta (fig. 2-1). Late Cretaceous uplift related to the Laramide orogeny is responsible for Gulfian series rocks, a regressive sequence of limestone to terrigenous shales and sandstones that overlie the Comanchean series.

Comanchean deposition began with a coarse-grained, basal conglomerate consisting of clasts of Paleozoic rocks. This conglomerate unconformably overlies the Pennsylvanian Tesnus Formation, where it is exposed in the Solitario, Big Bend Ranch State Park, and at Persimmon Gap, Big Bend National Park. In the Solitario, it is mapped separately as the Shutup Conglomerate; in Big Bend National Park, it is mapped as a lower part of the Glen Rose Limestone (McCormick and others, 1996). The Glen Rose Limestone is overlain by the Del Carmen Limestone, which is equivalent to the Edwards Limestone in Central Texas. Completing the Comanchean (Lower Cretaceous) series in West Texas above the Del Carmen Limestone are the Sue Peaks Formation (equivalent to the Kiamichi Formation) and the Santa Elena Limestone (equivalent to the Georgetown Limestone) (Maxwell and others, 1967; McCormick and others, 1996). The Comanchean series is overlain by the Gulfian series, a regressive sequence beginning with the Del Rio Clay, followed by the Buda Limestone, and the Boquillas, Pen, Aguja, and Javelina Formations.

# Cenozoic Geology

Sedimentation from the Late Cretaceous into the Early Tertiary was nearly continuous and shows a regressive sequence of marine limestone yielding to swampy shale and terrigenous sandstone (Maxwell and others, 1967). This sedimentation was contemporaneous with the compressive Laramide orogeny, which uplifted long, linear mountain chains (including the Del Norte-Santiago Mountains and the Chihuahua Tectonic belt) and was the most likely cause of the retreat of the epeiric seas in Trans-Pecos Texas (Lehman, 1986).

The Laramide orogeny, lasting from about 100 to 50 million years ago (Ma), began with the onset of subduction of the Farallon plate under the North American plate (Coney and Reynolds, 1977; McDowell and Clabaugh, 1979). Laramide compression produced folds, thrust faults, and high-angle reverse faults; shortened Trans-Pecos Texas about 3.4 percent; and reactivated west-northwest-striking, left-lateral strike slip basement faults (Barker, 1987). Laramide features in the Trans-Pecos include the Terlingua-Fresno monocline, Mariscal Mountain, the Del Norte-Santiago Mountains, and the Chihuahua Tectonic Belt (Muehlberger and Dickerson, 1989). Laramide compression peaked in the Late Paleocene and ended during the Eocene (Coney and Reynolds, 1977; Price and others, 1987). Magmatism began during the Eocene (~48 Ma) as compressive stresses began to wane and continued through the Oligocene and into the Miocene, ending about 17 Ma (Henry and McDowell, 1986).

Early magmatism began with the emplacement of a suite of intrusive rocks in the El Paso area around 48 Ma and was followed by activity in the Christmas Mountains and western Big Bend National Park between 45 and 40 Ma. Large-scale silicic volcanism began in the Davis Mountains and Sierra Vieja about 39 Ma and continued until about 35 Ma. The Crossen Trachyte and Star Mountain Rhyolite are generally accepted to be the oldest volcanic units within the Davis Mountains, erupting from unknown sources about 38.6 Ma (Henry and others, 1994). The Buckhorn caldera, the first recognized caldera in the Davis Mountains, formed about the same time, with the eruption of the Gomez Tuff (Parker, 1986). Volcanism was nearly contemporaneous in the Sierra Vieja and Van Horn Mountains, with the eruption of the Buckhorn Ignimbrite from unknown sources and the eruption of the Chambers and High Lonesome Tuffs from the Van Horn caldera (Henry and Price, 1986). The Solitario laccocaldera also erupted during this initial phase. In the southern Davis Mountains, several widespread flows of mafic to intermediate lavas erupted, producing the lavas of the Pruett Formation, including the Sheep Canyon Basalt, Potato Hill Andesite, and Cottonwood Springs Basalt. This activity was followed by the formation of the Paisano volcano, a large trachyte-rhyolite shield volcano, around 36 Ma (Parker, 1983). The Eagle Mountain and Quitman Mountain calderas are approximately contemporaneous with the Paisano volcano (Henry and Price, 1984). Large-scale volcanism in the Davis Mountains ended with the eruption of the Barrel Springs and Wild Cherry Formations from the Paradise Mountain–Pine Peak caldera between 36.4 and 35.4 Ma (Henderson, 1979).

Following the cessation of activity in the Davis Mountains, volcanism shifted southward to Big Bend National Park and the Chinati Mountains. The earliest known episode of volcanism in Big Bend National Park is represented by the 47 to 40 Ma Alamo Creek Basalt. Magmatism in Big Bend resumed between 35 and 32 Ma with the eruption of lavas and tuffs of the Chisos Group, followed by those of the South Rim Formation. From the Chinati Mountains area, lavas and tuffs of the Shelly Group and Morita Formation were erupted from unknown sources between 34 and 32 Ma. Between 33 and 32 Ma, the Chinati Mountains caldera formed as a result of the eruption of the Mitchell Mesa Tuff, the largest (>1000 $km^3$) ash-flow tuff in Trans-Pecos Texas. This eruption was followed by several hundreds of thousand years of continued volcanic activity that resulted in the accumulation of a large volcaniclastic alluvial apron around the Chinati Mountains, which is represented by the Tascotal Formation.

During the Oligocene, between 31 and 28 Ma, the compressive stress regime yielded to a tensional (extensional) environment, which persisted through the remainder of the

Cenozoic (Muehlberger and others, 1978; Henry and others, 1991). Volcanism during this phase was restricted to Big Bend Ranch State Park and adjacent Mexico and began with the eruption of the San Carlos and Santana Tuffs from the Sierra Rica caldera complex in northern Mexico at 30 and 27.8 Ma. Volcanism continued between 27.8 and 26 Ma with the eruption of the Rawls Formation, a complex series of mostly mafic to intermediate lavas. Faulting associated with extensional tectonics produced a discontinuous series of north-northwest-trending pull-apart grabens that terminate at west-northwest-trending strike-slip faults (Barker, 1987). Mafic volcanism closely associated with faulting continued until about 17 Ma.

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# Chapter 3

# Evaluation of Groundwater Recharge in Basins in Trans-Pecos Texas

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## Introduction

The Trans-Pecos Texas region, in the southeast part of the Basin and Range province, consists of topographically high plateaus and mountains separated by major normal faults from adjacent, topographically low desert basins. The basins were progressively filled by detritus eroded from adjacent ranges and from Colorado and New Mexico by the ancestral Rio Grande (Gustavson, 1990). The study area includes the Diablo Plateau, Hueco Bolson, and Eagle Flat and Red Light Basins (fig. 3-1). Depth to groundwater ranges from 100 to 900 ft (30 to 274 m) beneath the Diablo Plateau, 350 to 500 ft (107 to 152 m) in the Hueco Bolson, 100 to 1,100 ft (30 to 335 m) in Eagle Flat Basin, and 10 to 500 ft (3 to 152 m) in Red Light Basin.

Trans-Pecos Texas lies within the northern Chihuahuan Desert (King, 1948). The region has a subtropical, arid climate (Larkin and Bomar, 1983). Long-term average annual precipitation ranges from 11 inches (280 mm) in Hueco Bolson to 12.6 inches (320 mm) in Sierra Blanca in Eagle Flat and Red Light Basins. Most precipitation occurs in the summer months as thunderstorms.

The geology of the different regions is quite variable. The Diablo Plateau site consists of shallow alluvium over Cretaceous limestone. The Hueco Bolson is made up of about 0 to 75 ft (~0 to 23 m) of coarse-grained material over 600 ft (183 m) of clay interbedded with silts and sands. Basin-fill deposits in the Eagle Flat region are generally fine grained muds, which are overlain by sand sheets in some parts of the basin.

## Methods

The theoretical basis for the various techniques used to evaluate groundwater recharge in the study area is described next in order to provide the reader with a conceptual understanding of the different approaches. We used physical, chemical, and isotopic data from the unsaturated zone to determine whether groundwater recharge was occurring on

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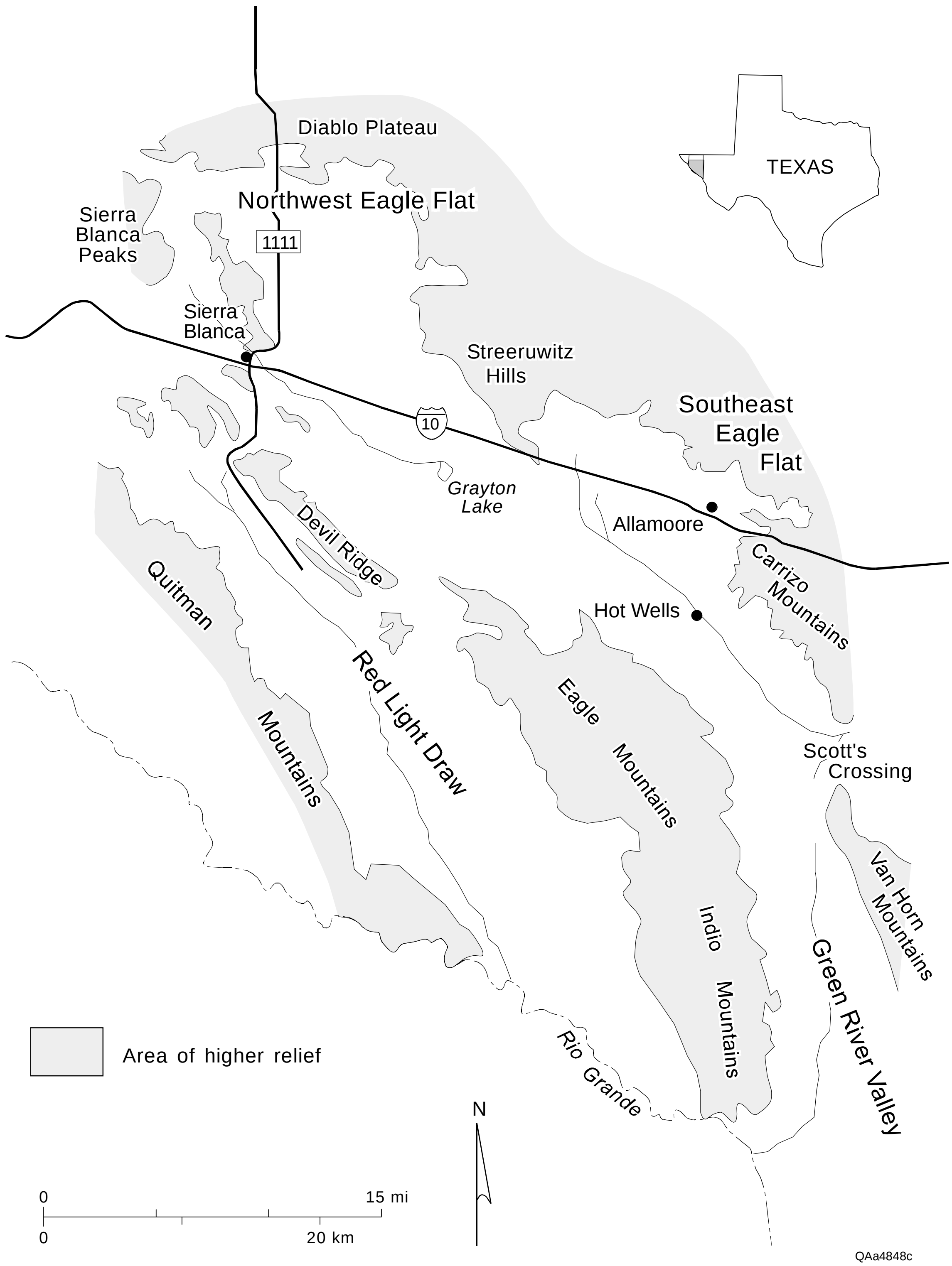


Figure 3-1: Location of the study area, including Diablo Plateau and Hueco Bolson and Eagle Flat and Red Light Basins.

the valley floors and on the Diablo Plateau. The unsaturated zone is the zone between the land surface and the water table, where pore spaces are filled with water and air. Water pressures are negative in the unsaturated zone and positive in the saturated zone. Physical data were used to determine the direction of water movement at a site. Chemical and isotopic tracers were used to quantify net water fluxes over long time periods. Groundwater tracers were also used on the Diablo Plateau and in the Eagle Flat and Red Light Basins to evaluate recharge.

## Direction of Water Movement

Increasing potential energy with depth in the subsurface in the study area indicates that water is moving upward and that there is no recharge there. We can understand this concept by considering the gravitational force field, which is related to elevation of an object relative to a datum. Gravitational potential energy increases with elevation above a datum, such as the land surface, and objects will move from higher to lower gravitational potential energy (i.e., from higher to lower elevations). In soils, potential energy is related to forces in the soil, such as capillary, adsorptive, and osmotic forces (water-potential energy), and water will move from regions of high to low potential energy. Energy can be expressed in different units such as pressure (mega Pascals [Mpa] bars, or atmospheres) or head (ft, m). One MPa is equivalent to 10 bars, or 102 m. Thermocouple psychrometers were used to measure the relative humidity of the soil air, which was converted to matric potential energy according to the Kelvin equation.

## Chemical and Isotopic Tracers

Chloride concentrations in unsaturated-zone pore water have been used to study recharge in semiarid systems over time scales ranging to thousands of years. Chloride concentrations in unsaturated-zone pore water are inversely proportional to water flux: high chloride concentrations in pore water indicate low water fluxes because chloride accumulates in the unsaturated zone as a result of evapotranspiration. In contrast, low chloride concentrations indicate high water flux because chloride is flushed through the unsaturated zone. The recharge rate, or water flux, is calculated by dividing the chloride input (precipitation × chloride concentration in precipitation) by the chloride concentration in the unsaturated zone. The age represented by chloride at any depth can be estimated by dividing the total amount of chloride from the surface to that depth by the annual chloride input.

Chlorine-36, a radioactive isotope of chlorine, has been used to a limited extent to date pore water in arid unsaturated zones. Chlorine-36 ($^{36}Cl$), which is produced naturally in the atmosphere (Bentley and others, 1986), has a half-life of 301,000 yr. The term *half-life* refers to the amount of time required for one-half of the atoms of a radioactive element to decay to half of the initial value. Nuclear-weapon tests conducted in the Pacific between 1952 and 1958 resulted in $^{36}Cl$ concentrations in rainfall that were as much as 1,000 times greater than natural fallout levels (fig. 3-2; Bentley and others, 1986). Bomb-pulse $^{36}Cl:Cl$ ratios have been used to estimate water fluxes during the past 40 yr and to evaluate preferential flow (Phillips and others, 1988; Scanlon, 1992;

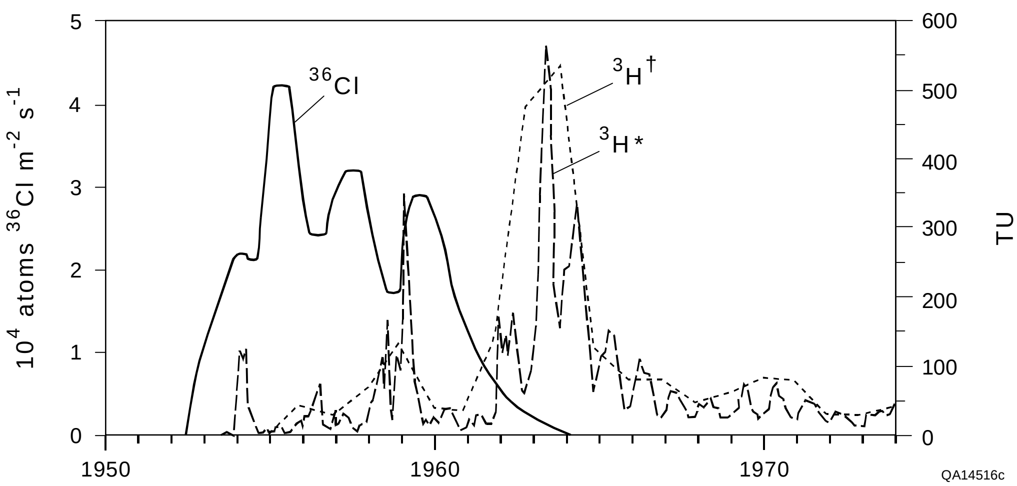


Figure 3-2: Temporal variations in $^{3}H$ and $^{36}Cl$ fallout from Ottawa, Canada (Scanlon, 1992).

Fabryka-Martin and others, 1993) (fig. 3-1). The depth of this high concentration of $^{36}Cl$:Cl ratios in the soil water can be used to estimate how far water has moved during the time between bomb-pulse fallout and soil-sample collection (~ 40 yr). Radioactive decay of $^{36}Cl$ can also be used to estimate water ages to 1 million yr.

Tritium ($^{3}H$), a radioactive isotope of the element hydrogen (H) and produced naturally in the upper atmosphere, results in concentrations in precipitation over the northern hemisphere of 1 to 20 tritium units (TU) (Michel, 1989; Solomon and Sudicky, 1991) and averages approximately 5 TU (Mazor, 1991, p. 151). We define 1 TU as one atom of $^{3}H$ in $10^{18}$ atoms of H, and the half-life of $^{3}H$ is 12.43 yr. Tritium concentrations increased from 10 to ≥ 2,000 TU during atmospheric nuclear testing (IAEA, 1983) that began in 1952 and peaked in 1963 to 1964 (fig. 3-2). Tritium generated by the above-ground detonations of nuclear weapons is referred to as *bomb-pulse tritium*. Tritium that recharged groundwater in 1953 would have decayed to less than 0.5 TU by 2001, assuming an initial average value of 5 TU. Therefore, tritium concentrations of less than 0.5 TU are often interpreted to indicate recharge before 1952, and higher levels of tritium are regarded as indicative of post-1952 recharge. Subsurface distribution of bomb-pulse tritium in the unsaturated zone can be used to estimate how deep water has moved in the past 40 yr—a procedure similar to the one described for $^{36}Cl$.

Carbon-14 ($^{14}C$), a radioactive isotope of carbon, is formed naturally in the upper levels of the atmosphere. Because of its long half-life, $^{14}C$ can be used to date groundwater with recharge ages to 10,000 yr. The concentration of $^{14}C$ is reported as percent modern carbon (pmc), which is the ratio of $^{14}C$ in a sample of groundwater measured against the concentration in an internationally accepted standard of oxalic acid. The apparent age of a water sample is inversely related to the concentration of $^{14}C$. A value of 100 pmc indicates modern water. A value of 50 pmc represents an apparent age of 1 half-life, or 5,730 yr.

Stable isotopes of oxygen ($^{18}O$) and hydrogen ($^{2}H$, or deuterium) have been used to evaluate unsaturated flow. Because of coupling between fractionation of oxygen and hydrogen, a plot of stable isotopes of water samples from precipitation, rivers, and lakes throughout the world follows a straight line called the *global meteoric water line* (Craig, 1961).

Although the isotopic composition of individual rain events is highly variable, the annual mean values are fairly constant (Gat, 1981). The isotopic composition of pore water in the unsaturated zone is generally either depleted or enriched relative to the mean isotopic composition of rainwater. Enriched values of the stable isotopes relative to the meteoric water line indicate evaporation from surface water before infiltration or from pore water in the unsaturated zone. Slopes of $^{2}H$ versus $^{18}O$ are about 5 for surface-water evaporation but can decrease to 2 for evaporation of pore water in the unsaturated zone (Allison, 1982).

## Field and Laboratory Methods

Sediment samples were collected for water-potential measurements in the laboratory from 8 boreholes to 46-ft (14-m) depth in the Hueco Bolson and 48 boreholes to 102-ft (31-m) depth in Eagle Flat Basin (Scanlon and others, 1991, 2000). The boreholes were drilled by using a hollow-stem auger, and samples were collected in split-tube core barrels (1.5 m [4.9 ft] long). To monitor water-potential variations, field psychrometers were installed in two locations in the Hueco Bolson and in one location in Eagle Flat Basin.

Sediment samples were collected from 10 boreholes to 31.5-ft (9.6-m) depth on the Diablo Plateau, 10 boreholes to 73.5-ft (22.4-m) depth in Hueco Bolson, and 52 boreholes to 102-ft (31-m) depth in Eagle Flat Basin for laboratory determination of chloride content (fig. 3-1). $^{36}Cl$:Cl ratios and $^{3}H$ were measured in soil samples collected in a shallow pit in Hueco Bolson and in five boreholes in Eagle Flat Basin. Samples for $^{36}Cl$:Cl were analyzed by Accelerator Mass Spectrometry at Lawrence Livermore National Laboratory. Samples for $^{3}H$ from Hueco Bolson were analyzed unenriched at the University of Waterloo, and those from Eagle Flat Basin were analyzed, enriched by standard direct scintillation methods, at the University of Arizona Tritium Laboratory or by gas proportional counting at the University of Miami Tritium Laboratory. Samples from seven boreholes in Eagle Flat Basin were analyzed for stable isotopes of hydrogen ($^{2}H$) and oxygen ($^{18}O$). The analyses were done by the Desert Research Institute (University of Nevada, Las Vegas).

Groundwater samples were collected from 30 wells on the Diablo Plateau for $^{3}H$ analysis. A total of 74 samples were collected from 59 wells in Eagle Flat and Red Light Basins for $^{14}C$ and $^{3}H$ analyses.

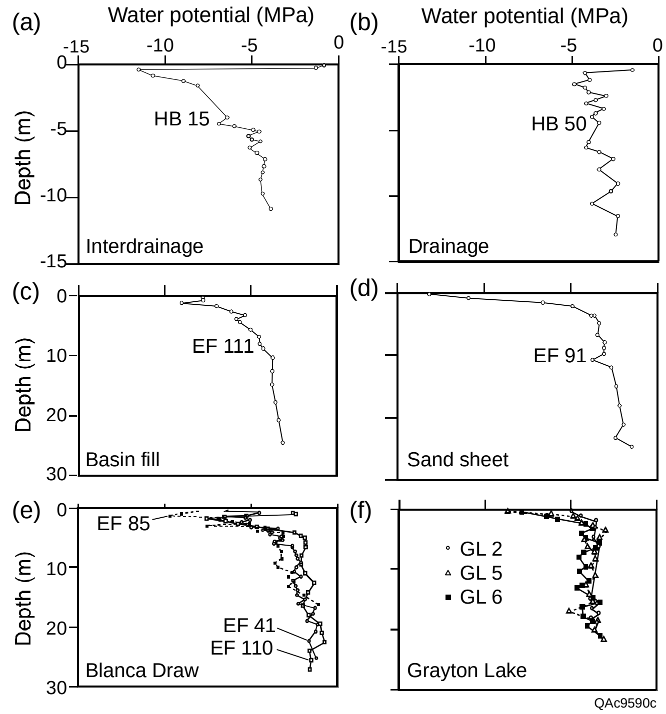


Figure 3-3: Typical water-potential profiles in (a, b) Hueco Bolson (interdrainage area, HB 15, drainage area, HB 50) and in (c–f) Eagle Flat Basin (interdrainage mud flats [EF 111], interdrainage sand sheet [EF 91], Blanca Draw [EF 41, 85, 110], and Grayton Lake [GL 2, 5, and 6]).

# Results

## Valley-Floor Settings in the Hueco Bolson and Eagle Flat Basin

Measured water potentials in interdrainage areas were extremely low in the Hueco Bolson and Eagle Flat Basin, indicating that the sediments are dry (Scanlon and others, 1991, 2000) (fig. 3-3). Water potential generally increased with depth, suggesting that water is moving upward in the profile as a result of long-term drying. Long-term monitoring of water potentials in the Hueco Bolson and Eagle Flat Basin in interdrainage areas indicates that during the past 7 to 10 yr, water did not move below the top 3 ft in a sandy location in Hueco Bolson or below 1 ft in a silt-loam site in the Hueco Bolson or a similar site in Eagle Flat Basin. Drainage areas in the Hueco Bolson and Eagle Flat Basin (Blanca Draw) also had low water potentials and upward gradients indicating upward water movement. The drainage areas are generally characterized by vegetation, such as mesquite, that can readily remove water from the subsurface and dry out the soil profile. However, water potentials at depths of more than 15 to 30 ft beneath Blanca Draw were much higher than those in adjacent regions, indicating downward water movement at these depths. Water potentials beneath Grayton Lake playa in Eagle Flat Basin were also

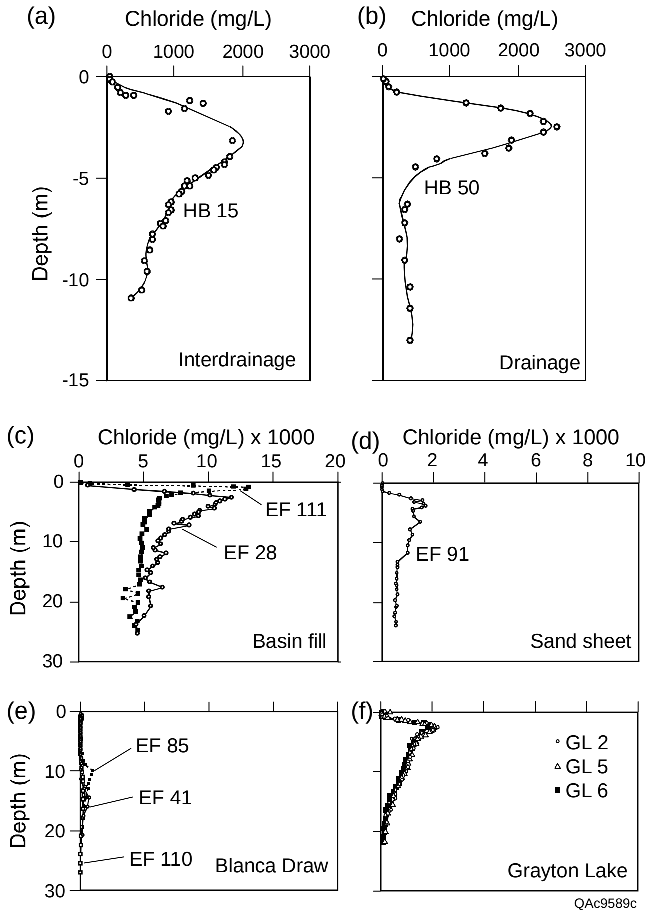


Figure 3-4: Typical chloride profiles in (a, b) Hueco Bolson (interdrainage area, HB 15; drainage area, HB50) and in (c–f) Eagle Flat Basin (interdrainage basin fill [EF 28, 111], interdrainage sand sheet [EF 91], Blanca Draw [EF 41, 85, and 110], and Grayton Lake [GL 2, 5, and 6]).

low, indicating dry conditions. Recent studies indicate that these upward water-potential gradients in interdrainage areas may take thousands of years to develop, and these profiles may reflect drying of the soils since Pleistocene time (Walvoord and others, 1999).

Chloride concentrations in the unsaturated zone in interdrainage areas were generally high, indicating low water fluxes (fig. 3-4). Maximum chloride concentrations in the Hueco Bolson profiles ranged from 1,858 to 9,343 mg/L. The chloride profiles are generally bulge shaped, with low concentrations near the surface increasing to a maximum at depth and decreasing below the peak to total depth. The bulge has been attributed to higher water fluxes before the last 10,000 yr (during Pleistocene glaciation) and reduction or change from downward to upward flow during the Holocene (~last 10,000 yr). Water fluxes at depth below the bulge in the Hueco Bolson profiles were as much as 0.04 inches/yr (1 mm/yr). Chloride profiles in the sand-sheet areas of Eagle Flat

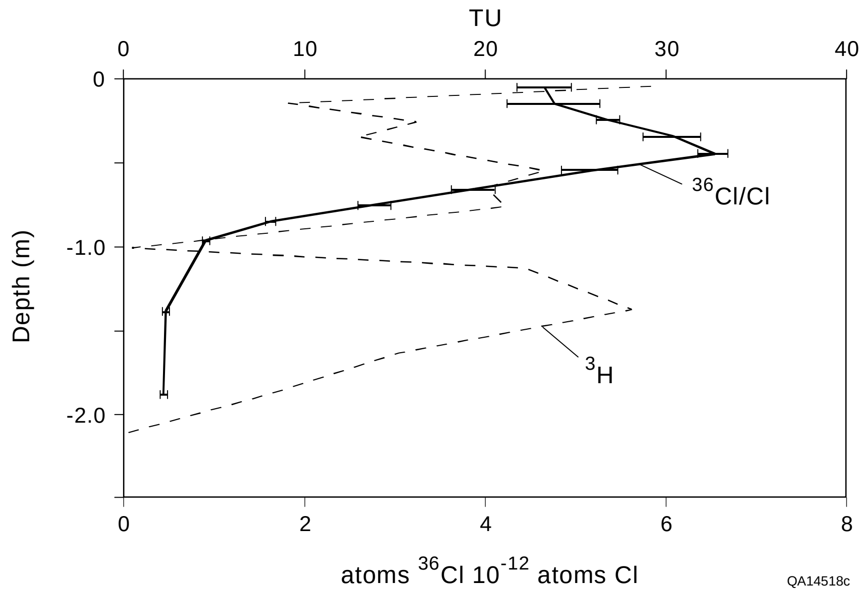


Figure 3-5: Vertical profile of $^{36}Cl$:Cl ratios in samples from borehole 51 and $^{3}H$ concentrations in samples from borehole 52 (Hueco Bolson).

Basin were similar to those in the Hueco Bolson, with maximum chloride concentrations ranging from 1,716 to 7,831 mg/L. Higher chloride concentrations (maximum concentrations in profiles: 5,912 to 17,821 mg/L) were found in the finer grained sediments in interdrainage areas of the Eagle Flat Basin. The age of the water in these profiles was as much as 130,000 yr at a depth of 25 m. Many of these profiles seem to show no response to Pleistocene climate change and remained uniformly high during that time. The calculated age of the water is consistent with dating results from radioactive decay of $^{36}Cl$ in an interdrainage profile.

Because water ponds in Grayton Lake and was ponded during the study for about 1 yr, chloride concentrations were expected to be negligible, but they were higher. All three chloride profiles in Grayton Lake were bulge shaped, and peak concentrations ranged from 1,084 to 1,315 mg/L. These concentrations indicate low water fluxes (~ 0.01 inches/yr [0.25 mm/yr]), which are attributed to the fine-grained sediments beneath the playa lake. Low chloride concentrations were found beneath the ephemeral stream setting in Eagle Flat Basin (Blanca Draw) (mean chloride, 349 mg/L), suggesting that chloride was flushed out during ponded conditions.

The depth distribution of bomb-pulse $^{36}Cl$ :Cl was examined in a small ephemeral stream in Hueco Bolson to determine how deep water migrated during the time between bomb fallout (mid-1950's) and the time of sampling (1989). Bomb-pulse $^{36}Cl$ :Cl was restricted to the upper 4.1 ft (1.25 m) in the Hueco Bolson site; therefore, water must not have moved deeper than 4.1 ft in the past 35 yr (fig. 3-5). The peak depth was located at 1.6 ft (0.5 m). A water velocity of 0.55 inches/yr (14 mm/yr) was calculated on the basis of the depth of the $^{36}Cl$ :Cl peak (4.1 ft [0.5 m]) and the time between peak fallout and soil-

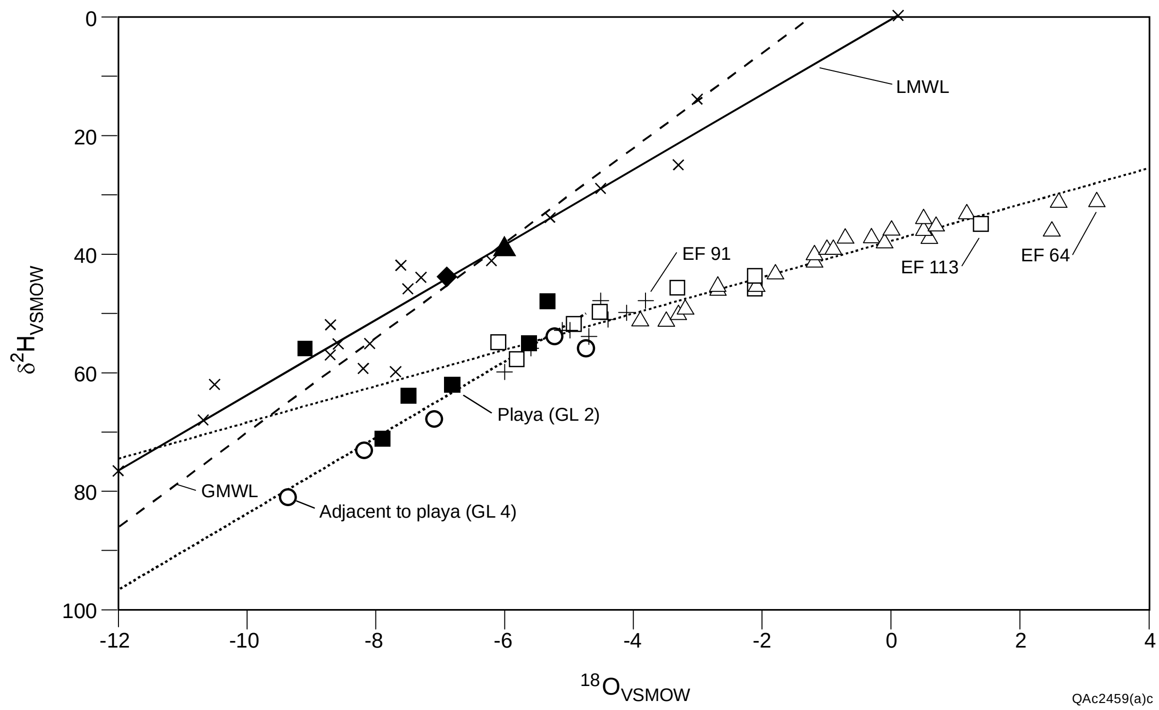


Figure 3-6: Deuterium versus oxygen-18 in soil samples collected beneath Grayton Lake and in interdrainage profiles in Eagle Flat Basin (GMWL, global meteoric water line, LMWL, local meteoric water line).

sample collection (35 yr). The corresponding water flux (0.055 inches/yr [1.4 mm/yr]) was calculated by multiplying water velocity by the average water content in the profile (0.1 $ft^3$/ $ft^3$). Distribution of bomb-pulse $^3H$ was also evaluated. The $^3H$ profile was multipeaked (fig. 3-5), with $^3H$ concentrations ranging from 23 to 29 TU. The deepest peak (1.4 m) was attributed to the 1963 peak fallout, which resulted in a water velocity of 2.2 inches/yr (56 mm/yr) (assuming a time of 25 yr since peak fallout). The corresponding water flux of 0.28 inches/yr (7 mm/yr) was calculated on the basis of an average volumetric water content of 0.13 $ft^3$/ $ft^3$. High $^3H$ concentrations beneath Grayton Lake playa were attributed to preferential flow along desiccation cracks in the floor of the playa.

Stable isotopes of oxygen and hydrogen were used to evaluate subsurface evaporation. A local meteoric water line was estimated from $^{18}O$ and $^2H$ data from precipitation that is similar to the global meteoric water line (fig. 3-6). Interdrainage profiles show enrichment of $\delta^{18}O$ relative to $\delta^2H$ that is described by $\delta^2H = 3.1\delta^{18}O$. This low slope is consistent with evaporation of pore water in the unsaturated zone. In contrast, stable isotope data from beneath Grayton Lake plot parallel the meteoric water line, indicating negligible evaporation in these sediments.

## Recharge on the Diablo Plateau

Results of tritium analysis in groundwater samples in the Diablo Plateau indicate that bomb-pulse tritium was found in most wells in the area (Kreitler and others, 1987). Tritium concentrations ranged from more than 0.8 to 32 TU. Approximately 18 percent of the wells had bomb-pulse tritium (> 0.8 TU). The most likely mechanism for rapid recharge to groundwater is through arroyos and depressions where water ponds at the surface. This hypothesis is supported by the chloride profiles, which were variable in arroyo settings. Low chloride concentrations were measured in one profile (135 mg/L), which corresponds to a water flux of 0.0688 inches/yr (1.75 mm/yr). The other two profiles in arroyo settings had higher chloride concentrations (2,286 to 3,292 mg/L), which resulted in lower water fluxes (0.0041 to 0.0028 inches/yr [0.10 to 0.07 mm/yr]). Low chloride concentrations were also measured in a closed depression (186 mg/L), which corresponds to a water flux of 0.0501 inches/yr (1.27 mm/yr). In contrast, chloride concentrations in interarroyo settings were high in all profiles. Chloride concentrations ranged from 4,211 to 10, 226 mg/L, and calculated water fluxes were low in these settings (0.0022 to 0.0009 inches/yr [0.056 to 0.023 mm/yr]).

## Mountain-Front Recharge in Eagle Flat and Red Light Basins

Results of $^{3}H$, $^{14}C$, and apparent $^{14}C$ ages of groundwater in Eagle Flat and Red Light Basins are shown in figure 3-7. Average values are shown for wells having multiple samples. The estimated ages are maximum values and are not based on adjustments for the effects of factors that are known to lower the concentrations of $^{14}C$ in groundwater. In this study, $^{14}C$ is not regarded as an indicator of absolute age, but of relative age.

The highest $^{3}H$ and $^{14}C$ values are found in groundwater along the mountain fronts (adjacent to Eagle Mountain) and where bedrock is exposed or covered by a thin layer of basin-fill sediments. The lowest values of $^{14}C$ and $^{3}H$ are found in the deepest groundwater of Eagle Flat and Red Light Basins.

In southeast Eagle Flat Basin, $^{14}C$ ranges from >107 pmc to 2 pmc, and $^{3}H$ ranges from 8 to 0 TU. The highest $^{14}C$ values occur in groundwater of Bean and Millican Hills and the Carrizo and Eagle Mountains. Depth to the potentiometric surface is generally less than 200 ft (<61 m) in these areas, and bedrock is either exposed or covered by a thin layer of basin-fill sediments. The lowest $^{14}C$ values are consistent with apparent ages that range from 23,000 to 31,000 yr. These age estimates are in line with estimates of recharge ages of groundwater in the Hueco Bolson (Fisher and Mullican, 1990) west of Eagle Flat Basin.

In northwest Eagle Flat, $^{14}C$ values range from 80.5 to 1.3 pmc, and $^{3}H$ ranges from 2.0 to 0 TU. The largest concentrations occur in groundwater near the margins of the basin where bedrock of Precambrian or Cretaceous age is exposed. Concentrations of 0 TU are characteristic of deep groundwater near the central area of the flow system.

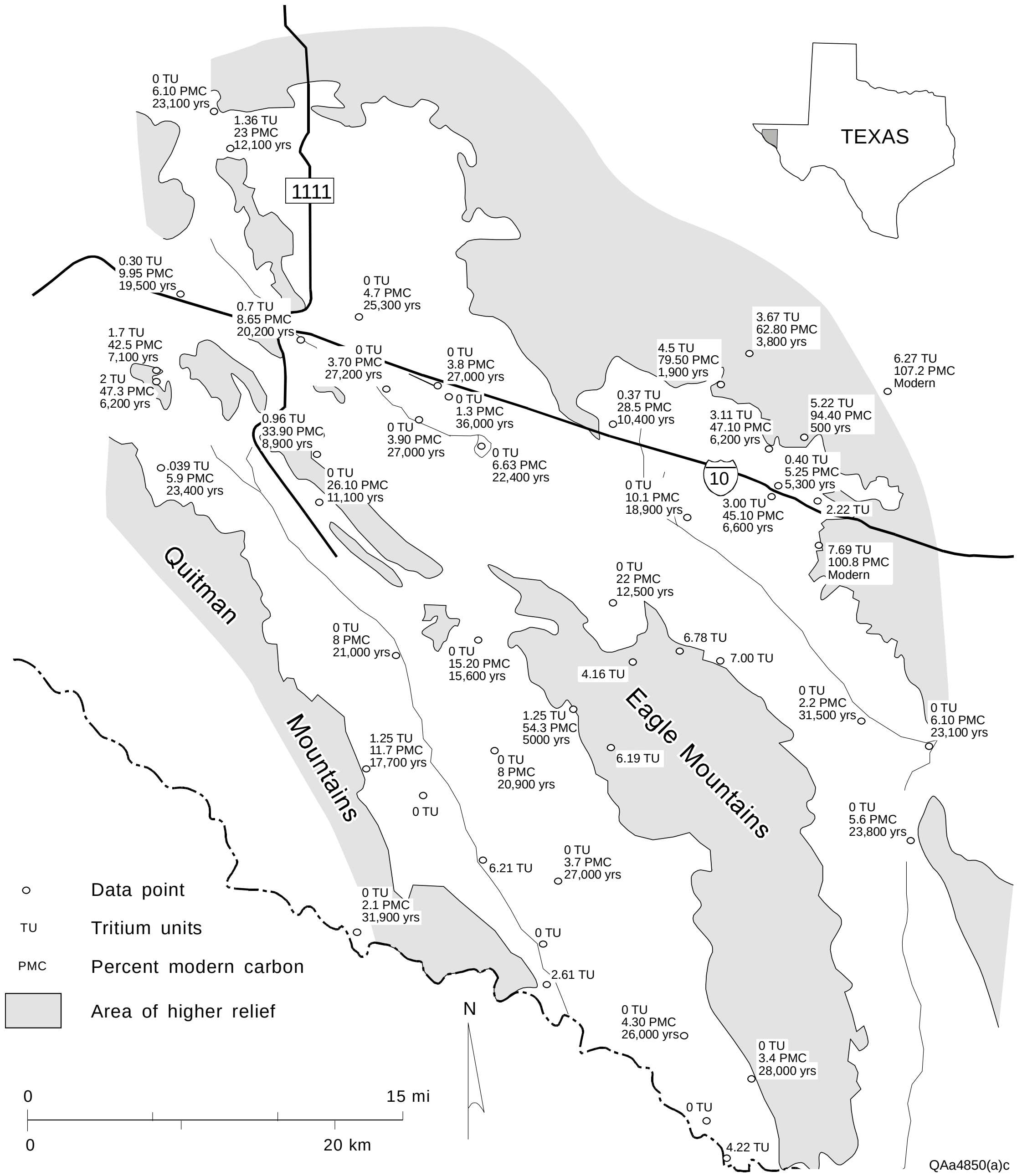


Figure 3-7: Map of groundwater $^3H$ (TU) and $^{14}C$ (pmc) concentrations and uncorrected $^{14}C$ ages in Trans-Pecos Texas.

In Red Light Basin, $^{14}C$ ranges from 54.3 to 2.1 pmc, and $^{3}H$ ranges from 6.9 to 0 TU. The highest $^{14}C$ values are associated with wells in the Eagle Mountains and along the northwest part of Devil Ridge. Spatial distribution is one of sharply decreasing $^{14}C$ with increasing distance from the Eagle Mountains toward the central area of the northwest-oriented basin. For example, $^{14}C$ decreases from 54.3 to 15.2 pmc or less in wells between the upper and middle alluvial fans of the Eagle Mountains (fig. 3-7). Most $^{14}C$ concentrations in this area of the basin are less than 8 pmc.

Six wells within Red Light Basin produce groundwater having measurable $^{3}H$. The largest values occur in groundwater within the upper elevations of the Eagle Mountains and the upper to middle alluvial fans of the southern Quitman Mountains. Two other wells less than 1 mi north of the Rio Grande produce tritiated groundwater (2.61 and 4.22 TU).

# Discussion

Unsaturated-zone studies conducted on the valley floors of the Hueco Bolson and Eagle Flat Basin clearly demonstrate that there is no groundwater recharge in interarroyo settings. Lack of recharge is evidenced by upward water-potential gradients; high chloride concentrations, which result from evapotranspiration, prebomb $^{36}Cl:Cl$ ratios and $^{3}H$ concentrations below the shallow subsurface; and enrichment in stable isotopes of oxygen and hydrogen relative to the local meteoric water line. Similar results were found in alluvial-fan settings in Ward Valley (California), Amargosa Valley (Beatty, Nevada) (Prudic, 1994) and the Nevada Test Site (Tyler and others, 1996). Chloride concentrations at the Beatty site in Nevada decreased to 50 mg/L at depths more than33 ft (≥10 m), indicating an increase in water flux from 0.25 inches/yr (0.01 mm/yr) (3- to 33-ft zone) to 0.08 inches/yr (2 mm/yr) (>33 ft [10 m] depth). The higher water fluxes at depth were attributed to the Amargosa River being more active during the Pleistocene (Prudic, 1994). Low chloride concentrations at depths of 100 to 197 ft (~30 to 60 m [mean 18 mg/L]) at one of the Nevada Test Site profiles were attributed to the site's location at the confluence of alluvial fans, which affected the system's response to higher precipitation during the Pleistocene.

Recharge occurs primarily on the Diablo Plateau and on the mountain fronts of Eagle Flat and Red Light Basins, judging from the $^{3}H$ concentrations in groundwater beneath the Diablo Plateau. These are attributed to recharge along arroyos and depressions where water ponds, as shown by chloride profiles. High recharge rates in the mountains and upland areas of Eagle Flat and Red Light Basins are indicated by high $^{3}H$ concentrations (post-1953 recharge) and high $^{14}C$ concentrations. Tritium concentrations in deep groundwater in these basins are typically 0 TU, whereas $^{14}C$ concentrations are low and indicate ages of 20,000 to 30,000 yr. The low $^{14}C$ values of the deeper groundwater indicate that flow rates within these basins are less than a few feet per year.

Similar results have been found at Yucca Mountain, Nevada: high recharge on the ridge tops and slopes and much lower or no recharge on the valley floors (Flint and others, 2000). The high recharge on the ridge tops and side slopes is attributed to the exposure of fractured rocks near the surface that allows water to migrate rapidly with depth and minimizes evapotranspiration.

# Conclusions

Regional evaluation of recharge in the Trans Pecos demonstrates that it occurs primarily in the Diablo Plateau and in the mountains and adjacent areas in Eagle Flat and Red Light Basins and that no recharge occurs in interdrainage areas on the valley floors. High recharge on the Diablo Plateau is evidenced by tritium in groundwater. The primary recharge mechanism is flow in depressions, such as arroyos through fractured rocks, where soils are thin or absent. Deep migration of water minimizes the amount of time that the water resides in the zone of evapotranspiration. High recharge rates in the mountains and adjacent regions and near bedrock exposures in Eagle Flat and Red Light Basins are shown by high $^{3}H$ and $^{14}C$ concentrations. In contrast, deep groundwater in these basins typically has 0 TU and low concentrations of $^{14}C$, indicating old groundwater (20,000 to 30,000 yr).

Absence of recharge in the interdrainage areas on the valley floors is evidenced by upward water-potential gradients, high chloride concentrations, evapotranspirative enrichment of stable isotopes of oxygen and hydrogen, shallow penetration of bomb-pulse tracers during the past 40 to 50 yr, and radioactive decay of $^{14}C$. The water-potential profiles may reflect upward water movement for the past several thousand years during the Holocene and suggest long-term drying of the sediments. The bulge-shaped chloride profiles in the Hueco Bolson indicate higher water fluxes during Pleistocene times (>10,000 yr), with accumulation of chloride since that time. Some of the chloride profiles in the sand-sheet areas of Eagle Flat Basin also showed higher water fluxes during the Pleistocene; however, most of the chloride profiles in Eagle Flat interdrainage areas did not show any response to Pleistocene climate change, probably because the sediments are too fine grained. The drainage areas of the basin floors did show evidence of higher water fluxes, which is attributed to ponding of water in these systems. The size of the drainage systems is important in affecting water fluxes. Small drainages in the Hueco Bolson with low topographic expressions had much higher chloride concentrations and correspondingly lower water fluxes relative to larger drainages in Eagle Flat Basin (e.g., Blanca Draw). Although water fluxes were expected to be high beneath Grayton Lake playa as a result of ponding, the fine-grained sediments in the playa floor effectively reduce water movement through the floor of the playa.

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# Chapter 4

# Regional Groundwater Flow Systems in Trans-Pecos Texas

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## Introduction

Regional flow systems (Hubbert, 1940; Toth, 1963) are defined as groundwater flow systems that extend both from a regional topographic high to a regional topographic low and beneath overlying local watersheds (catchments) and local flow systems within those watersheds. Regional flow systems are an important aspect in the hydrogeology of Trans-Pecos Texas, which includes the Mesilla, Hueco, Presidio, and Redford Bolsons adjacent to the Rio Grande, the Diablo Plateau/Otero Mesa, the Salt Basin and its southern extensions, the Delaware and Apache Mountains, the Wylie Mountains, the Davis and Barilla Mountains, the Rustler Hills, the Cenozoic Pecos alluvial fill (including the Toyah Basin), and the Stockton Plateau. The area (fig. 4-1) is bordered on the north and east by the Pecos River and is transitional eastward to the Edwards Plateau; it is bordered on the south and west by the Rio Grande. Surface water is essentially nonexistent in this area; both the Pecos River and the Rio Grande are generally now too salty to use, and locally important springs are widely scattered. Consequently groundwater is of paramount importance; groundwater systems are both local and regional. Extensive irrigation districts have been established in the Dell City, Wildhorse Flat, Lobo Flat, Toyah Basin, Balmorhea, Coyanosa, Leon, and Belding areas and along the Rio Grande. As a consequence, there have been a number of hydrological studies of this area, but it is only in the last several decades that an understanding of the regional groundwater flow systems has begun to emerge.

Geological studies of the area are also numerous, and it is the geology (e.g., Barnes, 1976, 1979, 1995a, 1995b; Dickerson and Muehlberger, 1985), which varies from basin-and-range to stable platform and intracratonic basins, that gives the Trans-Pecos groundwater systems their flavor. Uplift, faulting, salt dissolution, and Tertiary volcanism were the significant geological processes in the evolution of this area. Isotopic data also indicate the importance of paleo-climatic processes. Some groundwater is probably Pleistocene in age (Lambert and Harvey, 1987; Kreitler and others, 1987; LaFave and Sharp, 1987; Uliana and Sharp, in press). Modern recharge is low and irregularly distributed both spatially and temporally.

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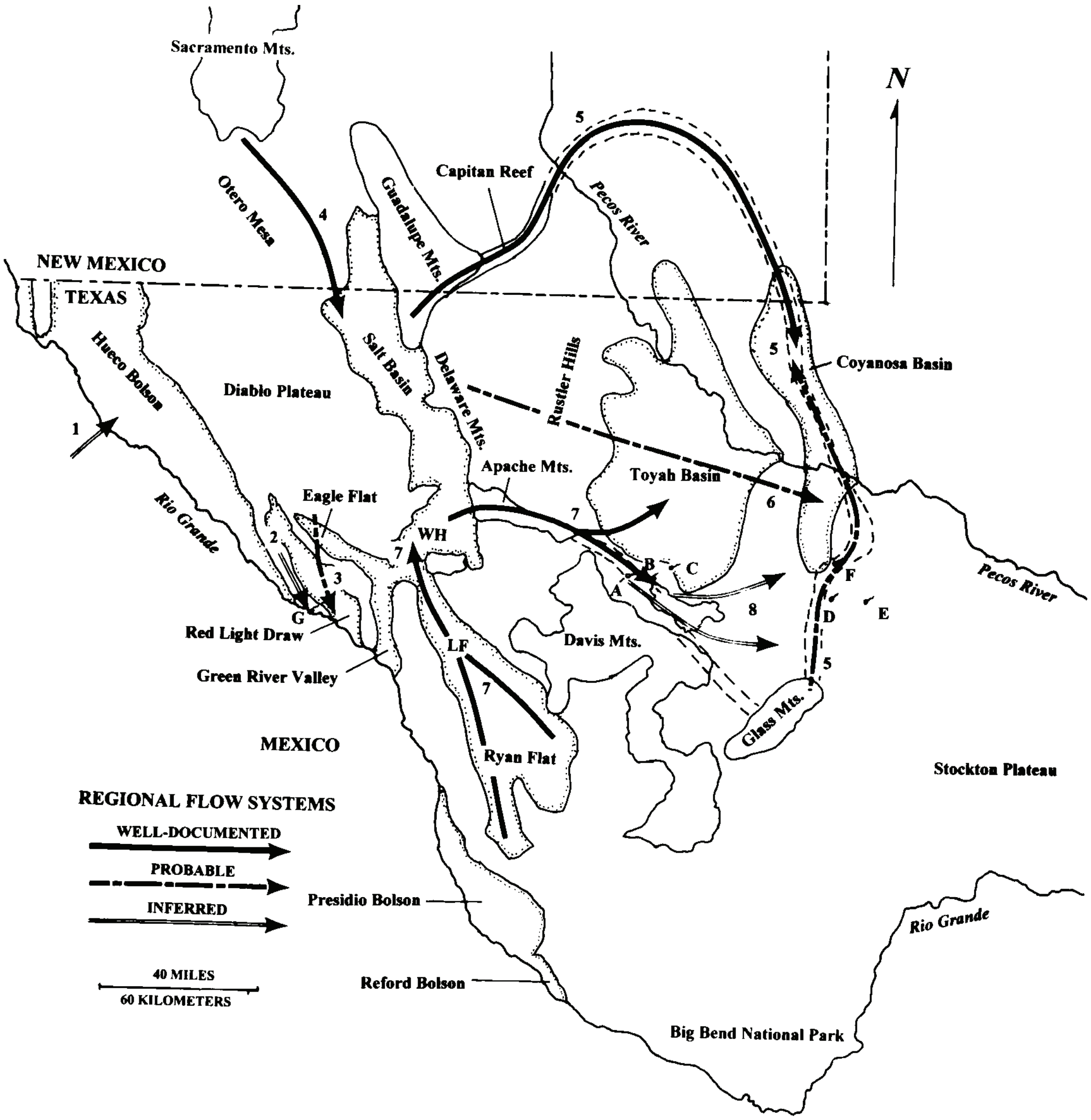


Figure 4-1: Regional flow systems of Trans-Pecos Texas. WH and LF denote Wild Horse Flat and Lobo Flat, respectively, of the Salt Basin. Springs are denoted by letters—A, Phantom Lake Spring; B, San Solomon and Giffin Springs; C, East and West Sandia Springs; D, Leon Springs; E, Comanche Springs; F, Diamond-Y Springs; and G, Indian Hot Springs. A, D, and E no longer flow. The regional flow systems are numbered—1 and 2, the inferred flow systems discharging at the Fabens artesian zone and Indian Hot Springs (G), respectively; 3, Eagle Flat–Red Light Draw flow system; 4, Sacramento Mountains–Dell City flow system; 5, flow systems in the Capitan Reef; 6, eastward flow in the Delaware Basin, perhaps discharging at Diamond-Y Springs (F); 7, the Salt Basin–Toyah Basin–Pecos River system that also feeds Balmorhea Springs (A, B, and C); and 8, speculative eastward extensions of this last flow system.

The most important aquifer systems are (1) the Cenozoic Pecos alluvial aquifer system (especially the Toyah Basin and Coyanosa Basin aquifers); (2) the Salt Basin bolson fill; (3) the bolson aquifers associated with the Rio Grande; and (4) the Permian-Cretaceous carbonate rock aquifers that supply Carlsbad, Dell City and Fort Stockton and that also feed (or fed) the major springs of Pecos (D and E on fig. 4-1) and Reeves (A, B, and C on fig. 4-1) Counties. All of these aquifer systems possess important regional flow components. Numerous other minor aquifers are in the area; they are mostly for domestic or livestock use. Among these, the volcanic rock aquifers of the Davis Mountains, designated the McCutcheon aquifer by Hart (1992), have become particularly important as that area has developed.

Regional flow systems become important in areas where recharge is limited, where there is a significant regional topographic gradient, and where high-permeability rocks exist at depth. These conditions are all met in Trans-Pecos Texas. In the following, the regional flow systems of Trans-Pecos Texas are discussed, with an emphasis on those that supply (or in the past supplied) the major springs of Pecos and Reeves Counties that also possess (or in the past possessed) unique biota, including Federally listed endangered species.

# Hydrostratigraphy

The oldest hydrogeologically significant rocks are Permian, although Pennsylvanian through Precambrian rocks are present at depth (McMahon, 1977). Permian strata are divided into four series—Wolfcampian, Leonardian, Guadalupian, and Ochoan. In the area of the Delaware Basin, these units can be subdivided into three hydrogeologic facies (Hiss, 1980; Nielson and Sharp, 1985)—the high-permeability Guadalupian Series shelf margin/reef; the variably permeable (fracture-dependent) Leonardian and Wolfcampian Series shelf facies, which crop out in the Diablo Plateau and Otero Mesa (in New Mexico); and the low-permeability Guadalupian and Ochoan\Series basin facies rocks that fill the Delaware Basin. Shelfal facies rocks form the aquifer that serves the Dell City irrigation district and ranches throughout the Diablo Plateau and Otero Mesa. Overlying and east of the shelf is the Capitan Reef facies consisting of the Capitan and underlying Goat Seep limestones of Guadalupian age. Carlsbad Cavern is a prime example of this highly permeable reefal facies that extends circumferentially around the Delaware Basin (Adams, 1944; Hiss, 1980). The reefal facies exerts a major control on regional ground-water flow systems. The Guadalupian and Ochoan basinal facies, located east of the Guadalupe Mountains in the center of the Delaware Basin, are generally low in permeability, except where exposed at the surface. Water quality is also generally poor. Included in the Ochoan Series are the Castile, Salado, and Rustler Formations. The Castile Formation is composed of gypsum, calcareous anhydrite, halite, and subordinate limestone (Ogilbee and others, 1962); it possesses numerous karstic features (Olive, 1957). The Rustler Formation produces poor-quality water that is used for irrigation and livestock. However, basinal facies rocks form an eastward-flowing regional flow systems, and the Rustler probably discharges at Diamond-Y Springs in Pecos County (Boghici, 1997). Dissolution of these evaporite rocks also created swales in which the Cenozoic Pecos alluvial aquifers, such as the Toyah Basin aquifer, were deposited (LaFave, 1987; Ashworth, 1990).

Overlying the Permian is the Triassic Dockum Group that yields good-quality water in the southeastern part of the Toyah Basin. Overlying the Dockum are Cretaceous Comanchean and Gulfian carbonates and sandstones. Hydrogeologically important units include the Cox Sandstone and the Edwards, Georgetown, and Boquillas limestones. The Cox Sandstone is an aquifer on the southwestern flank of the Apache Mountains and just north of the Wylie Mountains in Wild Horse Flat, as well as along the Pecos River in the eastern part of the Trans-Pecos. The Cretaceous carbonates provide water to ranches north of the Davis Mountains and water for irrigation and municipal use in Pecos County. Paleokarst features are present in these formations. The regional flow systems of the Trans-Pecos Texas are either contained within or are strongly influenced by these fractured and karstified Cretaceous and Permian rocks.

Rocks of Tertiary age form the Davis Mountains and include the McCutcheon volcanics that consist of interbedded lava flows, tuffs, and nonmarine sedimentary rocks. Generally these yield minor quantities of water for domestic and livestock use. The Davis Mountains rest nonconformably over Cretaceous and, perhaps, Permian sedimentary rocks. The volcanic rocks have been undergoing slope retreat in the Cenozoic and were present originally farther to the north and east than at present (Halamicek, 1951).

Quaternary and Tertiary alluvial sediments that veneer much of the study area and attain thicknesses of over 2,400 ft in the Salt Basin (Gates and others, 1980) and over 1,500 ft in the Toyah Basin (Maley and Huffington, 1953) are of great hydrogeological significance. These sediments are dominantly clastic, although gypsum and caliche are probably present (Ogilbee and others, 1962). The alluvium provides water for irrigation in Hueco Bolson, Redford Bolson, Wild Horse Flat, Lobo Flat, and the Toyah and Coyanosa Basins, and the undifferentiated alluvium/Permian limestone system in Wild Horse Flat supplies water to the towns of Van Horn and Sierra Blanca (farther to the west). Ryan Flat has been considered as a potential water source for El Paso. In many parts of the region, including the vicinity of Balmorhea, some shallow wells produce small amounts of water from undifferentiated alluvium/limestone systems, but most of the water needs in Balmorhea are provided by spring flow, which issues from fractures and solution cavities in the Cretaceous units. The bolson aquifers, Hueco Bolson, Red Light Draw, Green River Valley, Eagle Flat, Presidio Bolson, and the Redford Bolson, provide fresh to brackish waters to municipalities and to the public for domestic, livestock, and irrigation uses.

# Structural Setting

The northern Trans-Pecos is a transitional area from the Basin and Range province, exemplified by the Salt Basin graben, to the Permian Delaware Basin, a stable cratonic feature. The Delaware Basin contains more than 20,000 ft of Paleozoic sediments and is bounded by Capitan Reef rocks that are exposed in the Guadalupe Mountains, Apache, and Glass Mountains (fig. 4-1). The reef trend continues north-northeastward into New Mexico and southeastward in the subsurface. Basin-and-range style tectonics down-dropped the Salt Basin in the Cenozoic, and fault movement has continued to the present (Goetz, 1977, 1980, 1985). Important second-order structural features are the major fault and fracture sets in the carbonate units and the clastics-filled dissolution basins of the Cenozoic Pecos alluvial aquifer system and the bolson fills. For instance, LaFave and Sharp (1987) and Uliana and Sharp (in press) demonstrated how the regional flow system between the Salt and the Toyah Basins correlate with major fault/fractures systems in Permian/Cretaceous carbonates. Mayer and Sharp (1998) demonstrated a similar trend along Otero Break connecting the regional flow system between the Sacramento Mountains in New Mexico and the Dell City irrigation district. In the Salt Basin, major flexures/fracture systems correlate with groundwater divides (Neilson and Sharp, 1985). Finally, van Broekhoven and Sharp (1998) showed how recharge zones in Ryan/Lobo Flat correlate with fracture systems in the surrounding ranges. The Stocks Fault, which bounds the north-northeastern flank of the Apache Mountains, is one of a set of east-trending brittle fractures that are evident north of the Davis Mountains (LaFave and Sharp, 1987) and may, possibly, also be present beneath the Tertiary volcanics. DeFord (personal communication, 1986) and Wood (1965) stated that the large throw of the Stocks Fault, which abuts the Apache Mountains on the north, is the result of subsurface dissolution of Delaware Basin evaporites. The Rounsaville Syncline and Star Mountain Anticline parallel the Stocks Fault to the southeast. The large springs near Balmorhea are located near the syncline (A, B, C on fig. 4-1). The Toyah Basin and other units of the Cenozoic Pecos alluvial aquifer system (Ashworth, 1990) were created by late Tertiary and Quaternary dissolution of Permian evaporite-salts of the Castile and Salado Formations and anhydrite and gypsum from the Rustler Formation.

# Regional Flow Systems

Although there was no comprehensive study of area groundwater systems prior to the large-scale municipal and agricultural development after World War II, it is possible to approximately infer some predevelopment potentiometric surfaces from a series of reports and unpublished data in the files of the Texas Water Commission. These include reports on the Rio Grande bolsons (Gates and others, 1980; White and others, 1941), the Toyah Basin in 1940 (Lang, 1943; see also LaFave and Sharp, 1987), the Salt Basin including Wild Horse, Lobo, and Michigan Flats in the late 1950's (Hood and Scalapino, 1951; White and others, 1980; Nielson and Sharp, 1985), and the Dell City area (Scalapino, 1950; Mayer, 1995). Data on the Diablo Plateau were compiled by Kreitler and others (1987) and Mullican and Senger (1992). The recent regional study by Richey and others (1985) presents potentiometric surfaces in the Cenozoic alluvium, the Santa Rosa aquifer, the Rustler Formation, and the Capitan Reef aquifer. Groundwater in the Diablo Plateau and Otero Mesa has not been extensively developed, except near Dell City, so its present potentiometric surface is probably similar to that of predevelopment conditions. Flow systems in the Rustler Formation are still not well delineated.

Figure 4-1 is a generalized depiction of the regional flow systems in Trans-Pecos Texas. Those flow systems have been designated as well documented because the flow systems are consistent with potentiometric, geologic, structural, geochemical, and isotopic data. Probable regional flow systems have fewer data sets confirming them. They are most consistent with available data, but other interpretations are possible. Inferred regional flow systems are more speculative. They feed artesian wells or hot springs and are inferred from limited data.

There are three regional predevelopment discharge areas—the Rio Grande, the northern and middle sections of the Salt Basin, and the Pecos River on the northeastern boundary of the study area. The Hueco, Presidio, and Redford Bolson groundwater systems discharge to the Rio Grande, as do those of Red Light Draw and the southern part of the Green River Valley. Most of the other systems discharge to the Pecos River. The Salt Basin is divided into three flow systems: the northern section, the middle section, and the southern section, including Wild Horse, Lobo, and Ryan Flats. There are playas in the northern and middle sections of the basin that demonstrate evaporative discharge. This inference is supported by water-chemistry studies (Gates and others, 1980; Boyd, 1982; Chapman, 1984; Mayer and Sharp, 1998) that show increasing salinity in the direction of flow. In the playas, gypsum and halite are precipitated from groundwater in the capillary fringe. On the eastern margin of the northern section and on the eastern and western margins of the middle section, ground-water recharge occurs by influent streams that cross alluvial fans and by precipitation on the permeable shelf margin carbonates. On the eastern flank of the Salt Basin, slightly brackish water can be found at depths of more than 2000 ft in Capitan Reef rocks (Reed, 1965). In contrast, near Dell City, on the western flank, there is little topographic relief on the western margin of the graben that merges gradually with the Diablo Plateau and Otero Mesa. Most of the other systems discharge to the Pecos River. These include the southern section of the Salt Basin, the Rustler Hills, Ryan and Lobo Flat, and the Stockton Plateau.

## Rio Grande Bolsons

The flow systems in Hueco, Presidio, and Redford Bolson aquifers are (or were, prior to heavy groundwater pumpage) local flow systems that discharge to the Rio Grande. Recharge was concentrated on the basin margins, especially the proximal portions of alluvial fans, major fracture systems, and along the more perennial streams. However, the Fabens artesian zone and Indian Hot Springs (G on fig. 4-1) represent discharge from deeper artesian, regional flow systems (Kreitler and Sharp, 1990). These flow systems are designated as 1 and 2, respectively, on figure 4-1. The Fabens system is assumed to recharge in Mexico (Gates and others, 1980); the flow system at Indian Hot Springs could originate in either Texas (as suggested in fig. 4-1) or Mexico. Gabaldon (1991) suggested that deep brackish waters in the Presidio Bolson could arise from a similar regional system; an alternative explanation is that there are evaporite deposits at depth in this bolson.

## Eagle Flat–Red Light Draw

Darling (1997) and Hibbs and others (1998) investigated the flow systems under Eagle Flat where a low-level radioactive waste repository had been proposed. There is no natural groundwater discharge in Eagle Flat, but there is flow into it. The water table is deep (>400 ft), and karstic/fractured carbonate rocks are present at depth. A deep flow system (3 on fig. 4-1) exists in the carbonate rocks from northwest Eagle Flat beneath the Devil Ridge to discharge into Red Light Draw and into the Rio Grande.

## Diablo Plateau–Otero Mesa–Dell City

Scalapino (1950), the Groundwater Field Methods Class (1992b), Sharp and others (1993), Ashworth (1995), Mayer (1995), and Mayer and Sharp (1998) reported on the flow systems associated with the Dell City irrigation district. Irrigation pumpage near Dell City has created a cone of depression, and water quality has deteriorated from a TDS range of 1,100 to 1,800 mg/L to a range of 3,000 to 5,000 mg/L. It is unclear how much of the deterioration is due to irrigation return-flow and how much to a reversal of flow and salt-water intrusion from the basin, although return-flow is probably the major cause. Kreitler and others (1987) documented flow to the northeast in the Diablo Plateau, but hydrologic and geochemical data (Mayer and Sharp, 1998) indicate that the bulk of the flow comes from the area where the Sacramento River sinks and then flows southeastward along the Otero Break to Dell City. Mayer (1995) mapped fracture intensities and orientations in Otero Mesa. These are consistent with a zone of higher permeability rocks along the flow system (designated by 4 on fig. 4-1). A plume of relatively fresh water marks this path. Several paleolake basins exist along this regional flow path (Hawley, personal communication. 1996). In the southern margins of the Diablo Plateau, flow into the Hueco Bolson was documented by Mullican and Senger (1992) and shown as the proximal end (start) of flow path 3.

## Delaware Basin and Capitan Reef

Hiss (1980) documented the regional flow system in the Capitan Reef aquifers that flows from its outcrops along the Texas-New Mexico border northeastward along its trends and southeastward from the Apache Mountains. The flow paths (5 on fig. 4-1) follow the high-permeability reef facies and are enhanced by fracture systems that subparallel the reef trends (Uliana, 2000). The uplift of the western side of the Delaware Basin created a regional topographic gradient, and a regional groundwater flow system from west to east was hypothesized by Hiss (1980), Mazzullo (1986) and Richey and others (1985). This regional flow has been suggested as a process for hydrocarbon migration and mineralization in the deeper sections of the Delaware Basin. This flow system is designated as 6 on figure 4-1. Boghici (1997) suggested that discharge from deep Rustler Formation waters along a fault system in Pecos County is responsible for the flows at Diamond-Y Springs (F on fig. 4-1). The southern part of the Capitan reef system (5) and the regional flow system (6) are designated as probable herein because of the potential effects of petroleum-production-related depressurization.

## Ryan Flat–Lobo Flat–Salt Basin–Apache Mountains–Balmorhea–Toyah Basin–possible extension to Pecos County

The Ryan Flat–Lobo Flat–Salt Basin–Apache Mountains–Balmorhea–Toyah Basin is the longest regional flow system in Trans-Pecos Texas; it is designated as flow system 7 on figure 4-1. It extends from Ryan and Lobo Flats, which are southern extensions of the Salt Basin, and the groundwater boundaries are close to the Rio Grande. This flow system collects groundwater from the northern Green River Valley and southeastern Eagle Flat and flows into Wild Horse Flat in the Salt Basin near Van Horn, Texas. Recharge to the southern extensions occurs by infiltration at the proximal portions of alluvial fans and from subsurface flow in major fracture systems in Sierra Vieja on the western flank of Lobo and Ryan Flats (Darling and others, 1995; Darling, 1997, Hibbs and others, 1998; and van Broekhoven and Sharp, 1998). The magnitude of flow from Lobo Flat is uncertain, and a steepening of the water table gradient south of Van Horn is coincident with east-trending faults (Hay-Roe, 1958; Twiss, 1959; Sharp, 1989).

In Wild Horse Flat, additional recharge is gathered from the alluvial fans on the western side of the basin. This gathering is confirmed by isotopic analyses of the groundwater (Uliana and Sharp, in press) that show a Precambrian Sr-isotopic signature obtained from flow through these fans. The fans are largely derived from Precambrian rocks in the Carrizo Mountains, Beach Mountain, and the southern Sierra Diablo. Recharge also occurs from precipitation along ephemeral streams, such as Wild Horse Creek, and perhaps from irrigation return-flow. In Wild Horse Flat, the predevelopment water table was about 100 ft beneath the surface (Gates and others, 1980), in contrast to the evaporative-discharge playa systems in the northern portions of the Salt Basin. The main predevelopment, regional flow occurred eastward toward the Toyah Basin through shelf margin (reef) facies rocks of the Apache Mountains, which serve as a drain for Wild Horse Flat.

The structural setting is also conducive to interbasin flow not only because the rocks in the Apache Mountains are permeable but also because the trend of extensive, regional fractures is roughly east (LaFave and Sharp, 1987; Uliana, 2000). These rocks are highly anisotropic, and the direction of greatest permeability is subparallel to the Stocks Fault that is the northern border of the Apache Mountains. LaFave and Sharp (1987) concluded on the basis of regional geology and geochemistry that a significant portion of the flow of the Balmorhea Springs discharged from a regional aquifer system, recharged in part from interbasin flow through the Apache Mountains. This finding was reconfirmed by Uliana (2000), who used a much larger geochemical database augmented by Sr, $^{2}H$ (or D), and $^{18}O$ isotopic analyses. Balmorhea Springs (A, B, C on fig. 4-1) issue from orifices at elevations of about 3,300 ft. This elevation provides a reasonable hydraulic gradient of $10^{-3}$ to $10^{-4}$ between the springs and Wild Horse Flat. Spring flows enter Toyah Creek, which flows across the Toyah Basin but flows into the Pecos River only after major storms.

The regional flow system also discharges directly into the Toyah Basin aquifer, which produces groundwater for its extensive irrigation areas from the Cenozoic alluvium and, in the eastern section, from undifferentiated alluvium and Cretaceous limestones. In addition to the component of interbasin regional flow, recharge to the Toyah basin aquifer occurs in the Rustler Hills and from the ephemeral streams that drain the Davis and Barilla Mountains. LaFave (1987) noted that the aquifer produces Cl-dominated-facies water in the southwest and central portions of the Toyah Basin, whereas $SO_4$-dominated-facies water is produced from the western and northwestern portions. The Cl-facies is virtually identical chemically to groundwaters produced from Capitan Reef aquifers and from Balmorhea Springs. This hypothesis is in concurrence with the reports of Hiss (1980), Mazzullo (1986), and Richey and others (1985), although these authors did not address the possibility that regional flow recharges the Toyah Basin aquifer. The $SO_4$ facies indicates its origin in the Ochoan rocks of the Rustler Hills. It is not known whether these waters recharge the Toyah Basin aquifer chiefly by subsurface flow or by infiltration along the many draws that drain eastward from the Rustler Hills. Finally, on the eastern margins of Toyah Basin, better-quality (>1,500 mg/L) water is obtained. Recharge in these areas is by precipitation and by infiltration from waters in draws draining the Barilla and Davis Mountains.

The effects of humans on the Toyah Basin aquifer have been significant. Irrigation pumpage increased rapidly after 1945. Many springs in the area have since ceased to flow (Brune, 1981). Irrigation pumpage from the Toyah Basin lowered water-table elevations and created a cone of depression. Thus, pumpage totally altered the regional-flow-system discharge zone from the Pecos River to irrigation wells within the Toyah Basin (LaFave and Sharp, 1987; Schuster, 1997; Boghici, 1999). Water quality has remained relatively constant, but a perched water table has developed about the City of Pecos, with salinities of over 8,000 mg/L (Groundwater Field Methods Class, 1990a, 1992a). The Groundwater Field Methods classes (1990b, 1992c, 1995, 1996) found water-level declines near Balmorhea Springs of about 20 ft with respect to the 1932 data (White and others, 1938). Recent declines of pumpage for irrigation because of economic conditions has allowed partial recovery of water levels, but it seems doubtful that predevelopment conditions will be achieved.

Eastward extensions of this regional flow system (designated by 8 on fig. 4-1) were suggested by Boghici (1997) and Uliana (2000). Boghici's numerical model of Pecos County flow systems required additional subsurface recharge from northwest of the Glass Mountains. Uliana found that chemical and isotopic data are consistent with continued geochemical evolution of waters along the trend of the Stocks Fault/Rounsaville syncline trend consistent with the eastward extension of the flow system. These extensions, however, are still somewhat speculative.

### Other regional flow systems?

Other, yet-undocumented, regional flow systems may exist in Trans-Pecos Texas. For example, the flow systems in Mesozoic and Paleozoic units beneath the Davis Mountains may discharge to the Pecos River or, in part, south toward the Rio Grande. Permeable carbonate rocks are also present at depth toward the Big Bend area so that regional flow systems would be expected there, and transboundary (USA-Mexico) regional flow systems may exist. The delineation of these systems should be interesting.

## Discussion

Regional groundwater flow systems are a major hydrogeologic characteristic of Trans-Pecos Texas. Geologic processes of faulting, folding, and dissolution in semiarid Trans-

Pecos Texas have created the controlling framework for regional groundwater flow systems. There exist three regional discharge areas—the Rio Grande, the northern and middle sections of the Salt Basin, and the Pecos River. The feature common to all eight flow systems depicted on figure 4-1, whether well-documented, probable, or inferred, is the presence of Permian carbonate rocks or Cretaceous carbonate rocks in close proximity to Permian units. The carbonate rocks have been fractured by a variety of tectonic episodes, including the Laramide orogeny, Basin-and-Range extension, and subsidence caused by dissolution of underlying evaporite deposits. The region remains tectonically active, as evidenced by recent seismic activity. In some areas, this was followed by very effective karstification—the Capitan reef aquifers may be some of the Earth's most permeable rocks. Coupled with the low rainfall and consequent groundwater recharge, regional flow systems have become an integral part of the regional hydrogeology. With greater recharge, local flow systems would be more dominant and, perhaps, they were in the past as is suggested by the old apparent ages of waters in some of these regional flow systems.

Yet unknown and apparently unstudied is the evolution of these regional flow systems and the effects on spring flows, desert ecosystems, and potential hydrocarbon and mineral deposits. Understanding of the regional flow systems is also critical to the development and sustainability of water resources in this region.

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# Chapter 5

# Regional Ecology and Environmental Issues in West Texas

Gary P. Garrett[1] and Robert J. Edwards[2]

## Introduction

The Chihuahuan Desert of Mexico, Texas, and New Mexico contains a wide variety of habitats and many uniquely adapted plants and animals. The limited aquatic habitats of this ecosystem have undergone substantial modifications in the last hundred years, including: reduced water quality, diversion of surface water, overdrafting of groundwater, channelization, impoundment, and extensive introduction of nonnative species (Miller and Chernoff, 1979; IBWC, 1994; Propst and Stefferud, 1994; TNRCC, 1994; Lee and Wilson, 1997; Propst 1999; Edwards and others, in press).

One of the most heavily impacted habitats is the desert springs and their associated wetland habitats. In the American Southwest and northern Mexico these are known as *ciénegas*. These ecosystems were seldom damaged on purpose; put simply, water is rare in the desert and people want it for a variety of uses. The ways in which ecosystems have been destroyed include grazing and watering livestock, draining to move water more efficiently to agricultural fields, and over-pumping of aquifers. Impacts from these modifications are only now being documented, and few baseline data exist concerning the ecological requirements for most of the aquatic species.

Approximately half of the native fishes of the Chihuahuan Desert are threatened with extinction or are already extinct (Hubbs, 1990). Documented extinctions from this area include the Maravillas red shiner (*Cyprinella lutrensis blairi*), the phantom shiner (*Notropis orca*), the Rio Grande bluntnose shiner (*Notropis simus simus*), and the Amistad gambusia (*Gambusia amistadensis*) (Miller and others, 1989). Extirpations include the Rio Grande shiner (*Notropis jemezanus*) from the New Mexico portion of the Rio Grande (Propst and others, 1987) and the Rio Grande silvery minnow (*Hybognathus amarus*) and the Rio Grande cutthroat trout (*Oncorhynchus clarki virginalis*) and blotched gambusia (*Gambusia senilis*) in Texas (Bestgen and Platania, 1988, 1990; Hubbs and others, 1991). Endemic species other than fishes are also being lost from this area (Howells and Garrett, 1995).

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A number of species inhabiting the area have insufficient information about their status, but enough is known to raise concern among biologists. Not all have legal status (yet), and ideally we can find solutions before legal protection is needed. All serve as indicators of ecosystem integrity and for the quality of our lives.

# Extirpated Species

Rio Grande silvery minnow (*Hybognathus amarus*) is a federally endangered species that once occurred throughout the Rio Grande basin from northern New Mexico to Brownsville. There is now only a small population in the middle Rio Grande of New Mexico. This sort of range reduction, from common to near extinction, is indicative of much larger problems in the watershed. Recovery efforts include attempts to insure flow and water quality, as well as possible reintroductions into selected sites.

Rio Grande cutthroat trout (*Oncorhynchus clarki virginalis*) was apparently a Texas inhabitant until the late 1800's (Garrett and Matlock, 1991). Remnant populations were reported in the Guadalupe and Davis Mountains, and officers at Fort Hudson reported them in San Felipe Springs. Reduced water flows and introduction of rainbow trout (*O. mykiss*) led to their ultimate extirpation from Texas.

Blotched gambusia (*Gambusia senilis*) is listed as threatened by Texas, although it has apparently been extirpated since the filling of Amistad Reservoir. It originally occurred in springs of the Devils River arm of the reservoir but was wiped out when this habitat was inundated. The species is still found in Río Conchos and is a protected species in Mexico.

# Species of Concern

Sturgeon (*Scaphirhynchus platorynchus*) is known to occur in the Red River below Lake Texoma Reservoir. There are also historic records from the Rio Grande at Albuquerque, New Mexico (Cope and Yarrow, 1875). There are anecdotal accounts of occurrence in the Rio Grande in the vicinity of Big Bend National Park, but numerous collections have not yielded specimens. Some think that this animal may actually be an Atlantic sturgeon (*Acipenser oxyrhynchus*). Further surveys of the Rio Grande are planned, but, even if it is found, many questions will remain. Conservation decisions will hinge upon whether a viable population exists or whether there are simply some very old "relict" sturgeon cut off from the Gulf of Mexico by dams.

Mexican stoneroller (*Campostoma ornatum*) occurs throughout the Chihuahuan Desert, but populations are fractured and appear to be reduced in abundance. This reduction is primarily due to pollution and overpumping from aquifers. Determining solutions to problems for this species will also be beneficial to water issues for all organisms in this ecosystem. An ongoing survey of Chihuahuan Desert fishes may provide needed information to help conserve this species. A small, but secure population exists in Cienega Creek in the Big Bend Ranch State Park.

Proserpine shiner (*Cyprinella proserpina*) is a State threatened species. It is closely related to the more common red shiner (*C. lutrensis*) but has a range restricted to the Pecos and Devils Rivers; the San Felipe, Las Moras and Pinto Creeks; and the Río San Carlos in Mexico.

Devils River minnow (*Dionda diaboli*) is a Federally threatened species. It is now only found in the Devils River and San Felipe and Sycamore Creeks. The geographic location and historic stability of the Devils River have sustained a number of indigenous organisms. Because of limited access, the river has not been well studied. However, collections in the past decade by Garrett and others (1992) and others indicate a diminution in abundance of several flowing-water species, particularly the Devils River minnow. In 1953, a collection at Baker's Crossing showed the Devils River minnow to be the fifth-most-abundant fish species there. In the mid-1970's Harrell (1978) found it to be the sixth-most-abundant fish in the river. By 1989, collections from 24 locations throughout the range of the minnow yielded a total of only 7 individuals. Only one fish was obtained from Baker's Crossing, and no more than two were obtained at any site. In 1979, the Devils River minnow made up 6 to 18 percent of the *Dionda* population at the Headsprings area of San Felipe Creek. In 1989, none were present. Very little is known of the Devils River minnow, but some problem areas are apparent. Habitat loss has occurred by extirpation of the Las Moras Creek population, minimal flows in Sycamore Creek, and inundation of the lower Devils River by Lakes Walk and Devils and, ultimately, Amistad Reservoir. Many springs in the area have diminished flows, and some have totally stopped (e.g., Willow Springs, Beaver Springs, Juno Springs, and Dead Man's Hole), thus reducing the overall length of the Devils River, as well as the quantity of water flowing in it. Many of the perennial streams (Gray, 1919) of the area no longer flow. USGS data from the Pafford Crossing gauging station reveal a general decrease in daily mean discharge for the period between the study by Harrell (1978) and that of Garrett and others (1992). Brune (1981) attributed the reduced spring flows in this area to heavy pumping from wells and overgrazed soils with lowered capacity to absorb water and thus recharge aquifers.

Manantial roundnose minnow (*Dionda argentosa*) is a species closely related to the Devils River minnow. It is limited to the Devils River and San Felipe and Sycamore Creeks. It is not legally protected, and hopefully efforts to recover the Devils River minnow will also benefit the manantial roundnose minnow.

Blue sucker (*Cycleptus elongatus*) is a State threatened species. It is a big river fish found throughout the Mississippi Basin and large streams of Texas. Because these systems suffer from impoundment, pollution, and reduced water flows, abundances have decreased. Those in the Rio Grande may be a different, as yet undescribed, species.

West Mexican redhorse (*Scartomyzon austrinus*) is closely related to the common gray redhorse (*S. congestus*). It occurs from Pacific coast drainages in Mexico to the mid-Rio Grande in Texas. Those in the Rio Grande may also be a new, undescribed species.

Rio Grande blue catfish (*Ictalurus furcatus* ssp.) is very likely a unique subspecies of blue catfish. We have taken it only in the Rio Grande from Presidio to Laredo. The unique spotting pattern and head shape of this fish make it different from other blue catfish. Unfortunately very little else is known of this creature.

Headwater catfish (*Ictalurus lupus*) originally ranged throughout the streams of the Edwards Plateau and Pecos River and Rio Grande. It is now uncommon and can now be found only in subsegments of the Pecos River and Rio Grande.

Leon Springs pupfish (*C. bovinus*) were first collected in 1851 by the U.S. and Mexican Boundary Survey at Leon Springs (Baird and Girard, 1853). Leon Springs no longer exists because of impounding, inundation, and groundwater pumping (Hubbs, 1980). *Cyprinodon bovinus* was extirpated from Leon Springs as early as 1938 (Hubbs, 1980) and presumed extinct (Hubbs, 1957). In the early 1900's Leon Springs flowed at approximately 20 cfs, but heavy groundwater pumping reduced the flow to 0 by 1962 (Echelle and Miller, 1974). Today the only water source for the species is the Diamond-Y Springs and outflows north of Fort Stockton, Pecos County, Texas.

Conchos pupfish (*Cyprinodon eximius*) is a State threatened species occurring in the Conchos basin of Mexico and Rio Grande from Alamito Creek to the Devils River. Populations in the Devils River were nearly eliminated in the 1950's, but recovery efforts have led to a thriving population in the Devils River State Natural Area. Although the reestablished population in Dolan Creek is thriving (Hubbs and Garrett, 1990), most of the other Rio Grande tributary populations are sparse.

Comanche Springs pupfish (*C. elegans*) originally inhabited two isolated spring systems approximately 90 km apart in the Pecos River drainage of western Texas (Baird and Girard, 1853). The type locality, Comanche Springs inside the city limits of Fort Stockton, is now dry, and the population at this locality is extinct. The other population is restricted to a small series of springs, their outflows, and a system of irrigation canals interconnecting Phantom Lake Springs (located in easternmost Jeff Davis County, Texas), San Solomon Springs, Giffin Springs, and Toyah Creek near Balmorhea, Reeves County, Texas. The habitat of Comanche Springs pupfish has been markedly altered into an irrigation network of concrete-lined canals with swiftly flowing water and dredged, earth-lined laterals. Waters from Phantom Lake Springs originally emerged from a cave and formed a ciénega that drained back into a cave. Water is now captured in an irrigation canal as it emanates from the cave. Water from San Solomon and Giffin Springs flows into additional irrigation systems, some of which is stored in an irrigation supply lake known as Lake Balmorhea. This habitat is highly impacted, ephemeral, and very dependent upon local irrigation practices and other water-use patterns. For the most part, the irrigation canals provide little suitable habitat for *C. elegans*. The species is wholly dependent upon failing spring flows in the area and suffers as well from threats of hybridization and competition with introduced sheepshead minnow (*C. variegatus*).

Pecos pupfish (*C. pecosensis*) once occurred throughout the Pecos River in New Mexico and Texas. It now suffers from habitat degradation and hybridization with the introduced *C. variegatus*.

Big Bend gambusia (*Gambusia gaigei*) was described in 1929 (Hubbs, 1929) on the basis of specimens taken from a spring-fed slough across from Boquillas, Mexico, in Brewster County, Texas (now Big Bend National Park). Big Bend gambusia went under an extreme genetic bottleneck approximately 50 yr ago, when their only habitat was contaminated with *G. affinis*. All *G. gaigei* are descendents of three individuals taken in 1956 (Hubbs and Broderick, 1963). At present, several thousand Big Bend gambusia inhabit two spring-pool refugia and a spring-fed drainage ditch. Smaller populations also occur in the presumed original habitat and the spring's outflow channel. The limited quantities of warm spring waters available, park campground development in the area, and the loss of the species' natural habitats in Boquillas Spring and Graham Ranch Warm Springs further limit this species' recovery.

Pecos gambusia (*Gambusia nobilis*) was described by Baird and Girard (1853) from Leon and Comanche Springs, Pecos County, Texas. Leon Springs was later designated the type locality (Hubbs and Springer, 1957). The species is endemic to the Pecos River basin in southeastern New Mexico and western Texas. At present, the species is restricted to four main areas, two in New Mexico and two in Texas. Where suitable habitats exist, Pecos gambusia populations can be dense, ranging from 27,000 to 900,000 individuals in the isolated environments in which they occur (Bednarz, 1975). Pecos gambusia face severe threats from spring-flow declines and habitat modifications throughout their range and from competition with *G. geiseri*.

Rio Grande darter (*Etheostoma grahami*) is a State threatened species. It is found in the lower Pecos and Devils Rivers, San Felipe and Sycamore Creeks, and the intervening Rio Grande. It is also part of the unique fauna of the region, and efforts to protect the Devils River minnow should also help this species.

# Hope for the future

The Texas Parks and Wildlife Department is working with Federal, State, local agencies, and especially private landowners to resolve endangered species problems. With 97 percent of the land in Texas being privately owned, this is the only way we can achieve long-term benefits.

## San Solomon Ciénega

A cooperative project in West Texas has recreated a unique and valuable type of aquatic habitat that is rapidly vanishing from the desert Southwest. The main purpose of the restoration project was to create vital habitat, not only for the two endangered fishes, Comanche Springs pupfish and Pecos gambusia, but also for all of the plants and animals that lived in these fragile desert wetlands. Additional benefits include educational opportunities, boost to a local economy, and protection of an agricultural lifestyle.

Few ciénegas, or desert wetlands, have survived intact to this day. San Solomon Springs continues to flow at Balmorhea State Park, but its associated ciénega was destroyed long ago, along with its great diversity of wildlife. San Solomon Springs is currently the largest spring in the Trans-Pecos and the sixth largest in the state. Comanche Springs, in nearby Fort Stockton, used to be able to claim that title, but groundwater pumping caused its perennial flow to cease entirely in 1961.

The native inhabitants of the San Solomon ciénega ecosystem have suffered greatly. When the original wetlands were modified and for the most part destroyed, the only aquatic habitat remaining was in concrete irrigation canals. Although better than no habitat at all, the irrigation canals, at best, provided a tenuous existence for many of the aquatic species.

People also suffer when their water sources vanish. Farmers who depended on surface irrigation from Comanche Springs lost everything when the springs went dry. Farmers in the Balmorhea area also rely on surface irrigation from springs, and if serious conflicts on groundwater use were to occur, local agriculture would suffer. The effects on the community of Balmorhea also would be catastrophic because the community depends on the aquifer and the spring flows for everything from drinking water to tourist dollars.

Somewhat ironically, the one thing that can prevent over-pumping of this aquifer is the Federal Endangered Species Act. The Endangered Species Act protects the fish, the fish need the water, and as long as the water is flowing from the springs, it is also available to humans downstream. Through a pragmatic understanding of the basic relationship between the natural and human communities, biologists and Balmorhea community leaders chose to come together to work out a solution that would benefit all concerned, rather than adopt adversarial roles, which so often occurs today.

A plan was formulated to create a ciénega that would look and function like a natural ecosystem. In this way, the survival of the fishes and a dependable water supply could be assured. Water, of course, was the most important element of the whole plan. The Reeves County Water Improvement District and the agricultural community it represents agreed to provide the essential water needed to create a secure environment for the endangered species. Water is a rare and precious commodity in far West Texas, particularly for farmers, but by each of the users giving up a small amount, they could ensure that they all would have water for the future.

An additional benefit for the farmers was that, because of their help in creating a permanent habitat for the endangered fishes, the Texas Department of Agriculture, the U.S. Fish and Wildlife Service, and the U.S. Environmental Protection Agency worked out a plan to reduce some the extra pesticide restrictions that had been in place to protect the endangered species in the irrigation canals. The fish have a better place to live, and the farmers can continue to raise their crops economically.

Biologists, engineers, and resource managers from universities and government agencies joined forces to make the project work. The USDA Natural Resource Conservation Service provided soil analysis and, along with the Texas Agricultural Extension Service and the Texas Department of Agriculture, gave expert advice on some of the intricacies of the project. The expertise of the Texas Department of Transportation also was crucial. Their surveyors, design engineers, and equipment operators transformed biological ideas into reality. The Texas Department of Criminal Justice provided inmate manpower to do such things as build the observation deck and retaining walls, as well as install the plant materials selected for the initial ciénega vegetation restoration. Botanists at Sul Ross State University provided container-grown native plants for the project.

## Devils River Minnow Conservation Agreement

*Dionda diaboli* was recommended for listing as endangered by the U.S. Fish and Wildlife Service because of its extremely low numbers and reduced habitat. It is somewhat of a mystery why this fish almost disappeared after being one of the more abundant species present. Certainly the mere act of putting it on the Endangered Species list would not do much for the species. Something needed to be done to determine the causal factors and protect not only the fish, but also the health of the rivers and creeks in which it occurred. Private landowners and the city of Del Rio were extremely interested in working with the Texas Parks and Wildlife Department to determine and resolve these problems. As a result, the species was listed as threatened and a Conservation Agreement was developed.

In formulating the Conservation Agreement, the Texas Parks and Wildlife Department and the U.S. Fish and Wildlife Service agreed to work closely with landowners and the City of Del Rio to determine and resolve life-history requirements and restore populations to natural levels. As a result, there will also be additional protection for the quality of the Devils River and associated streams.

Specifically, the conservation actions outlined in the agreement are designed to (1) assess the current status of wild populations, (2) provide immediate security for the Devils River minnow, (3) implement actions needed for long-term conservation of the Devils River minnow, and (4) fill in gaps in pertinent information.

## Pecos Pupfish Conservation Agreement

*Cyprinodon pecosensis* was recommended for listing by the U.S. Fish and Wildlife Service because of its loss of habitat and massive hybridization with *C. variegatus*. In this situation, the rationale of the Texas Parks and Wildlife Department was to fix the problem and preclude the need to list. If the State were to fail, the species would be listed.

The approach has three components: (1) amend baitfish regulations to prevent further introductions of nuisance fishes; (2) protect the existing natural population, and (3) create new habitat through a landowner-incentive program that turns stock ponds into ciénegas, thus creating alternate habitat on private land. To date, baitfish regulations have been changed, progress in habitat protection has been made, and, perhaps most importantly, two wetlands on private land have been created that have thriving populations of Pecos pupfish.

## Summary

Exploitation of limited resources, particularly groundwater pumping, has degraded the West Texas environment, caused extirpation and extinction of species, and, ultimately, loss of habitat and ecosystems. Many of the fishes of this region could serve well as biological indicators of the overall integrity of the ecosystem. The few remaining relatively intact faunas and unmodified localities need careful management if they are to be preserved. In addition, information gained by studying aquatic communities can be used to provide useful baseline data for future actions and decisions affecting the management of the Chihuahuan Desert ecosystem within the larger bi-national border region. By involving individuals and local governments, we are more likely to achieve long-term benefits for natural resources, as well as public health and quality of life.

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# Chapter 6

# The Hueco Bolson: An Aquifer at the Crossroads

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## Introduction

The Hueco Bolson is a thick pocket of sediments derived from nearby mountains that extends from New Mexico, through Texas, and into Mexico in the El Paso and Ciudad Juarez area. Over time, these sediments filled with water and became the Hueco Bolson aquifer: an oasis of plentiful water in the northern part of the Chihuahuan Desert. El Paso and Ciudad Juarez have relied on the Hueco Bolson aquifer as a primary source of drinking water for several decades (Sayre and Penn, 1945; White and others, 1997). Ciudad Juarez, several communities in New Mexico, and the Fort Bliss Military Reservation currently depend on the Hueco Bolson aquifer as their sole source of drinking water (Sheng and others, 2001). Because of the desert climate and the local geology, the aquifer is not easily replenished, and recharge is low. Low recharge and high pumping rates have caused large water-level declines and large decreases in fresh-water volumes in the aquifer.

The aquifer and the El Paso-Ciudad Juarez area are at the crossroads. With current trends, groundwater models predict that El Paso will pump the last of its fresh water by 2025, and Ciudad Juarez will pump the last of its fresh water by 2005 (Sheng and others, 2001). The El Paso Water Utilities/Public Service Board (EPWU) has recognized the nature of limited groundwater resources in the area and has investigated and invested in several strategies to increase the longevity and usefulness of the aquifer. The purpose of this paper is to briefly summarize the hydrogeology of the Hueco Bolson aquifer and discuss several of the management strategies to protect and responsibly use the aquifer.

## Hydrogeology

The Hueco Bolson aquifer is coincident with the Hueco Bolson, a long, sediment-filled trough that lies between the Franklin, Organ, and San Andres Mountain ranges and the Quitman, Malone, Finlay, Hueco, and Sacramento Mountain ranges (fig. 6-1). Hill (1900) defined the Hueco Bolson as including the Tularosa Basin (as shown in fig. 6-1).

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[2] Texas Water Development Board
[3] El Paso Water Utilities

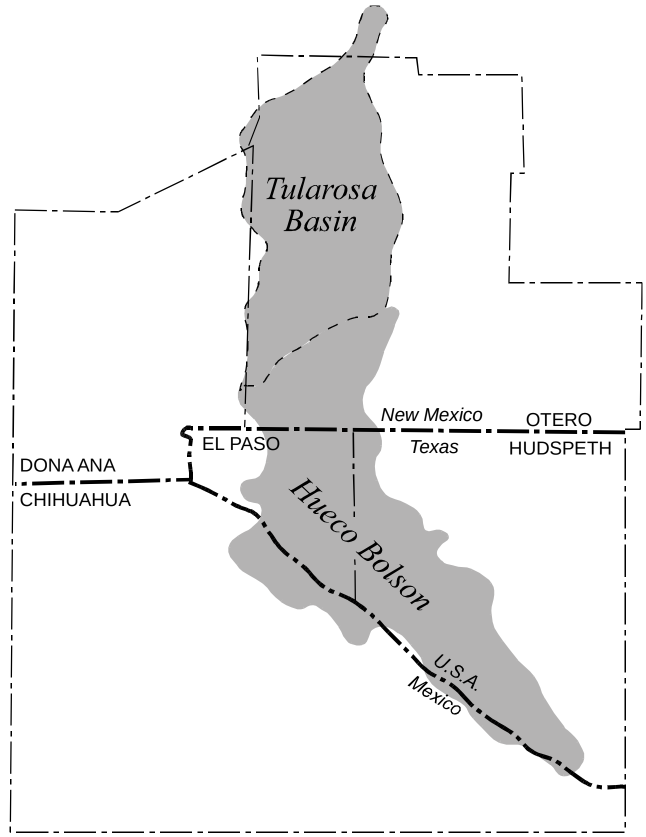


Figure 6-1: Location of the Hueco Bolson aquifer in Texas, New Mexico, and Mexico.

However, Richardson (1909) divided the bolson into two parts: the Tularosa Basin to the north and the Hueco Bolson to the south. The topographic divide between these two basins is about 7 mi north of the Texas–New Mexico border. However, the Hueco Bolson and the Tularosa Basin are hydraulically connected to each other (Wilkins, 1986) and have been combined into the Hueco-Tularosa aquifer (Hibbs and others, 1997).

The Hueco Bolson is about 200 mi long and 25 mi wide. The Hueco Bolson aquifer consists of unconsolidated to slightly consolidated deposits composed of fine- to medium-grained sand with interbedded lenses of clay, silt, gravel, and caliche. Sediments in the bolson are fluvial, evaporitic, alluvial fan, and aeolian in origin and have a maximum thickness of 9,000 ft (Mattick, 1967; Cliett, 1969; Abeyta and Thomas, 1996). The bottom part of the Hueco Bolson is primarily clay and silt. Therefore, only the top several hundred feet produce good-quality water.

## Recharge

The Hueco Bolson aquifer is recharged by mountain-front recharge; seepage from the Rio Grande, canals, and agricultural drains; and deep-well injection (Knorr and Cliett, 1985;

Land and Armstrong, 1985; White and others, 1997). Mountain-front recharge is the seepage of surface run-off after rainfalls into the aquifer where the bolson laps up against bordering mountains. Before the aquifer was heavily pumped, water in the aquifer naturally discharged to the Rio Grande. After pumping caused water levels to decline, the Rio Grande began to lose water into the aquifer, so much so that a part of the river through El Paso-Ciudad Juarez has been lined with concrete to minimize leakage. Unlined irrigation canals and drains also leak water into the aquifer, although the water is usually of poor quality. EPWU has taken treated wastewater and injected it up-gradient of one of El Paso's well fields to increase recharge to the aquifer.

Meyer (1976) estimated that mountain-front recharge (from the Organ and Franklin Mountains in New Mexico and Texas and the Sierra de Juarez in Mexico) to the aquifer in El Paso County is 5,640 acre-ft/yr. White (1987) estimated that about 33,000 acre-ft/yr of water is recharged into the Rio Grande alluvium overlying the bolson aquifer. Recharge from the Rio Grande was reduced significantly when the bottom of the Rio Grande was lined in 1973 and 1998 in the El Paso-Ciudad Juarez area (Hibbs and others, 1997; Heywood and Yager, in review).

Treated wastewater is injected at the Fred Harvey Wastewater Treatment Plant in El Paso and provided about 3,800 acre-ft of water per year in 1995 (USEPA, 1995) and about 1,800 acre-ft in 1999 (Sheng and others, 2001).

## Well Yields

Well yields in the Texas part of the Hueco Bolson aquifer are as much as 1,800 gpm (Hibbs and others, 1997). In New Mexico, yields are higher in alluvial fans that flank the basin (~1,400 gpm) and lower in the interior of the basin (300 to 700 gpm) (Hibbs and others, 1997). In the well field for Ciudad Juarez in Mexico, yields range between 300 and 1,500 gpm (Hibbs and others, 1997). Hydraulic conductivity in the Hueco Bolson, as determined with 73 aquifer tests, varies from 6.4 to 98.9 ft/day (Hibbs and others, 1997).

## Pumping

The Hueco Bolson aquifer is pumped at a much greater rate than the aquifer is recharged. Groundwater withdrawals from the aquifer in Texas amounted to about 69,000 acre-ft in 1999 (Sheng and others, 2001): about nine times greater than the amount of recharge in El Paso County. Over the past 20 yr, pumping from the Hueco and Mesilla Bolsons in Texas has ranged from 96,000 to 138,000 acre-ft/yr (Mace, this volume).

## Water Quality

Water quality in the Hueco Bolson varies depending on location and depth. Water quality in the Texas part of the Hueco Bolson tends to be better to the west than to the east, although there are pockets of good-quality water in the eastern part of the bolson (Gates and others, 1980). North of the Texas-New Mexico border, water tends to have total dissolved solids (TDS) greater than 1,000 mg/L except near mountain fronts where there is active recharge (Hibbs and others, 1997). The upper part of the aquifer tends to be fresher with TDS ranging between 500 and 1,500 mg/L, with an average of about 640 mg/L (Ashworth and Hopkins, 1995). Water quality has been affected by the large water-level declines in the aquifer, which have induced flow of poor-quality water into areas of fresh water. Water quality in the shallow part of the aquifer along the Rio Grande in the alluvium has degraded because of leakage of poor-quality irrigation return-flow into the aquifer (Sheng and others, 2001). Water quality beneath Ciudad Juarez is generally less than 1,000 mg/L TDS (Hibbs and others, 1997), however, water-quality deterioration has been observed in wells along the border and in the downtown area.

## Water Levels and Groundwater Flow

Depth to water in the Hueco Bolson aquifer ranges from very shallow to very deep. Depth to groundwater near the Cities of Tularosa and Alamogordo is between 20 and 150 ft, whereas depth to water below El Paso ranges from 250 to 400 ft in depth, and depth to water below Ciudad Juarez ranges between 100 and 250 ft (Hibbs and others, 1997). Depth to water below the Rio Grande is less than 70 ft. Groundwater flows from the Tularosa Basin southward into the Hueco Bolson and into Texas (Hibbs and others, 1997, their fig. 3.8). Little drawdown has been recorded in the northern part of the aquifer. The drawdown in Hueco Bolson along the Texas-New Mexico border has been relatively small, not exceeding 30 ft (Hibbs and other 1997). In heavily developed parts of the Hueco Bolson aquifer, drawdowns since predevelopment in 1903 are up to 170 ft. Focal points of drawdown are beneath the City of El Paso and Ciudad Juarez (Hibbs and others, 1997).

The model by Heywood and Yager (in review) suggests that about 6,000 acre-ft/yr of groundwater flowed in the Hueco Bolson aquifer from New Mexico into Texas before large-scale pumping by El Paso in the 1960’s. Since then, the amount of flow has increased to about 18,000 acre-ft/yr. In the El Paso-Ciudad Juarez area, groundwater flows toward cones of depression. Between 1910 and 1960, groundwater flowed from Mexico into Texas toward pumping centers in El Paso (Sheng and others, 2001). Since 1960, groundwater, generally of poor quality, has flowed from Texas into Mexico (Sheng and others, 2001).

## Groundwater Models

Several groundwater flow models have been constructed for the Hueco Bolson aquifer system. These models include an early electric-analog model of the El Paso area (Leggat and Davis, 1966) and three numerical models developed by the U.S. Geological Survey, including (1) Meyer and Gordon (1973) and Meyer (1975, 1976) (later updated by Knowles and Alvarez, 1979), (2) Groschen (1994), and (3) an as yet unpublished model (Heywood and Yager, in review). Mullican and Senger (1990, 1992) developed a model of the southeastern part of the Hueco Bolson. Mexico has also developed a groundwater flow model for part of the area. Wilson and others (1986) used a preexisting model to predict water resources through 2060.

Models by Groschen (1994) and Heywood and Yager (in review) simulate potential water-level declines, as well as changes in water quality due to pumping. Groschen (1994) showed that water quality in the bolson is most likely affected by horizontal movement of saline water in response to pumping.

The integrated flow and water-quality model by Heywood and Yager (in review) represents the cooperation of EPWU, the USGS, the International Boundary and Water Commission (IBWC), Fort Bliss Military Reservation, JMAS (Junta Municipal de Agua y Saneamiento de Ciudad Juarez), and CILA (Comision Internacional de Limites y Aguas). Binational coordination has included the exchange of aquifer information and comparison of water-resource management plans. The model is being used to assess (1) water storage in the aquifer, (2) the optimization of pumping for fresh and brackish water, (3) the location of new production wells, (4) the control of brackish-water intrusion, (5) the design of an aquifer storage and recovery program, and (6) the planning of water resources among Texas, New Mexico, and Mexico (Sheng and others, 2001).

# Groundwater Availability

Groundwater availability represents the amount of water that can be used from an aquifer. Groundwater availability can be defined in many different ways depending on the local socioeconomic needs (Mace and others, 2001). In the El Paso area, groundwater availability has been defined using a systematic depletion approach, where the total amount of recoverable water is considered the amount of water available for use. In general, groundwater availability is assessed for the fresh-water part of the aquifer. However, as water resources become scarcer in the state, more and more areas, including El Paso, are also evaluating the usable amounts of slightly saline water for ongoing or potential desalination projects.

## Fresh Water

The approximate volume of recoverable freshwater in the entire Hueco Bolson aquifer is about 7.5 million acre-ft, with 3 million acre-ft in Texas, 3.9 million acre-ft in New Mexico, and 600,000 acre-ft in Mexico (Sheng and others, 2001, on the basis of a review of USGS publications). The Far West Texas Planning Group estimated that there were about 3 million acre-ft of fresh water in the Hueco Bolson and 2.5 million acre-ft of slightly saline water for desalination (FWTPG, 2001). Recoverable fresh water accounts for economic and geologic constraints and does not represent all of the fresh water in the aquifer.

Other studies have suggested differing volumes of fresh water. Knowles and Kennedy (1956) estimated that the Hueco Bolson in Texas had about 7.4 million acre-ft of recoverable water, with less than 250 mg/L chloride (~750 mg/L TDS). Meyer (1976) estimated the recoverable amount of fresh water in the Texas part of the Hueco Bolson to hold 10.64 million acre-ft. White (1987) estimated that the Hueco Bolson aquifer in Texas holds about 9.95 million acre-ft of recoverable fresh water. The TWDB (1997) estimated that there was about 9 million acre-feet of fresh water in the Texas part of the Hueco Bolson.

### Slightly Saline Water

Slightly saline water may be a large potential water resource in the El Paso area. There is an estimated 20 million acre-ft of slightly saline water (TDS between 1,000 and 3,000 mg/L) in the Hueco Bolson aquifer in El Paso County (Sheng and others, 2001). Similar volumes of slightly saline water may also exist in New Mexico and Mexico (Sheng and others, 2001). Sheng and others (2001) recommended additional studies to quantify a more exact volume of poor-quality water in the aquifer.

## Strategies to Increase Groundwater Availability

Although recent modeling work suggests that the Hueco Bolson in the El Paso area will run out of fresh water by 2025, it is not a forgone conclusion. For prediction purposes, the model assumes that current trends and practices will remain the same. However, the life of the fresh groundwater resource can be extended by implementing strategies to increase the availability of groundwater.

### Increase Surface-Water Use

By increasing the use of surface water, groundwater use can be minimized, thus extending the useful life of the fresh-water part of the aquifer. In this case, surface water is relied upon when plentiful, and groundwater is relied upon when surface water is not plentiful. Regional water providers are pursuing this strategy by the implementation of the Regional Sustainable Water Project (IBWC and EPWU, 2000). The Far West Texas Planning Group also identifies the pursuit of additional surface-water supplies as a recommended water management strategy for the area (FWTPG, 2001). However, the planning group noted that El Paso cannot rely on the Rio Grande for water during times of severe drought (FWTPG, 2001).

### Hydraulic Control and Desalination

To reduce the degradation of groundwater quality due to laterally flowing poorer quality water, wells can be installed to hydraulically control the migration of poorer quality water by capturing the poorer quality water before it mixes with fresher water. The produced water can then be desalinated. EPWU and the Department of Defense at the Fort Bliss Military Reservation are investigating this approach in existing wells in the Airport/Montana well field (Sheng and others, 2001). To maximize the water supply, the desalinated water (~200 mg/L TDS) can then be blended with slightly saline water (~1,500 mg/L TDS) to produce a water with a TDS of about 900 mg/L TDS. Hydraulic control and desalination extend the life of the fresh-water part of the aquifer by protecting existing fresh-water resources from further intrusions of poor-quality water and decreasing the reliance on the fresh-water part of the aquifer. Hydraulic control and desalination are also being considered in other El Paso wellfields (Sheng and others, 2001).

### Pumping Optimization

Pumping of wells can be optimized to minimize the migration of poor-quality water and the depth of cones of depression around pumping centers. Pumping of water-supply wells should be optimized aquiferwide to minimize the effects of pumping on the migration of poor-quality water into areas of fresh water. An operational priority list for the Hueco well fields has been developed and used in well-field operation for over a year (Sheng and others, 2001). Results of the optimization program will be evaluated to further improve operation of the well fields. Pumping optimization extends the life of the fresh-water resource by minimizing the impacts of poor-quality water intrusions.

### Aquifer Storage and Recovery

Aquifer Storage and Recovery (ASR) is when treated surface water is injected into an aquifer when it is plentiful and demand is low, and then recovered the stored water from the aquifer when demand is high or during times of drought. ASR extends the life of the aquifer by maximizing the use of surface water and recharging the aquifer. In addition, it will also prevent brackish water intrusion if injection wells are located along the transition zone of marginal quality groundwater.

### Blending high-grade water with poor-quality water

Using the best quality water first has often been the preferred method of groundwater production. However, by blending good quality water with poorer quality water up to the Safe Drinking Water Act standards for TDS, chloride, and sulfate secondary maximum contamination levels, water providers can enhance their production capacity. The blending method extends the life of the aquifer by maximizing the use of the freshwater resource. When combined with hydraulic control, existing freshwater resources can also be additionally protected.

## Conclusions

The Hueco Bolson aquifer and the El Paso-Ciudad Juarez area are at the crossroads. Several scientific studies and recent modeling projects suggest that, under current trends, fresh water from the Hueco Bolson aquifer in Texas will be depleted by 2025. However, using groundwater more strategically can extend the longevity of fresh-water resources in the aquifer. EPWU and FWTPG are actively researching and implementing a number of strategies to do just this, including increased surface-water use, hydraulic control and desalination, pumping optimization, aquifer storage and recovery, and blending to increase freshwater supplies. The area will need to continue to follow this path to ensure that future water needs of El Paso are met.

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# Chapter 7

# The Mesilla Basin Aquifer System of New Mexico, West Texas, and Chihuahua—An Overview of Its Hydrogeologic Framework and Related Aspects of Groundwater Flow and Chemistry

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## Introduction

Our brief overview of the hydrogeology and geohydrology of basin-fill aquifers in the Mesilla Basin (Bolson) covers a large area of south-central New Mexico and adjacent parts of western Texas and northern Chihuahua (figs. 7-1, 7-2). Emphasis is on the hydrogeologic framework of this major intermontane basin and the controls exerted by basin-fill stratigraphy and structure on the distribution of major aquifer zones, the groundwater-flow regime, and related aspects of water chemistry. The 1,100-mi$^2$ Mesilla Basin is near the southern end of the river-linked series of structural basins that form the Rio Grande rift (RGR) tectonic province (Keller and Cather, 1994). The RGR extends southward from the San Luis Basin, which is flanked by the southern Rocky Mountains, to at least as far south as the Hueco Bolson in the southeastern sector of the Basin and Range province (Hawley, 1978; 1986).

The broad structural depression that forms the Mesilla Basin is bounded on the east by the Organ-Franklin-Juarez Mountain chain, and its western border includes fault-block and volcanic uplands that extend northward from the East Potrillo Mountains and West Potrillo basalt field to the Aden and Sleeping Lady Hills (figs. 7-1, 7-2). The entrenched Mesilla Valley of the Rio Grande, which has a valley-floor area of about 215 mi$^2$, crosses the eastern part of the basin. The metropolitan areas of Las Cruces and northwestern El Paso-Ciudad Juarez are located, respectively, in the northern part and at the southern end of the Mesilla Valley. The Robledo and Doña Ana mountains bound the northern end of the valley, but the northeastern basin boundary is transitional with the Jornada del Muerto Basin (Seager and others, 1987). The southern basin-boundary with the Bolson de los Muertos in north-central Chihuahua has still not been studied in detail. Regional groundwater and surface flow is toward “El Paso del Norte,” the topographic and structural gap between the Franklin Mountains and Sierra Juarez that separates the

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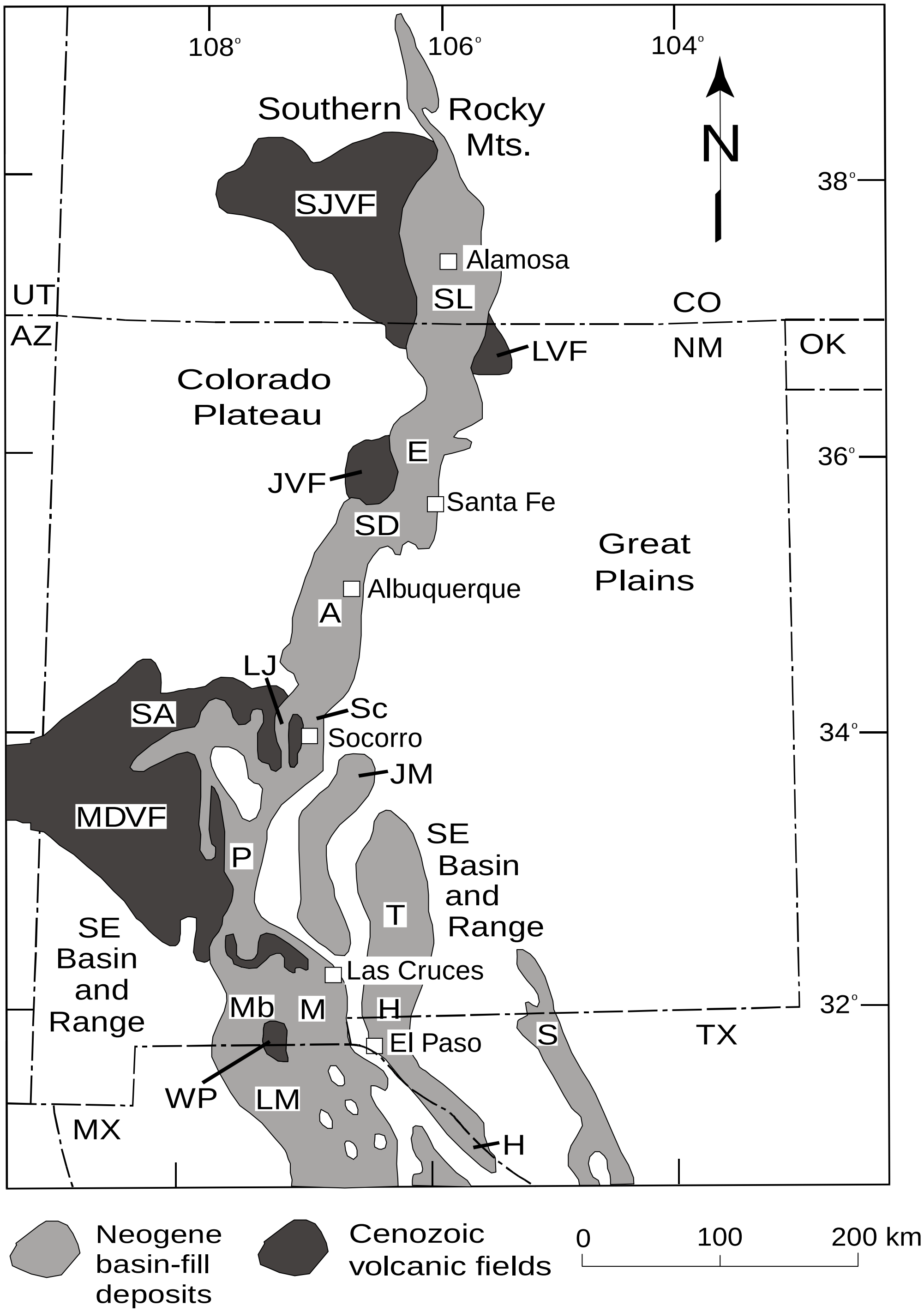


Figure 7-1: Index map showing location of the Mesilla Basin in the context of other basins and volcanic fields within the Rio Grande rift structural province. Basin abbreviations from north to south: San Luis (SL), Española (E), Santo Domingo (SD), Albuquerque (A), Socorro (Sc), La Jencia (LJ), San Agustin (SA), Jornada del Muerto (JM), Palomas (P), Tularosa (T), Mimbres (Mb), Mesilla (M), Los Muertos (LM), Hueco (H), and Salt (S). Cenozoic volcanic fields: San Juan (SJVF), Latir (LVF), Jemez (JVF), Mogollon-Datil (MDVF), and West Potrillo (WP) (modified from Keller and Cather, 1994).

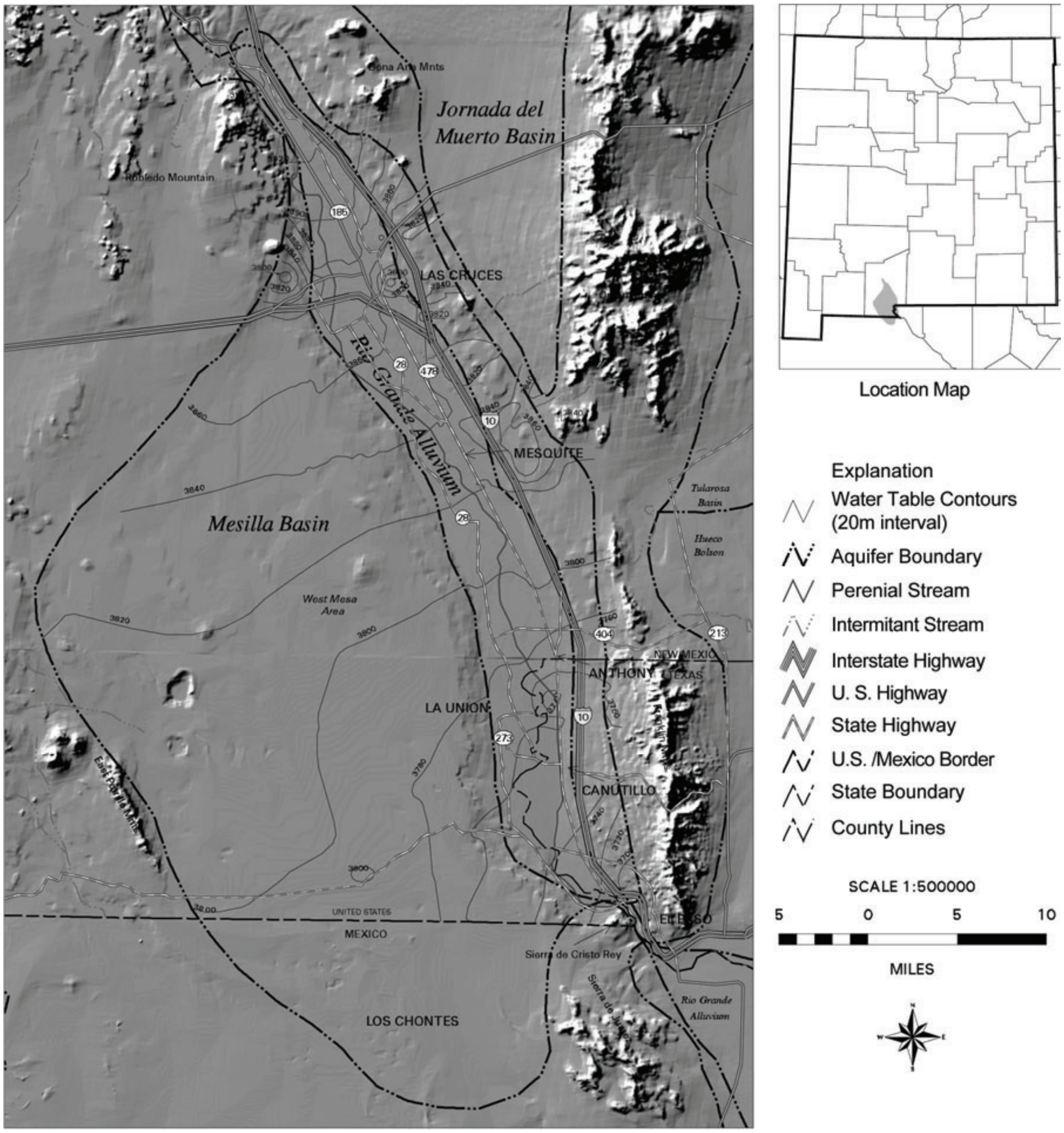


Figure 7-2: Shaded-relief index map of the Mesilla Basin area of southern New Mexico and adjacent parts of Texas and Chihuahua showing extent of modeled basin-fill (Santa Fe Gp) and Mesilla Valley aquifer systems. General water-table configuration and groundwater-flow direction in upper aquifer units are also illustrated (adapted from Hibbs and others [1997], with shaded relief from latest available U.S. Geological Survey DEM database).

Mesilla Basin from the Hueco Bolson (fig. 7-2). Underflow contribution from the Jornada Basin is restricted by a buried bedrock high between the Doña Ana and Tortugas Mountains east of Las Cruces (King and others, 1971; Wilson and others, 1981; Woodward and Myers, 1997).

## Background—Development of Basic Hydrogeologic and Geohydrologic Models

We open this discussion with a brief review of how the present conceptual model of basin hydrogeology has developed over the past century. In terms of modern concepts of groundwater flow in basin-fill aquifers, W. T. Lee (1907) and Kirk Bryan (1938) are the two most important early workers to characterize the Rio Grande Valley and RGR region between Colorado and Trans-Pecos Texas. However, we must note here that the contributions of Lee and Bryan are just one product of the great amount of cross-fertilization of geological and hydrological concepts that occurred throughout the American Southwest during the late 19$^{th}$ and early 20$^{th}$ centuries. Very important contributions by contemporary workers include Hill (1896, 1900), Slichter (1905), Richardson (1909), Tolman (1909, 1937), Meinzer and Hare (1915), Darton (1916), Schwennesen (1918), Dunham (1935), Theis (1938), and Sayre and Livingston (1945).

Lee (1907) developed an early scenario for evolution of the Rio Grande Valley system in New Mexico and emphasized the potential for building a large dam at the Elephant Butte site for irrigation water storage. Bryan's (1938) most significant hydrogeologic contributions include development of the earliest synoptic models of the RGR structural province (his "Rio Grande depression") and evolution of the northern Rio Grande fluvial system. He observed that (1) the main body of sedimentary deposits of the Rio Grande depression, from the north end of the San Luis valley to and beyond El Paso, is considered to be the same general age and to belong to the Santa Fe Formation (p. 205); (2) in general, the basins appear to have been elongated into ovals and to be divisible into two major types— basins with a through-flowing river and basins with enclosed drainage (p. 205); and, (3) [Rio Grande depression basins] differ from other basins [in the Basin and Range province] principally in being strung like beads on a string along the line of the Rio Grande (p. 221).

On the basis of observations in adjacent parts of Mexico and the American Southwest, Tolman (1909, 1937) also made a major contribution in developing a better definition of the fundamental hydrogeologic distinction between depositional systems in aggrading intermontane basins with topographic closure (bolsons) and those that are open in terms of both surface and subsurface flow (semibolsons). Both Bryan and Tolman recognized three basic classes of lithofacies assemblages in this continuum of closed and open basin landforms. Piedmont-slope facies (e.g., alluvial-fan) are present along the margins of both basin types, while basin floors in closed systems include alluvial flats that grade to terminal, playa-lake plains. Floors of basins that are integrated with surface-flow systems, in marked contrast, include alluvial flats and fluvial plains that grade to basin outlets.

Figures 7-3 and 7-4 illustrate the Bryan-Tolman conceptual models of the hydrogeologic framework and groundwater-flow regimes in basin-fill aquifer systems of the Basin and Range province. Figure 7-3, adapted from Bryan (1938, his figs. 51, 52), clearly demonstrates his basic understanding of the integrated groundwater and surface-water flow system in basins of the "Rio Grande depression." Figure 7-4, adapted from Eakin and others (1976), illustrates the Bryan-Tolman conceptual model in a more general hydrogeologic sense for the entire Basin and Range province, and it incorporates subsequent work in the Great Basin section (e.g., Mifflin, 1988), as well as the Trans-Pecos Texas and Chihuahua bolson region (Hibbs and others, 1998). The topographic terms *closed* and *open* are here used only in reference to the surface flow into, through, and from intermontane basins, whereas the terms *undrained*, *partly drained*, and *drained* designate classes of groundwater flow involving intrabasin and/or interbasin movement. *Phreatic* and *vadose*, respectively, indicate saturated and unsaturated subsurface conditions. *Phreatic playas* (with springs and seeps) are restricted to floors of *closed* basins (*bolsons, bolsones*) that are *undrained* or *partly drained;* and *vadose playas* occur in both *closed* and *open*, *drained* basins. In the Mesilla Basin region, as well as in most other intermontane basins of western North America, the intermediate basin class referred to as *partly drained* probably represents the major groundwater-flow regime. Few intermontane basins (*bolsons* and *semibolsons*) are truly *undrained* in terms of groundwater discharge, whether they are *closed* or *open* in terms of surface flow.

Under predevelopment conditions, groundwater discharge in the region occurred mainly through subsurface leakage from one basin system into another, discharge into the gaining reaches of perennial or intermittent streams, discharge from springs, or by evapotranspiration from *phreatic playas* and cienegas (valley-floor wetlands). Most recharge to basin-fill aquifers occurs by two mechanisms, (1) "mountain front," where some precipitation falling on bedrock highlands contributes to the groundwater reservoir along basin margins, and (2) "tributary," where the reservoir is replenished along losing reaches of larger intrabasin streams (Hearne and Dewey, 1988; Kernodle, 1992; Wasiolek 1995; Anderholm, 2000). Recharge estimates in this paper are based on the assumption that (1) less than 2 percent of average annual precipitation contributes to recharge and (2) this contribution is distributed very unevenly over higher watersheds and in major stream valleys.

## Developments Since 1945

Scientific and technological breakthroughs since 1945 include development of modern geophysical-survey and deep-drilling methods and advances in geochemistry. These breakthroughs contributed to much better characterization of basin-fill aquifers and groundwater-flow systems in the southern New Mexico–Trans-Pecos Texas region by the U.S. Geological Survey, Texas Water Commission, City of El Paso, U.S. Soil Conservation Service, and New Mexico State University (e.g., Conover, 1954; Knowles and Kennedy 1958; Leggat and others, 1962; Cliett 1969; Hawley, 1969; Hawley and

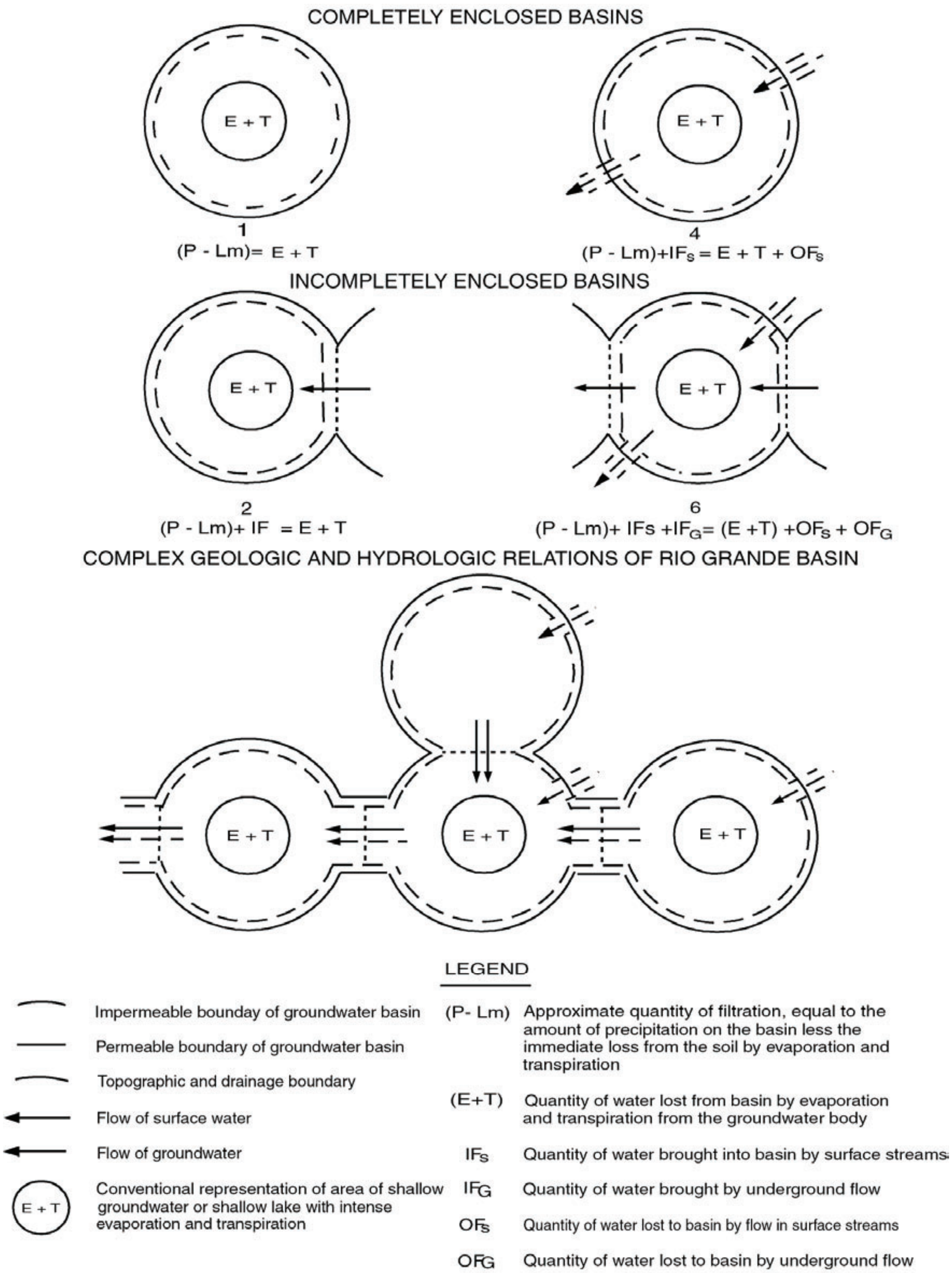


Figure 7-3: Kirk Bryan's conceptual models of "hydraulic regimes" in groundwater reservoirs of the "Rio Grande depression" (modified from Bryan, 1938, his figs. 51 and 52).

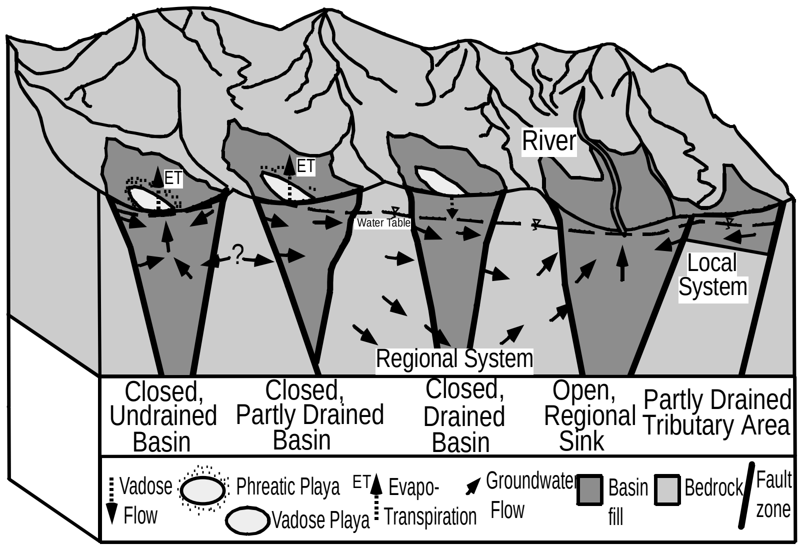


Figure 7-4: Schematic diagram showing hydrogeologic framework and groundwater-flow system in interconnected group of *closed* and *open, undrained, partly drained, and drained* intermontane basins (modified from Eakin and others, 1976, and Hibbs and others, 1998).

others, 1969; King and others, 1971; Gile and others, 1981; Wilson and others, 1981; Wilson and White, 1984).

Recent investigations are characterized by the increased availability of high-quality geophysical and geochemical data and deep borehole sample and core logs. We are now in an era dominated by the exponentially increasing power of computers and the evolution of numerical modeling and GIS technology. In the Mesilla Basin region, as elsewhere, the bridge between the early-20$^{th}$-century conceptual world and the present will continue to be *hydrogeologic ground truth*. Our surface and underground view of geohydrologic systems must now be expressed in units that modelers of groundwater-flow systems can understand and computers can process. The rapid improvements in our understanding of subsurface geophysical and geochemical systems, geochronology, and the definition of the hydrogeologic units described herein now allow modelers to join forces effectively with hydrogeologists, geophysicists, and geochemists in meeting the incredible water-resource challenges that face Third Millennium society in this and other arid and semiarid regions.

Recent and ongoing studies in the Mesilla Basin area that have provided much of the background material for our paper are reviewed or described in detail in the following reports and maps: Peterson and others (1984), Seager and others (1987), Hawley and Lozinsky (1992), Wade and Reiter (1994), Heywood (1995), Seager (1995), Gile and others (1996), Hibbs and others (1997; 1998; 1999), Hibbs (1999), Collins and Raney (2000), Hawley and others (2000), and Kennedy and others (2000). Discussion of a large geothermal system located near Tortugas Mountain, east of Las Cruces, is beyond the scope of this paper. The system results from very deep circulation of meteoric-source groundwater in a high heat-flow environment. Research on this complex (regional and local) groundwater-flow regime is in progress (e.g., Ross and Witcher, 1998).

### Recent Developments in Groundwater-Flow Models

Models of groundwater flow in the Mesilla Basin aquifer system (e.g., Peterson and others, 1984; Frenzel and Kaehler, 1990; West, 1996; Balleau, 1999; ) must be examined in terms of the hydrogeologic constraints placed on flow regimes by lithofacies, stratigraphic, and structural-boundary conditions that are either well documented or reasonably inferred. The critique of "U.S. Geological Survey Ground-Water-Flow Models of Basin-Fill Aquifers in the Southwestern Alluvial Basins Region" (Kernodle, 1992) relates directly to this concern. "As a rule identifiable geologic features that affect groundwater-flow paths, including geologic structure and lithology of beds, need to be represented in the model," (p. 65) and major categories of geohydrologic boundaries in alluvial basins include "(1) internal boundaries that alter flow paths, including small-permeability beds, fissure-flow volcanics and faults; (2) recharge boundaries, primarily around the perimeter of basins (mountain-front recharge), and along the channels of intermittent streams, arroyos, and washes (tributary recharge); [and] (3) recharge and discharge boundaries associated with semipermanent surface-water systems in the flood plains of major streams." (p. 66) Finally, "although two-dimensional models may successfully reproduce selected responses of the aquifer, they fail to accurately mimic the function of the system." (p. 59) In comparison, "three-dimensional models more accurately portray the flow system in basin-fill [aquifers] by simulating the vertical component of flow. However, the worth of the model is still a function of the accuracy of the hydrologist's concept of the workings of the aquifer system." (p. 59)

We must also emphasize that short- and long-term climatic changes have significant impacts on all water resources. This observation is well documented by both modern meteorological data and the historic and prehistoric tree-ring record (Thomas, 1962; Schmidt, 1986; D'Arrigo and Jacoby, 1992; U.S. Dept. of Commerce, 1999; and Hawley and others, 2000). For example, the region experienced prolonged droughts from the late 1940's until the late 1970's, and the following two decades were abnormally wet.

## Conceptual Hydrogeologic-Framework Model

The hydrogeologic framework of basin-fill aquifers in the RGR region, with special emphasis on features related to environmental concerns, is described here in terms of three basic conceptual building blocks: lithofacies assemblages (LFA's), hydrostratigraphic units (HSU's), and structural-boundary conditions. A conceptual hydrogeologic model of an interconnected, shallow, valley-fill/basin-fill and deep-basin aquifer system was initially developed for use in groundwater-flow models of the Mesilla and Albuquerque basins (Peterson and others, 1984; Frenzel and Kaehler, 1990; Hawley and Haase, 1992; Hawley and Lozinsky, 1992; Kernodle, 1992, 1998; Thorn and others, 1993; Hawley and others, 1995; Kernodle and others, 1995). However, basic design of the conceptual model is flexible enough to allow it to be modified for use in other basins of the Rio Grande rift and adjacent parts of the southeastern Basin and Range province (Hawley and others, 2000).

Hydrogeologic models of this type are simply qualitative to semiquantitative descriptions (graphical, numerical, and verbal) of how a given geohydrologic system is influenced by (1) bedrock-boundary conditions, (2) internal-basin structure, and (3) lithofacies characteristics of various basin-fill stratigraphic units. They provide a mechanism for systematically organizing a large amount of relevant hydrogeologic information of widely varying quality and scale (from very general drillers' observations to detailed borehole logs and water-quality data). Model elements can then be graphically displayed in combined map and cross-section (GIS) formats so that basic information and inferences on geohydrologic attributes (e.g., hydraulic conductivity, transmissivity, anisotropy, and general patterns of unit distribution) may be transferred to basin-scale, three-dimensional numerical models of groundwater-flow systems. As emphasized by Hawley and Kernodle (2000), however, this scheme of data presentation and interpretation is normally not designed for site-specific groundwater investigations.

## Lithofacies Assemblages

Lithofacies assemblages (LFA's) are the basic building blocks of the hydrogeologic model (fig. 7-5, table 1), and they are the primary components of the hydrostratigraphic units (HSU's) discussed below. These sedimentary-facies classes are defined primarily on the basis of grain-size distribution, mineralogy, sedimentary structures, and degree of postdepositional alteration. The secondary basis for facies-assemblage definitions is according to inferred environments of deposition. LFA's have distinctive geophysical, geochemical and hydrologic attributes, and they provide a mechanism for showing distribution patterns of major aquifers and confining units in hydrogeologic cross sections. In this study, basin and valley fills are subdivided into 13 major assemblages that are ranked in decreasing order of aquifer potential (tables 1 to 3; LFA's <u>*1-10*</u>, <u>*a-c*</u>). Figure 7-5 is a schematic illustration of the distribution pattern LFA's observed in the New Mexico Basin and Range Region. Lithofacies properties that influence groundwater flow and production potential in this region are summarized in tables 2 and 3. Note that *Roman numeral* notations (I–X) originally used in previous hydrogeologic framework models (Hawley and Lozinsky, 1992; Hawley and others, 1995) have been changed to *Arabic* style in order to facilitate the development of alphanumeric attribute codes that can be used in both conceptual and numerical models of basin-fill aquifer systems.

## Hydrostratigraphic Units

As already noted, most of the RGR basin fill in the south-central New Mexico border region have been subdivided into formation-rank, lithostratigraphic units of the Santa Fe Group (e.g., Hawley and others, 1969; Hawley, 1978; Gile and others, 1981), Seager and

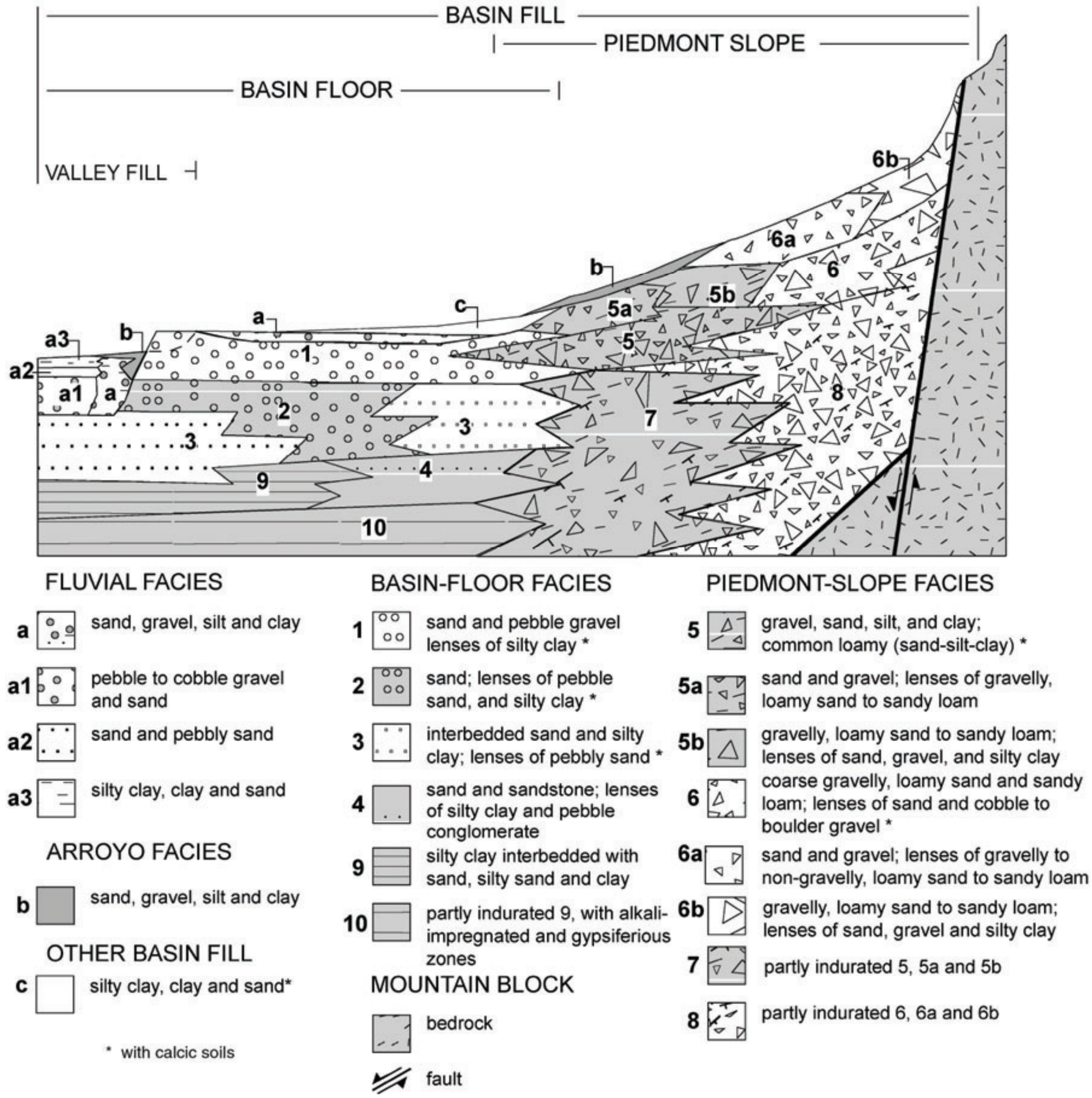


Figure 7-5: Schematic distribution pattern of major lithofacies assemblages (tables 1 through 3) in basin and valley fills of the Rio Grande rift region (from Hawley and others, 2000).

others, 1987; Mack and others, 1998b; Keller and Cather, 1994; Collins and Raney, 2000; Hawley and others, 2000). However, a clear distinction has rarely been made between deposits simply classed as “bolson” or “basin” fill and contiguous Santa Fe Group subdivisions. As a first step in organizing available information on basin-fill stratigraphy with emphasis on aquifer characteristics, a provisional hydrostratigraphic classification system has been developed during the past 20 yr that is applicable to all basins of the southeastern Basin and Range province.

Table 7-1: Stratigraphic units that comprise the aquifers of Loving, Pecos, Reeves, Ward and Winkler Counties.

| Lithofacies | | Dominant depositional settings and process | Dominant textural classes |
|---|---|---|---|
| 1 | | Basin-floor fluvial plain | Sand and pebble gravel, lenses of silty clay |
| 2 | | Basin-floor fluvial, locally eolian | Sand; lenses of pebble sand, and silty clay |
| 3 | | Basin-floor, fluvial-overbank fluvial-deltaic and playa-lake;eolian | Interbedded sand and silty clay; lenses of pebbly sand |
| 4 | | Eolian, basin-floor alluvial | Sand and sandstone; lenses of silty sand to clay |
| 5 | | Distal to medial piedmont-slope, alluvial fan | Gravel, sand, silt, and clay; common loamy (sand-silt-clay) |
| | 5a | Distal to medial piedmont-slope, alluvial fan; associated with large watersheds; alluvial-fan distributary-channel primary, sheet-flood and debris-flow, secondary | Sand and gravel; lenses of gravelly, loamy sand to sandy loam |
| | 5b | Distal to medial piedmont-slope, alluvial-fan; associated with small steep watersheds; debris-flow, sheet-flood and distributary-channel | Gravelly, loamy sand to sandy loam; lenses of sand, gravel, and silty clay |
| 6 | | Proximal to medial piedmont-slope, alluvial-fan | Coarse gravelly, loamy sand and sandy loam; lenses of sand and cobble to boulder gravel |
| | 6a | Like 5a | Sand and gravel; lenses of gravelly to non-gravelly, loamy sand to sandy loam |
| | 6b | Like 5b | Gravelly, loamy sand to sandy loam; lenses of sand, gravel, and silty clay |
| 7 | | Like 5 | Partly indurated 5 |
| 8 | | Like 6 | Partly indurated 6 |
| 9 | | Basin-floor --alluvial flat, playa, lake, and fluvial-lacustrine; distal-piedmaont alluvial | Silty clay interbedded with sand, silty sand and clay |
| 10 | | Like 9, with evaporite processes (paleophreatic) | Partly indurated 9, with gypsiferous and alkali-impregnated zones |
| a | | River-valley, fluvial | Sand, gravel, silt and clay |
| | a1 | Basal channel | Pebble to cobble gravel and sand (like 1) |
| | a2 | Braided plain, channel | Sand and pebbly sand (like 2) |
| | a3 | Overbank, meander-belt, oxbow | Silty clay, clay, and sand (like 3) |
| b | | Arroyo channel, and valley-border alluvial-fan | Sand, gravel, silt, and clay (like 5) |
| c | | Basin floor, alluvial flat, cienega, playa, and fluvial-fan to lacustrine plain | Silty clay, clay and sand (like 3,5, and 9) |

Hydrostratigraphic units defined in the RGR region are mappable bodies of basin fill and valley fill that are grouped on the basis of origin and position in both lithostratigraphic and chronostratigraphic sequences. The informal upper, middle, and lower Santa Fe hydrostratigraphic units (HSU's: USF, MSF, LSF) form the major basin-fill aquifer zones, and they correspond roughly to the (formal and informal) upper, middle, and lower lithostratigraphic subdivisions of the Santa Fe Group used in local and regional geologic mapping (fig. 7-6). Dominant lithofacies assemblages in the upper Santa Fe HSU are *LFA's 1-3, 5 and 6*. The middle Santa Fe HSU is characterized by *LFA's 3, 4, 7-9*, and the lower Santa Fe commonly comprises *LFA's 7-10*. Basin-floor facies assemblages *3* and *9* are normally present throughout the Santa Fe Group section in closed-basin (bolson) areas.

Table 7-2: Summary of properties that influence groundwater production potential of Gila and Santa Fe Group (modified from Haase and Lozinsky 1992) [>, greater than, <, less than].

| **Lithofacies** | Ratio of sand plus gravel to silt plus clay [1] | Bedding thickness (meters) | Bedding configuration [2] | Bedding continuity (meters) [3] | Bedding connectivity [4] | Hydraulic conductivity (K)[5] | Groundwater production potential |
|---|---|---|---|---|---|---|---|
| 1 | High | > 1.5 | Elongate to planar | > 300 | High | High | High |
| 2 | High to moderate | > 1.5 | Elongate to planar | > 300 | High to moderate | High to moderate | High to moderate |
| 3 | Moderate | > 1.5 | Planar | 150 to 300 | Moderate to high | Moderate | Moderate |
| 4 | Moderate to low* | > 1.5 | Planar to elongate | 30 to 150 | Moderate to high | Moderate | Moderate |
| 5 | Moderate to high | 0.3 to 1.5 | Elongate to lobate | 30 to 150 | Moderate | Moderate to low | Moderate to low |
| 5a | High to moderate | 0.3 to 1.5 | Elongate to lobate | 30 to 150 | Moderate | Moderate | Moderate |
| 5b | Moderate | 0.3 to 1.5 | Lobate | 30 to 150 | Moderate to low | Moderate to low | Moderate to low |
| 6 | Moderate to low | 0.3 to 1.5 | Lobate to elongate | 30 to 150 | Moderate to low | Moderate to low | Low to moderate |
| 6a | Moderate | 0.3 to 1.5 | Lobate to elongate | 30 to 150 | Moderate | Moderate to low | Moderate to low |
| 6b | Moderate to low | 0.3 to 1.5 | Lobate | < 30 | Low to moderate | Low to moderate | Low |
| 7 | Moderate* | 0.3 to 1.5 | Elongate to lobate | 30 to 150 | Moderate | Low | Low |
| 8 | Moderate to low * | > 1.5 | Lobate | < 30 | Low to moderate | Low | Low |
| 9 | Low | > 3.0 | Planar | > 150 | Low | Very low | Very low |
| 10 | Low* | > 3.0 | Planar | > 150 | Low | Very low | Very low |

[1] High >2; moderate 0.5-2; low < 0.5

[2] Elongate (length to width ratios > 5); planar (length to width ratios 1-5); lobate (asymmetrical or incomplete planar beds).

[3] Measure of the lateral extent of an individual bed of given thickness and configuration.

[4] Estimate of the ease with which groundwater can flow between individual beds within a particular lithofacies. Generally, high sand + gravel/silt+ clay ratios, thick beds, and high bedding continuity favor high bedding connectivity. All other parameters being held equal the greater the bedding, connectivity, the greater the groundwater production potential of a sedimentary unit (Hawley and Haase 1992, VI).

[5] High 10 to 30 m/day; moderate, 1 to 10 m/day; low, < 1 m/day; very low, < 0.1 m/day.

* Significant amounts of cementation of coarse-grained beds (as much as 30%)

The other major hydrostratigraphic units comprise channel and floodplain deposits of the Rio Grande (HSU–RG) and its major tributaries. These valley fills of Late Quaternary age (<130 ka) form the upper part of the region's most productive shallow-aquifer system. Surficial lake and playa deposits, fills of larger arroyo valleys, and piedmont-slope alluvium are primarily in the *vadose* zone. However, they locally form important groundwater discharge and recharge sites. Historical *phreatic* conditions exist, or have recently existed, in a few playa remnants of large pluvial lakes of Late Quaternary age (Hawley, 1993). Notable examples are gypsum or alkali flats in the Tularosa, Jornada del Muerto, and Los Muertos Basins, which are contiguous to but outside the area of discussion.

Table 7-3: Summary of properties that influence groundwater production potential of Gila and Santa Fe Group [>, greater than, <, less than].

| Lithofacies | Ratio of sand plus gravel to silt plus clay [1] | Bedding thickness (meters) [3] | Bedding configuration [2] | Bedding continuity (meters) [3] | Bedding connectivity [4] | Hydraulic conductivity (K) [5] | Groundwater production potential |
|---|---|---|---|---|---|---|---|
| a | High to moderate | > 1.5 | Elongate to planar | > 300 | High to moderate | High to moderate | High to moderate |
| a1 | High | > 1.5 | Elongate to planar | > 300 | High | High | High |
| a2 | High to moderate | > 1.5 | Planar to elongate | 150 to 300 | Moderate to high | Moderate | Moderate |
| a3 | Moderate to low | > 1.5 | Planar to elongate | 30 to 150 | Moderate to high | Moderate to low | Moderate to low |
| b | Moderate to low | 0.3 to 1.5 | Elongate to lobate | <100 | Moderate | Moderate to low | Moderate to low |
| c | Low to moderate | 0.3 to 1.5 | Elongate to lobate | 30 to 150 | Low | Low | Low |

[1] High >2; moderate 0.5-2; low < 0.5

[2] Elongate (length to width ratios > 5); planar (length to width ratios 1-5); lobate (asymmetrical or incomplete planar beds).

[3] Measure of the lateral extent of an individual bed of given thickness and configuration.

[4] Estimate of the ease with which groundwater can flow between individual beds within a particular lithofacies. Generally, high sand + gravel/silt+ clay ratios, thick beds, and high bedding continuity favor high bedding connectivity. All other parameters being held equal the greater the bedding, connectivity, the greater the groundwater production potential of a sedimentary unit (Hawley and Haase 1992, VI).

[5] High 10 to 30 m/day; moderate, 1 to 10 m/day; low, < 1 m/day; very low, < 0.1 m/day.

## Bedrock and Structural Boundary Components

Structural and bedrock features that influence aquifer composition and behavior include basin-boundary mountain uplifts, bedrock units beneath the basin fill, fault zones and flexures within and at the edges of basins, and igneous-intrusive and extrusive rocks that penetrate or are interbedded with basin fill. Tectonic evolution of the fault-block basins and ranges of the Mesilla Basin area (many with a half-graben structure and accommodation-zone terminations) has had a profound effect on the distribution of lithofacies assemblages and the timing and style of emplacement of all major hydrostratigraphic units (figs. 7-5, 7-6). Discussion of this topic is beyond the scope of this paper, however, and the reader is referred to particularly pertinent reviews in Seager and Morgan (1979), Keller and Cather (1994), Faulds and Varga (1998), and Mack and others (1998a). Moreover, most of the significant bedrock- and structural-boundary features in the area are well documented on geologic maps and sections by Seager and others (1987), Seager (1995), Woodward and Myers (1997), and Collins and Raney (2000).

# Mesilla Basin Aquifer Systems

Figure 7-7 is a schematic hydrogeologic cross section of the south-central Mesilla Basin, which is approximately aligned along the 32nd parallel. The section is based on (1) geologic mapping, primarily by Seager and others (1987) and Seager (1995), and (2) subsurface geophysical, hydrogeologic, and water-quality information compiled by Hawley and Lozinsky (1992). Major contributors to the hydrogeologic interpretations shown in figure 7-6 include Leggat and others (1962), Cliett (1969), Hawley and others (1969), King and others (1971), Wilson and others (1981), Peterson and others (1984),

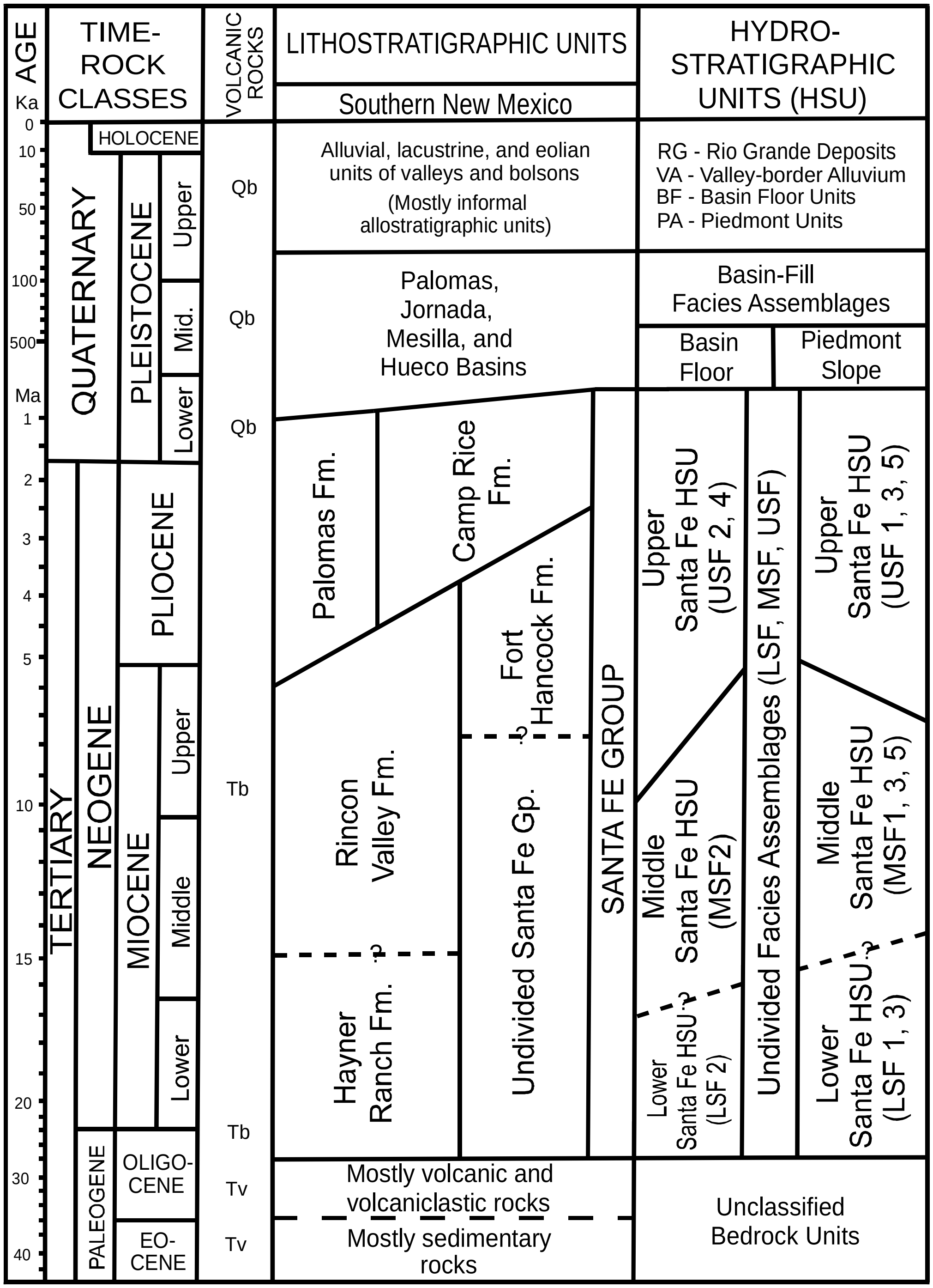


Figure 7-6: Regional summary and correlation of major chronlogic, lithostratigraphic, and basin-fill hydrostratigraphic units in the Messilla Basin-Hueco Bolson region of southern New Mexico and Trans-Pecos Texas. Igneous rock symbols: Qb-Quarternary basalt, Tb-Tertiary mafic volcanics, and Tv-older Tertiary intermediate and silcic volcanics and associated plutonic and sedimentary rocks (modified from Hawley and Kernodle, 2000).

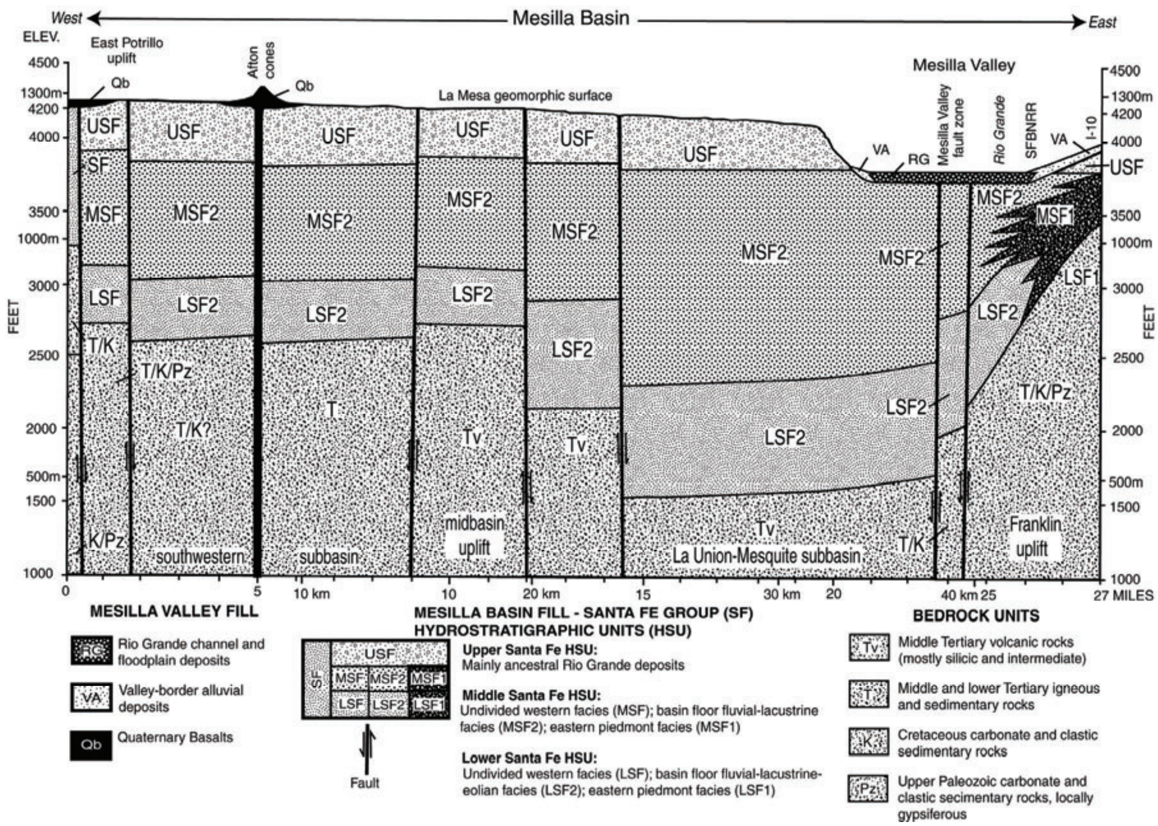


Figure 7-7: Schematic hydrogeologic cross section of the south-central Mesilla Basin near the 32nd Parallel in Doña Ana County, New Mexico, and El Paso County, Texas; with a vertical exaggeration of about 10× (modified from Hawley and Lozinsky, 1992).

Wilson and White (1984), Myers and Orr (1986), Seager and others (1987), Seager (1995), and Tom Cliett (EP-WUD) and Ken Stevens (USGS-WRD)—unpublished.

Major aquifers in the Mesilla Basin groundwater system occur in (1) Upper Quaternary alluvium of the inner river valley (valley-fill aquifer) and (2) poorly consolidated sedimentary deposits of the Santa Fe Group (basin-fill aquifer). The surface-water supply is derived from the Rio Grande, a few large tributary arroyo systems, and a network of canals, laterals, and drainage ditches that discharge to the river. The watershed of the Mesilla Basin covers approximately 11,000 $mi^2$.

## Mesilla Valley Aquifer System

The Rio Grande "alluvial" aquifer (fig. 7-7, HSU-RG, *LFA's a & b*) underlies the Mesilla Valley floor between Leasburg Dam and the El Paso narrows. This hydrostratigraphic unit comprises river-channel and overbank facies ranging in texture from sand and gravel to silt and clay. The base of these fluvial deposits is about 60 to 80 ft below the inner-valley floor, which is locally as much as 5 mi wide. In many places, the fluvial facies extends laterally for hundreds of feet beyond the valley floor. The basal-channel gravel and sand layer, which is as much as 30 to 40 ft thick, was deposited during the interval of maximum valley incision near the end of the Late Pleistocene ice age (about 15 to 30 thousand yr ago). The valley-fill HSU extends continuously from Elephant Butte and Caballo reservoirs, through the Rincon and Mesilla Valleys, to the Fort Quitman area at the lower end of the Hueco Bolson.

Groundwater within the Mesilla Valley fill is generally unconfined and typically moves southward down the valley at an average gradient of about 4 to 6 ft per mile; however, local-flow direction is influenced by nearby hydraulic conditions, such as the river, drains, canals, well pumpage, and heavily irrigated fields. The water table is approximately 10 to 25 ft below the land surface in much of the valley-floor area. Recharge to the valley-fill aquifer occurs primarily as vertical flow from the surface water system (river, canals, laterals, and drains) and irrigated cropland fields except in times of extreme drought. The inner-valley aquifer zone is, in turn, the major source of recharge to underlying and laterally adjacent basin fill of the Santa Fe Group. Most of the discharge from the valley fill occurs through evapotranspiration of irrigated crops, flow to drain system, and irrigation and industrial pumping. Transmissivity values range from 10,000 to 30,000 $ft^2/d$, hydraulic conductivities vary from 100 to 350 ft/d, and estimated specific yield is 0.2. Specific capacities of large production wells range from 10 to 217 gpm/ft of drawdown, with an average value of 69 gpm/ft of drawdown. The quality of the water generally reflects the quality of the surface-water system, ranging from about 500 mg/L TDS to over 1,000 mg/L TDS. At the extreme southern end of the Mesilla Valley, however, TDS values locally exceed 10,000 mg/L.

## Major Aquifer Zones in Santa Fe Group Basin Fill

A distinctive feature of Santa Fe Group basin fill in the Mesilla Basin is that it is relatively thin (maximum saturated thickness about 3,000 ft) in comparison to fills in adjacent parts of the Hueco-Tularosa and Mimbres Basin systems. Moreover, the major sources of fresh–to–slightly saline groundwater in the Mesilla Basin are from basin-floor facies assemblages in the middle to upper parts of the fill sequence. The dominant central-basin facies group comprises (1) thick sequences of fine-grained alluvial and lacustrine sediments that interfinger with and is overlapped by (2) coarser grained, ancestral-river deposits. Along basin margins both of these facies units are transitional with piedmont-slope alluvium (fig. 7-7).

The most-productive aquifer zones vary in thickness from about 300 ft in the northern and southernmost parts of the basin to over 2,000 ft in and near the eastern basin sector, which underlies the Mesilla Valley corridor from the Las Cruces metropolitan area to near Canutillo, Texas and La Union, New Mexico. Basic aquifer properties of the Mesilla Basin fill are very similar to properties of Hueco-Tularosa and Jornada Basin fills. The extent of these partly connected aquifer systems and the amount of interbasin groundwater flow is controlled in great part by the hydraulic properties of basin-boundary faults and lithofacies distribution patterns (depending, of course, on existing flow gradients). Fault zones and fine-grained facies commonly form effective barriers to interbasin flow. However, a small amount of flow may enter or leave the basin at low barrier points associated with zones of relatively high permeability.

The Mesilla Basin aquifer system comprises three major hydrostratigraphic subdivisions (HSU's) of the Santa Fe Group (Hawley and others, 1969; Hawley, 1978, charts 1 and 2). These units are ordered in upper to lower (younger to older) stratigraphic sequence (fig. 7-6). The upper Santa Fe unit (USF1,2) is generally correlative with the Camp Rice Formation (fig. 7-6), and its most productive aquifer zone consists of ancestral Rio Grande channel sand and gravel (HSU-USF2). However, the lower part of this unit is only saturated in the northeastern basin area near Las Cruces (Hawley and Lozinsky, 1992). The middle Santa Fe unit (MSF1,2) correlates with much of the Fort Hancock Formation in the Hueco Bolson, which is dominated by fine-grained, alluvial-flat, and playa-lake sediments. In the Mesilla Basin, however, the dominant facies assemblage (MSF2) includes extensive layers of clean sand that are interbedded with silty clay. The middle unit is less permeable than the upper unit because of a greater degree of cementation and the widespread presence of the fine-grained interbeds. HSU-MSF2, however, probably forms the major aquifer zone in the basin because it is almost entirely below the water table. The long-recognized "medial aquifer" zone of Leggat and others (1962) below the southern Mesilla Valley forms part of this unit (Cliett, 1969).

The lower Santa Fe unit (LSF) is primarily fine grained and party consolidated throughout much of the basin, and it only forms a significant part of the aquifer system in the lower Mesilla Valley area that extends from near Mesquite, New Mexico to Canutillo, Texas and La Union, New Mexico. This LSF unit was first identified in the El Paso Water Utilities-Canutillo well field by Leggat and others (1962) and was informally named the "deep aquifer" zone (HSU-LSF 2, fig. 7-6). The major component of the zone is a distinctive eolian-sand facies (LFA 4) that intertongues mountainward with piedmont fanglomerates (LFA's 7-8) and basinward with basin-floor facies assemblages (LFA's 3, 9 and 10?). The latter facies are here interpreted as fluvial-deltaic and playa-lake deposits (fig. 7-5, table 7-1). The sand facies is locally as much as 600 ft thick, and its base ranges from 1,000 to 1,500 ft below the Mesilla Valley floor. This extensive basin-floor to distal piedmont-slope deposit is interpreted as a buried dune field with an extent and thickness similar to that of *los Médanos de Samalayuca* dune complex in north-central Chihuahua (Cliett, 1969; Schmidt and Marston, 1981; Wilson and others, 1981; Hawley and Lozinsky, 1992).

## Concluding Comments on Groundwater Flow and Quality Conditions

The near-surface components (general elevation and direction) of the groundwater-flow system are shown on figure 7-2. Hydraulic conditions range from unconfined to semiconfined to confined in most basin-fill aquifer zones. In the central part of the basin west of the Mesilla Valley, which is designated the West Mesa in many reports, a transmissivity of 5,900 $ft^2/d$ was calculated for a well screened at selected depth intervals between 710 and 1,210 ft. In the northern part of the West Mesa area, aquifer transmissivity was estimated to be 10,000 $ft^2/d$, with a (confined) storage coefficient of $2 \times 10^{-5}$. According to aquifer tests, maximum values of transmissivity in the central Mesilla Basin ranged from 10,900 to 40,000 $ft^2/d$. The average horizontal hydraulic conductivity was 67 ft/d. This range in values, however, is probably only representative of the upper to middle parts of the Santa Fe Group aquifer system because these aquifer tests also provided evidence that the horizontal hydraulic conductivity decreases with depth. Vertical hydraulic conductivity values were found to range from 0.21 ft/d to 3.0 ft/d for the entire thickness of the confining layers at the aquifer-test sites.

Because of the limited scope of this paper, only a few comments on groundwater quality can be made. Water quality in the upper Santa Fe unit (HSU-USF2) in the eastern part of the basin generally reflects groundwater chemistry in the shallow valley-fill aquifer (HSU-RG) because this unit is the most significant recharge source for the upper part of the basin-fill aquifer system. Much of the groundwater pumped for irrigation is derived from the unconfined to semiconfined part of the (shallow) aquifer system that includes the river-valley fill (RG) and contiguous parts of HSU's USF2 and MSF2. A major influence on basinwide spatial variability in quality is due to irregular distribution patterns of fine-grained confining zones. Water in the middle Santa Fe unit (MSF2**)** is generally of better quality than in overlying valley-fill and basin-fill units, particularly in the northern part of the basin. Near the basin's southern end, however, available information indicates a significant deterioration in groundwater quality. The middle unit is the most heavily developed aquifer zone in terms of public and private drinking-water production. Water quality in the lower Santa Fe unit (LSF) is generally poorer than the middle unit except beneath the Mesilla Valley area between Mesquite and Canutillo. The majority of the discharge from the lower Santa Fe unit occurs as municipal and industrial pumping in the Anthony to Canutillo, Texas, area.

On the basis of review of data in the Frenzel and Kaehler (1992) groundwater-flow model, Balleau (1999, p. 46) estimated that about 14 million acre-ft of available water is stored in the upper 100 ft of saturated basin fill in the West Mesa area (~ 360,000 acres in New Mexico). This value is about twice our estimate, which assumes an effective aquifer porosity of 20 percent. Because saturated parts of HSU's USF2 and MSF2 in the West Mesa area range up to 1,000 ft in thickness, there is an enormous amount of potable to slightly saline groundwater stored in this part of the basin. Available fresh to slightly saline water stored in the upper 1,000 ft of Santa Fe Group hydrostratigraphic units, much of it very old, is probably no more than $100 \times 10^6$ acre-ft. Moreover, it has probably not been effectively recharged during the past 10,000 to 15,000 yr, except in areas contiguous to major streams.

The majority of recharge occurs through mountain-front mechanisms and through vertical groundwater flow from river-valley fill that forms the "shallow" alluvial aquifer. Except for a few perennial springs and seeps and short reaches of intermittent mountain streams, there are no permanent surface-water bodies in the small highland watersheds that flank the Mesilla Basin. Mountain-front recharge is, therefore, very low; and losing reaches of the Rio Grande channel and associated irrigation-canal systems are the major present sources of groundwater replenishment. Annual aquifer recharge in the 1,100-$mi^2$Mesilla Basin, exclusive of the 215-$mi^2$ Mesilla Valley area, is probably less than 10,000 acre-ft.

This estimate is based on the assumption that about 2 percent of the mean annual precipitation of 8 to 9 inches actually contributes to recharge outside the inner river valley. It must be noted in conclusion that present and projected basinwide groundwater use greatly exceeds this amount.

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**Chapter 8**

# An Overview of the Edwards-Trinity Aquifer System, Central–West Texas

Roberto Anaya[1]

## Introduction

Following a statewide drought during 1996, the passage of Senate Bill 1 in 1997 created a renewed public interest in the state's water resources not experienced since the 1950's drought of record. Senate Bill 1 also provided State funding to initiate the development of groundwater availability models for all of the major aquifers of Texas. The development and management of Groundwater Availability Modeling (GAM) has been tasked to the Texas Water Development Board (TWDB) to provide reliable and timely information on the state's groundwater resources. TWDB staff is currently conducting a GAM study for the Edwards-Trinity aquifer system, due for completion in late 2002. For most of West Texas, the 1996 drought has been prolonged and continues to impact West Texans. The purpose of this paper is to provide a brief overview of the Edwards-Trinity aquifer. Most of the information presented here is summarized from U.S. Geological Survey Regional Aquifer-Systems Analysis (RASA) reports.

## Geographic Setting

### Areal Extent

The Edwards-Trinity aquifer extends over an area of about 35,000 $mi^2$ and beneath all or parts of 38 counties (Ashworth and Hopkins, 1995) in central-west Texas (fig. 8-1). Most of the counties have relatively sparse populations concentrated in small towns, usually the county seats.

### Physiography

The aquifer sediments occupy the southeastern margin of the Great Plains physiographic province within an area known as the Edwards Plateau region and portions of the Trans-Pecos Basin and Range and the Llano Estacado regions. The area of the Edwards

[1] Texas Water Development Board

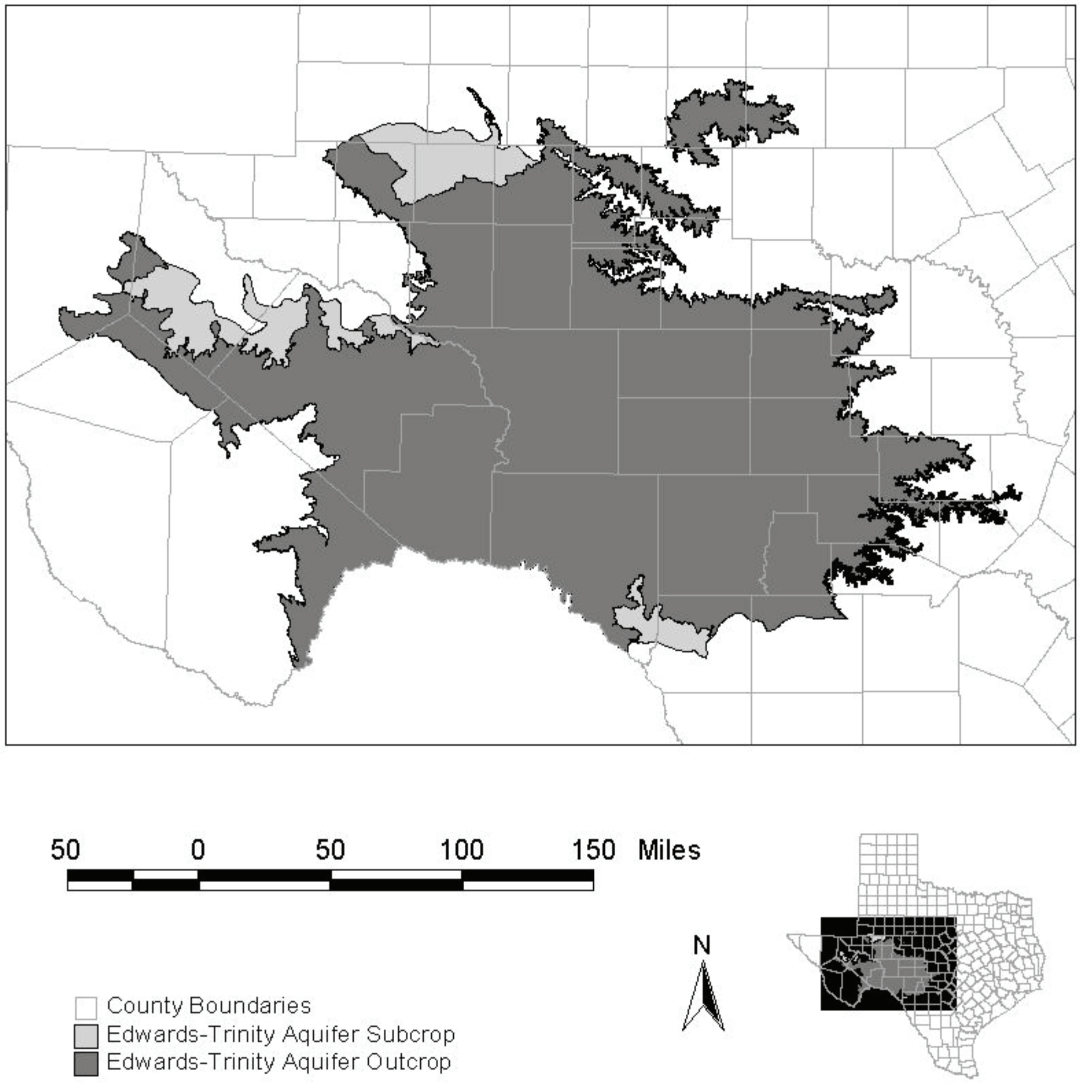



Figure 8-1: Areal extent of the Edwards-Trinity aquifer.

Plateau southwest of the Pecos River is often referred to as the Stockton Plateau. Northwest of the Pecos River, the area is commonly referred to as the Toyah Basin.

## Landforms

The headward erosion of streams transecting the Balcones Escarpment form canyon lands traditionally known as the Texas Hill Country that delineate the southern and eastern edge of the Edwards Plateau. The eastern flanks of the Delaware, Apache, Davis, Glass, and Santiago Mountains mark the western edge of the plateau. Playa lakes, characteristic

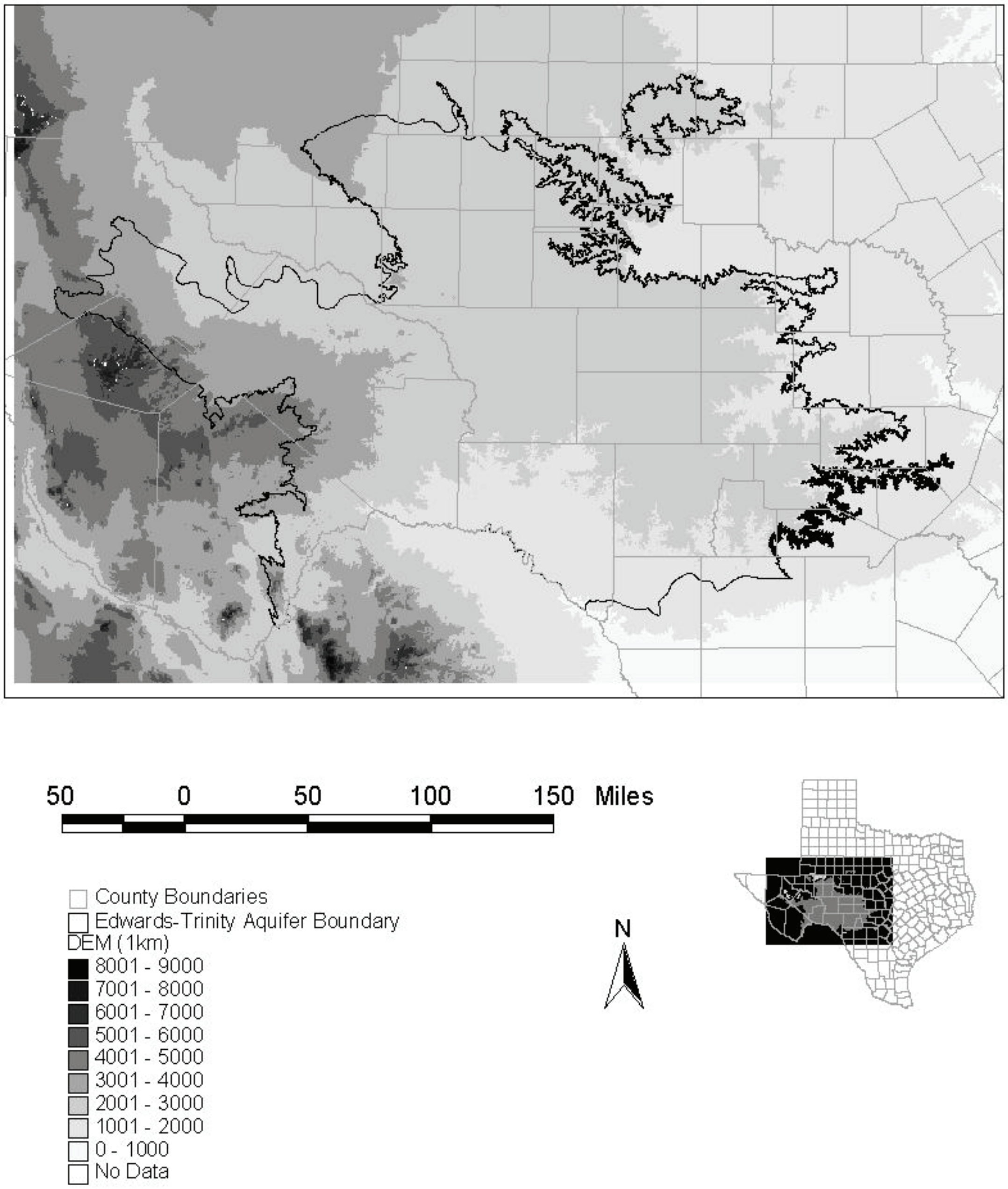



Figure 8-2: Surface topography of the Edwards Plateau.

of the High Plains, extend down into northern portions of the Edwards Plateau. The southwestern edge of the plateau extends into the northeastern margins of the Chihuahua Desert. The northeastern edge of the Edwards Plateau is adjacent to geologically complex Paleozoic and Precambrian landforms of the Central Mineral Region.

## Topography

The topography of the Edwards Plateau may be described as a flat tableland with stream-cut canyons in the southern and eastern portions of the plateau. Surface elevations range from about 5,000 ft near the mountains in the west to about 1,100 ft near the Rio Grande in the south (Barker and Ardis, 1996). The greatest surface relief occurs in the southern and western parts of the Edwards Plateau (Walker, 1979). Most of the plateau, however, has a flat surface sloping from the northwest to the southeast between 3,000 and 2,000 ft above sea level (fig. 8-2).

## Soils

Soils of the Edwards Plateau generally have thin and stony characteristics except in the northernmost portion of the plateau within the Llano Estacado region, where soils thicken into more sandy, loamy soils.

## Surface Drainage

Streams draining the Edwards Plateau have a dendritic or branchlike pattern, with stream density decreasing significantly toward the west (fig. 8-3). Tributary streams of the Colorado River, such as the Concho, San Saba, and Llano Rivers, drain the northeastern portion of the Edwards Plateau. Headwater tributaries of the Guadalupe, San Antonio, and Nueces Rivers drain the southern and southeastern portion of the plateau. The Pecos River and Devils River, major tributaries to the Rio Grande, drain the entire southwestern half of the Edwards Plateau. Although there are several small surface water bodies (<1 $mi^2$) in the central portion of the plateau, the only significant water bodies within the plateau are Big Lake in Regan County, Orient Reservoir in Pecos County, and Balmorhea Lake in Reeves County. Other much larger water bodies along the edge of the Edwards Plateau include Amistad Reservoir in Val Verde County, Twin Buttes and San Angelo Reservoirs in Tom Green County, and E. V. Spence Reservoir in Coke County.

## Climate

The climate ranges from subhumid in the eastern to semiarid in the western plateau (Walker, 1979). The long-term mean annual precipitation (1895–2000) ranges from about 25 inches in the east to about 12 inches in the west (fig. 8-4). Precipitation is greatest during late spring and early fall in the eastern two-thirds of the Edwards Plateau as a consequence of cool northern frontal air masses colliding with warm moist Gulf air masses from the south (fig. 8-5). However, in the western third of the plateau, most of the precipitation occurs during July, August, and September as a result of convectional thunderstorms (fig. 8-5). The variation of total monthly precipitation is greatest for the month of September throughout the plateau. Evaporation rates are high throughout the plateau and range between 43 inches in the east (Walker, 1979) to 80 inches (Rees and Buckner, 1980) in the west. Droughts are common on the Edwards Plateau, with about 10 moderate to severe droughts during the last 100 yr (fig. 8-5). The drought of record

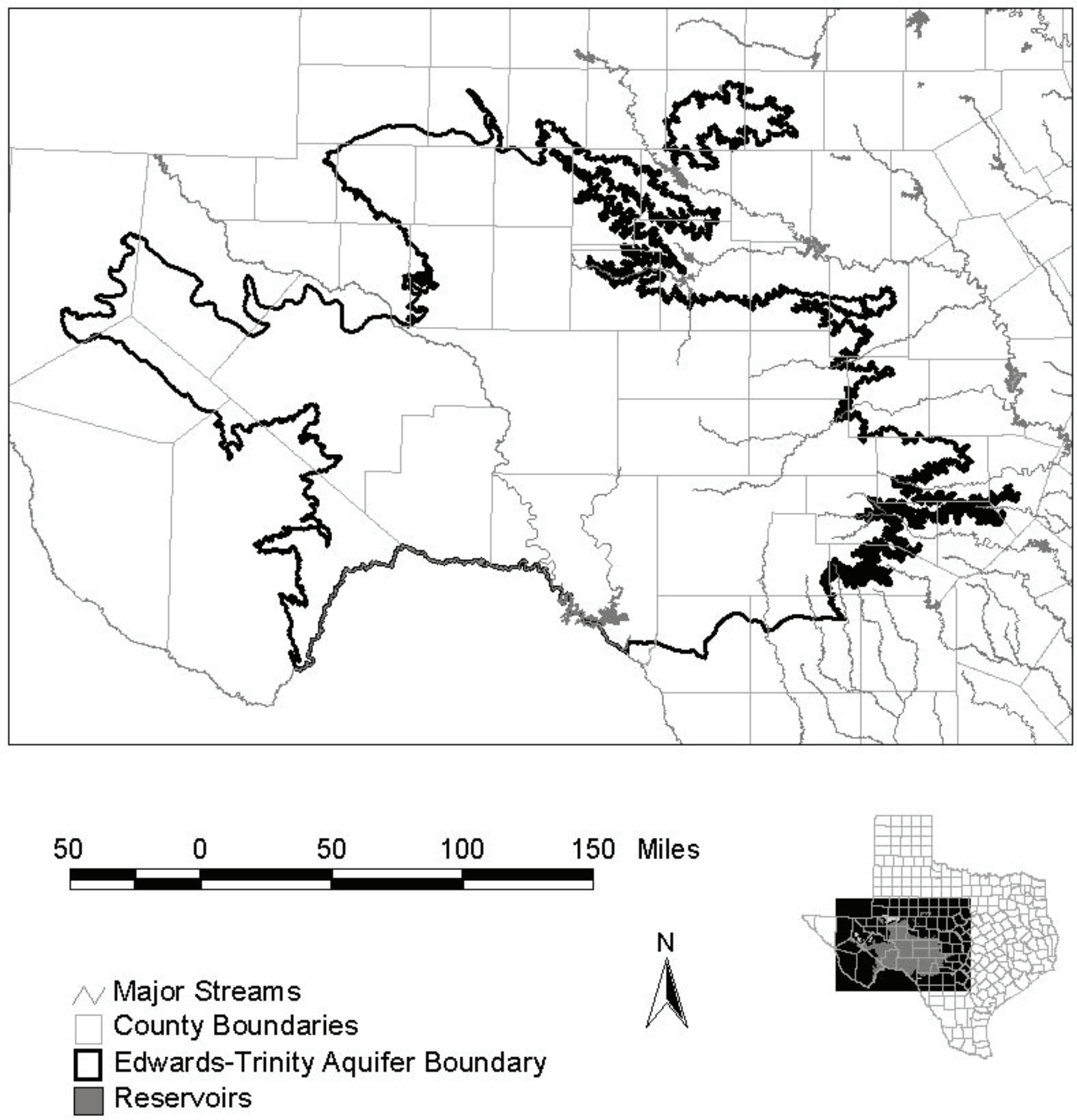
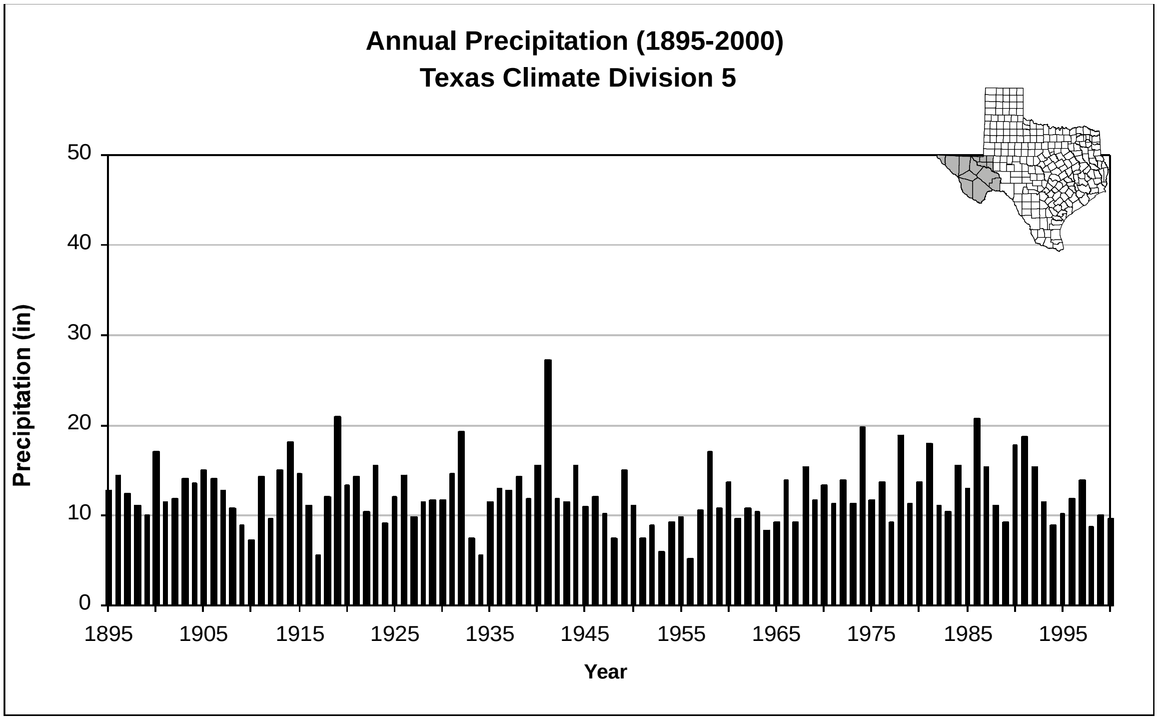



Figure 8-3: Surface-water drainage of the Edwards Plateau.

occurred during the 1950's, consistent with the rest of the state. However, the current drought that began during the mid- to late 1990's may potentially replace the 1950's record.

## Land Cover/Land Use

The Edwards Plateau is covered by scrubby savanna of oak-juniper-grass in the north and east and desert shrub and brush in the southwest. Salt cedar and other phreatophytes cover the stream valleys, contributing significant amounts of evapotranspiration. Some



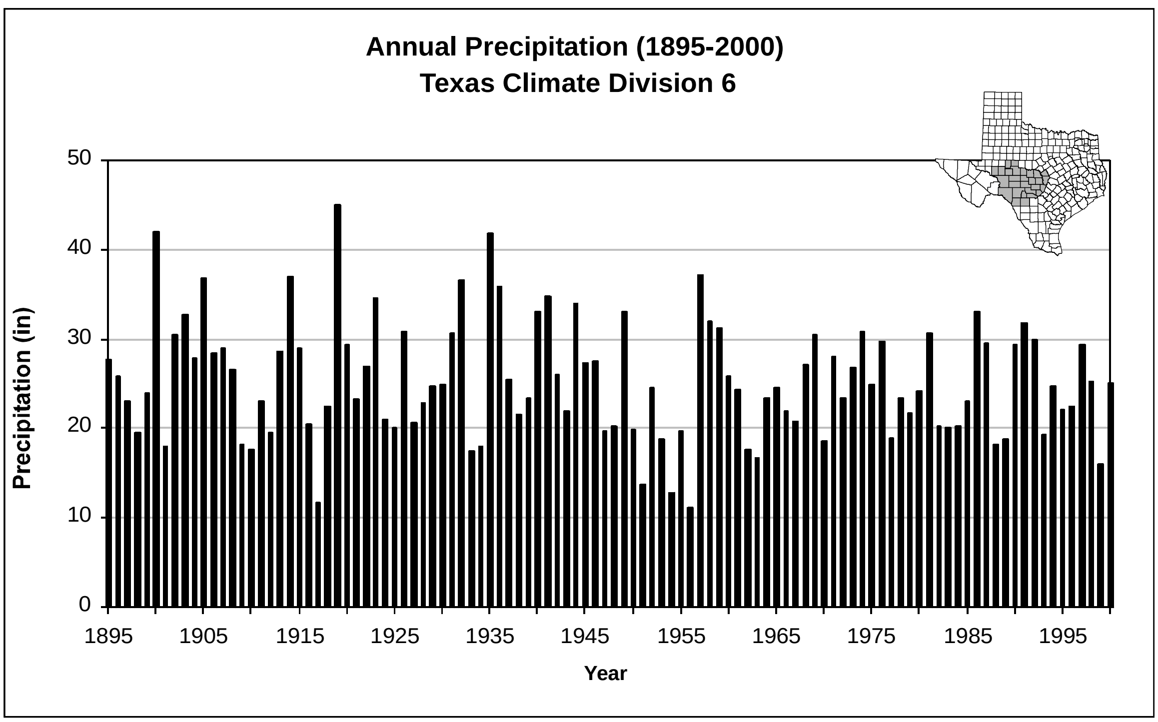
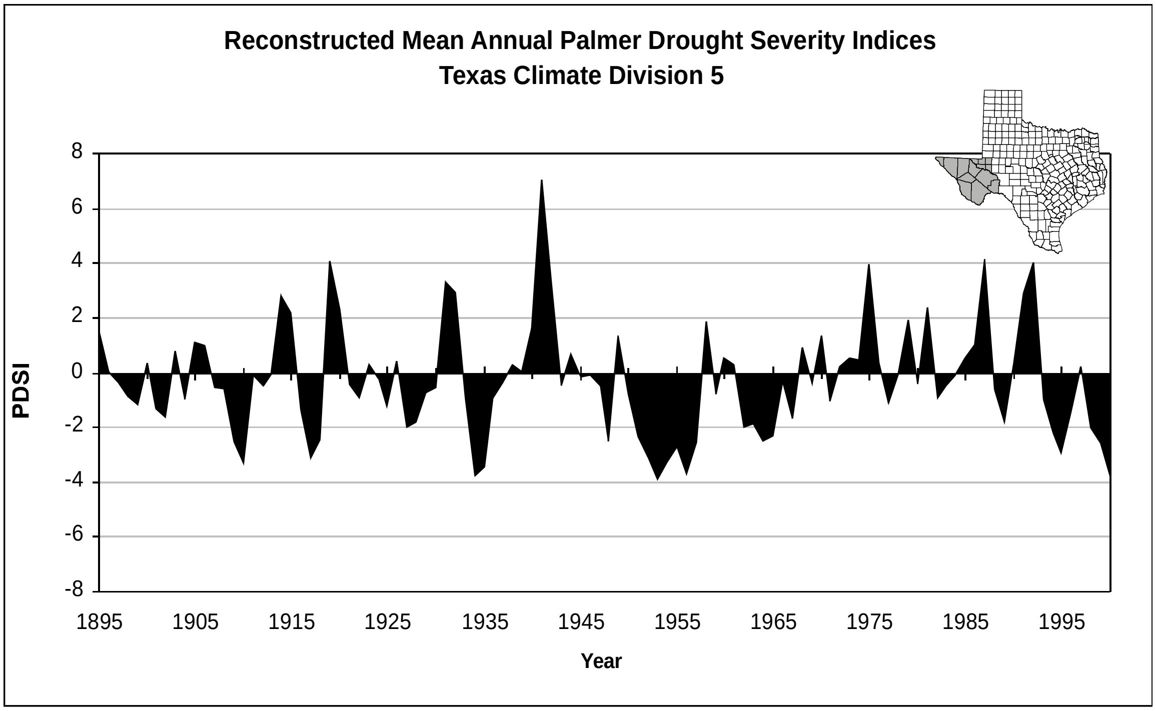




Figure 8-4: Long-term mean annual precipitation.



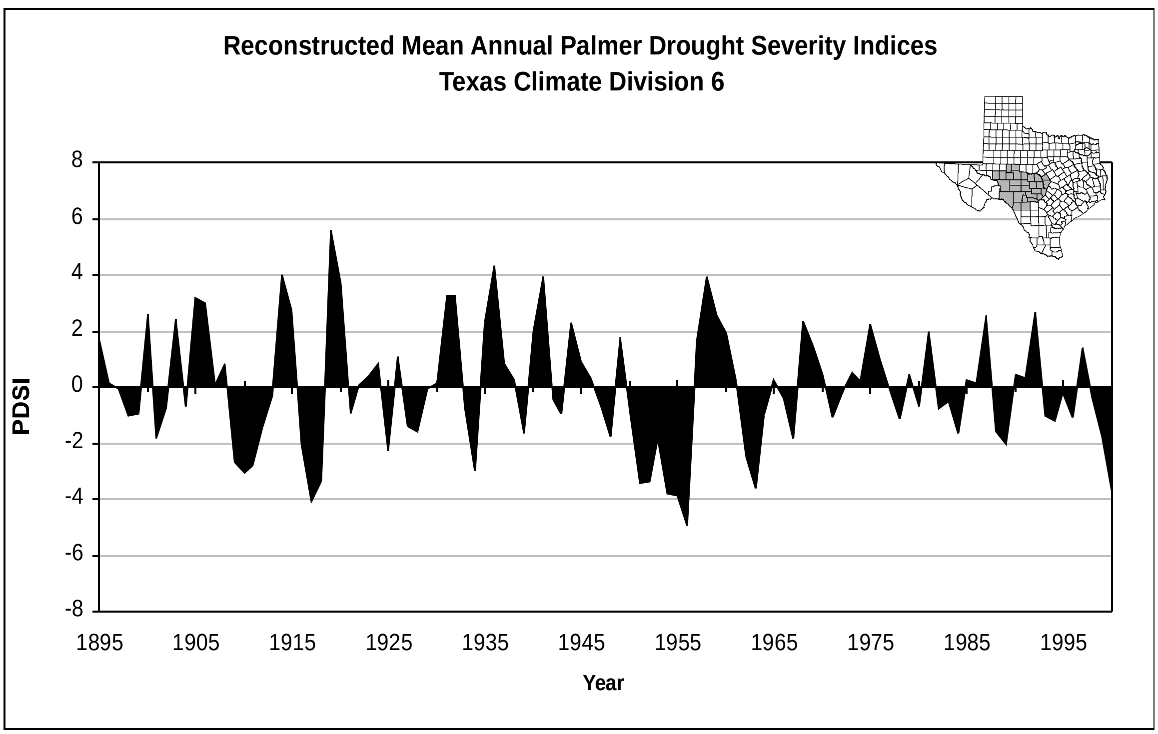




Figure 8-5: Long-term mean annual Palmer Drought Severity Indices.

native land cover has been replaced by cropland in northern portions of the Edwards Plateau. Ranching of cattle, sheep, and goats, along with wild-game-hunting leases are the principle land use, except for the northern portion of the plateau, where irrigated cropland is the dominant land use. Oil and gas production from the Permian Basin is common in the northern and western portions of the plateau.

## Groundwater Management Districts

The Edwards-Trinity aquifer falls within four Senate Bill 1 Regional Water Planning Groups (E, F, J, and K), although most of the aquifer falls within regions F and J. There are also about 24 groundwater-management districts with jurisdiction over the aquifer (fig. 8-6).

# Geologic History

## Paleozoic

The Paleozoic Era ended with a tectonic event known as the Ouachita Orogeny. The orogenic event resulted in the formation of a structural fold belt of sediments deposited during the Ordovician, Silurian, Devonian, and Mississippian Periods. The sediments were uplifted, faulted, and folded into a late Paleozoic mountain range that extended from northern Mexico along the present-day Balcones Escarpment up into the Ouachita Mountains of Oklahoma and Arkansas (Barker and Ardis, 1996). Before a final uplift during the Paleozoic Era, an arid and restricted shallow marine sea deposited Upper Permian sediments and evaporites into the Permian Basin of West Texas.

## Triassic

During the Triassic Period, terrigenous clastic red beds were deposited over the Paleozoic rocks as the Dockum Group sediments. The area of the Edwards Plateau was then exposed to erosion during the Jurassic Period to form a rolling peneplain known as the Wichita Paleoplain (Barker and Ardis, 1996). By the end of the Jurassic Period, the Gulf of Mexico had begun to open, and tilting of the peneplain toward the southeast provided a structural base for the deposition of the Cretaceous-age Edwards-Trinity sediments.

## Cretaceous

As the Gulf of Mexico continued to open and the Cretaceous seas advanced from the southeast, a broad continental shelf known as the Comanche Shelf began to form. The Llano Uplift, a tectonically active structural feature since the Precambrian, became a prominent structural shelf element for the deposition of the Trinity Group sediments (Barker and Ardis, 1996). The Early Cretaceous seas advanced across the pre-Cretaceous structural base in three cycles of transgressive-regressive stages to deposit the Trinity Group sediments (Barker and others, 1994). The Stuart City Reef trend began to form

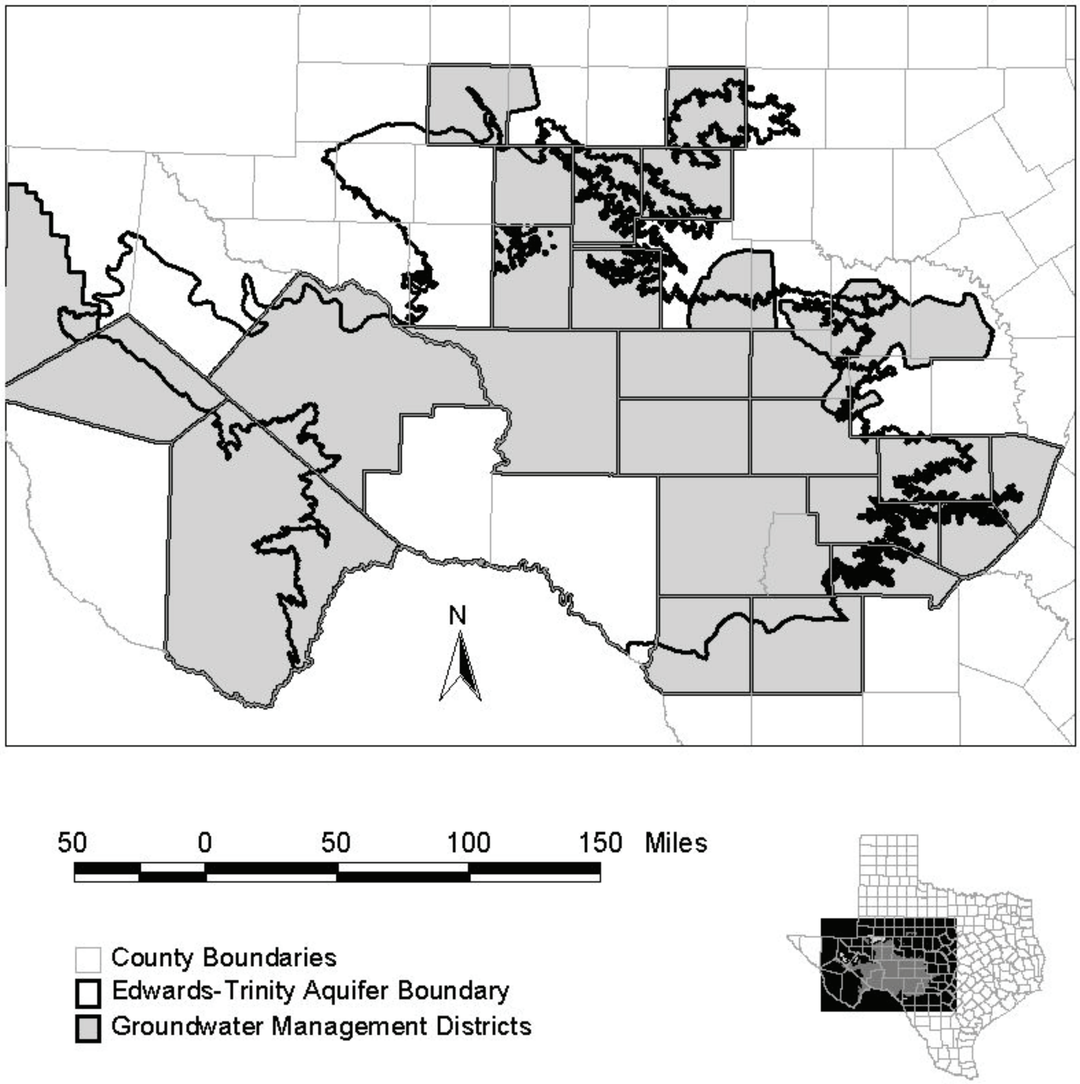



Figure 8-6: Groundwater management districts of the Edwards-Trinity aquifer.

along the present-day Balcones Fault Zone, enabling the carbonate platform deposits of the Edwards Group sediments to accumulate. Other structural shelf elements that formed behind the Stuart City Reef trend and controlled the depositional environments and lithologic characteristics of the Edwards Group formations include the Central Texas Platform, the San Marcos Arch, the Devils River Reef trend, Maverick Basin, and the Fort Stockton Basin. Prior to the deposition of Upper Cretaceous Del Rio, Buda, Boquillas, and Austin Group sediments, much of the Central Texas Platform was subaerially exposed, allowing for an initial karstification of Lower Cretaceous carbonate sediments (Barker and others, 1994).

## Cenozoic

Toward the end of the Cretaceous and beginning of the Tertiary Periods, the Laramide orogenic event and the dissolution of Upper Permian sediments resulted in the structural collapse and erosion of overlying Triassic and Cretaceous sediments along the Pecos River valley (Barker and others, 1994). These sediments were then redeposited as the Cenozoic Pecos Alluvium throughout the Tertiary and into the Quaternary Periods. The Ogallala sediments were deposited over a portion of the Edwards-Trinity sediments in the northern region of the plateau during the late Tertiary Period. During the mid-Tertiary Period, regional uplift and continued deposition of sediments into the Gulf of Mexico provided tensional stresses along the ancient Ouachita fold belt. Consequently, the development of the Balcones Fault Zone occurred and displaced Cretaceous and Lower Tertiary sediments by 900 to 1,200 ft (Barker and others, 1994). During the Quaternary, the headward erosion of streams began to reduce the plateau into its current form.

# Hydrogeologic Units

The vertical and lateral organization of the various hydrogeologic units of the Edwards-Trinity aquifer system is presented in a stratigraphic chart (fig. 8-7) and the following discussion.

## Paleozoic

The Hickory, the Ellenburger-San Saba, and the Marble Falls aquifers are hydraulically connected to the Edwards-Trinity aquifer system along the eastern margin of the plateau surrounding the Llano Uplift. The Permian-age Capitan and Rustler sediments are hydraulically connected to the Edwards-Trinity sediments in the Trans-Pecos portion of the aquifer (Bush and others, 1994). In general, most of the underlying Paleozoic rocks provide a relatively impermeable base for the Edwards-Trinity aquifer sediments (Barker and Ardis, 1992).

## Triassic

The Dockum Group consists of Lower (Tecovas Formation), Middle (Santa Rosa Formation), and Upper (Chinle Formation) members (Walker, 1979). Only where the Chinle Formation is missing, allowing for the Basal Cretaceous sands to be in hydraulic communication with the underlying Santa Rosa Formation, is the Santa Rosa Formation considered to be an aquifer (Walker, 1979).

## Cretaceous

The Trinity Group sediments are divided into Lower, Middle, and Upper Trinity sediments in the southeastern portion of the plateau (Ashworth, 1983). The Lower Trinity consists of the Hosston and Sligo Formations, the Sycamore Sand, and the Hammett Shale; the Middle Trinity consists of the Cow Creek Limestone, the Hensell Sand, and

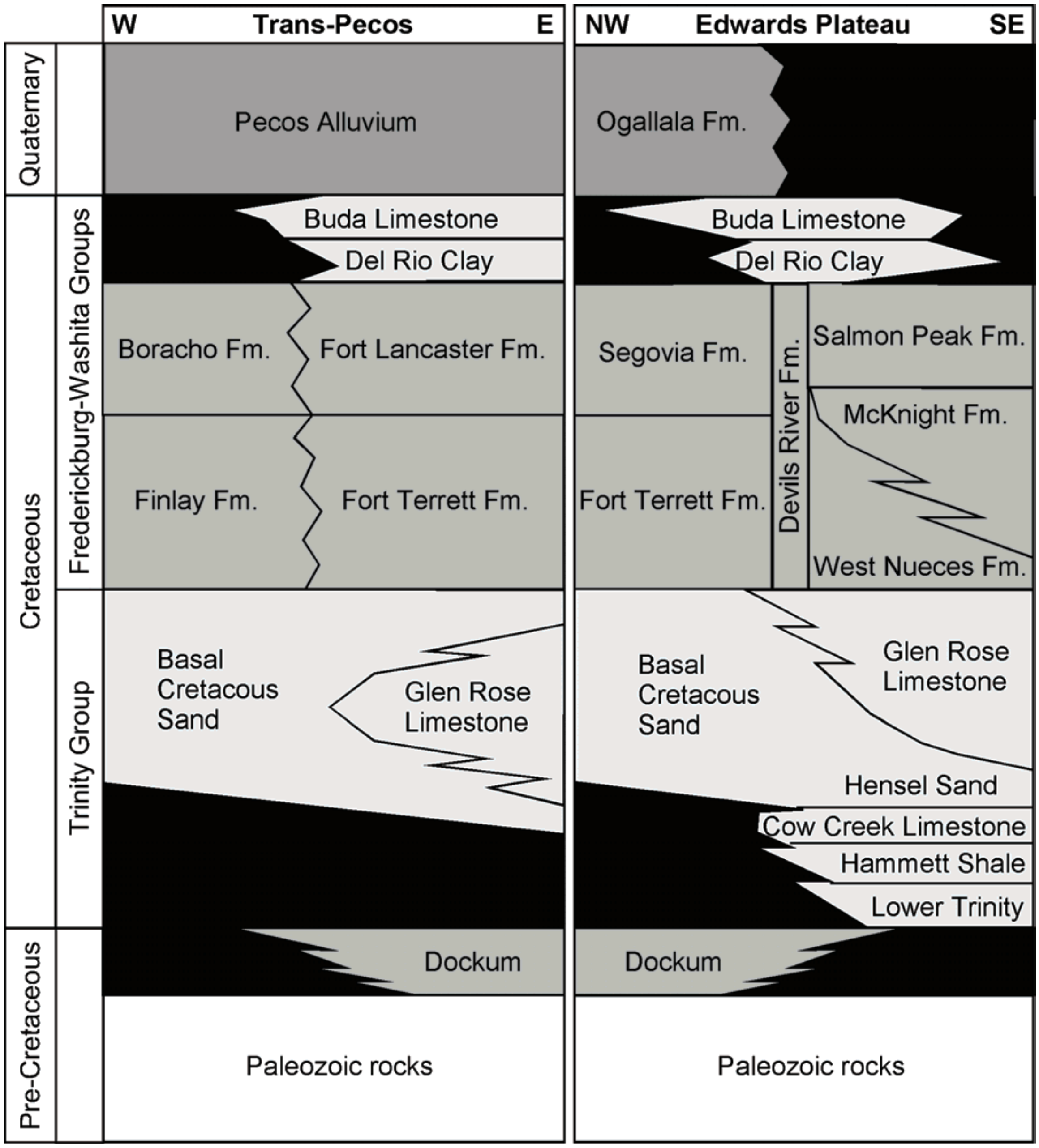



Figure 8-7: Stratigraphic chart for the hydrogeologic units of the Edwards-Trinity aquifer.

the lower member of the Glen Rose Limestone; and the Upper Trinity consists of the upper member of the Glen Rose Limestone (Mace and others, 2000). In the far northwest portion of the plateau, the Trinity Group sediments are divided into the Yearwood Formation and the Cox Sandstone. Elsewhere on the plateau, the Trinity Group sediments are divided into the Basal Cretaceous sand, the Glen Rose Limestone, and the Maxon

Sand. The Basal Cretaceous sand and Maxon Sand are sometimes lumped together and referred to as the Antlers Sand or Trinity Sands where the Glen Rose Limestone is absent.

The Edwards Group and equivalent sediments consist of the Fredericksburg and Washita Group sediments. The Fredericksburg Group is composed of the Finlay Formation within the Fort Stockton Basin, the Fort Terrett Formation within the central Texas platform, the Devils River Formation within the Devils River Reef trend, the West Nueces and McKnight Formations within the Maverick Basin, and the Kainer Formation within the San Marcos Arch. The Washita Group sediments are composed of the Boracho Formation within the Fort Stockton Basin, the Segovia Formation within the Central Texas Platform, the Devils River Formation within the Devils River Reef trend, the McKnight and Salmon Peak Formations within the Maverick Basin, and the Person Formation within the San Marcos Arch.

The Upper Cretaceous sediments include the uppermost section of the Washita Group sediments (the upper confining Del Rio Clay and the Buda Limestone), along with the Eagle Ford Group (Boquillas Formation) and the Austin Group sediments.

## Cenozoic

The Cenozoic Pecos Alluvium is hydraulically connected to the Edwards-Trinity aquifer in the northwestern edge of the aquifer. The late Tertiary-age Ogallala sediments are hydraulically connected only in the northernmost portion of the Edwards-Trinity aquifer.

# Aquifer Characteristics

## Structural Geometry

The initial base depositional surface of the Cretaceous sediments is generally flat and tilted toward the Gulf of Mexico, except for the area surrounding the Llano Uplift in the eastern plateau and areas of the western edge of the plateau along the eastern flanks of the mountains within the Trans-Pecos Basin and Range (fig. 8-8). Consequently, the Edwards-Trinity sediments form a wedge that thickens from the north and northwest toward the south and southeast. The exceptions to this structural trend are in the areas near the Llano Uplift and the mountains of the Trans-Pecos Basin and Range. The wedge of Cretaceous sediments pinches out beneath the Ogallala sediments in the northern portion of the plateau (Barker and Ardis, 1996). The Cretaceous wedge of sediments is terminated along the south and southeast by the Balcones Fault Zone. The entire Edwards Group and equivalent sediments and sections of the Upper Trinity sediments have been removed in the canyon-land areas of the Texas Hill Country. A small portion of the aquifer is confined in Val Verde County, where the Late Cretaceous-age Del Rio Clay overlies the Edwards Group sediments. The semipermeable Upper Cretaceous sediments of the Buda Limestone and Boquillas Formation form a thin cap over the Edwards Group and equivalent sediments in the central and southern portions of the aquifer.

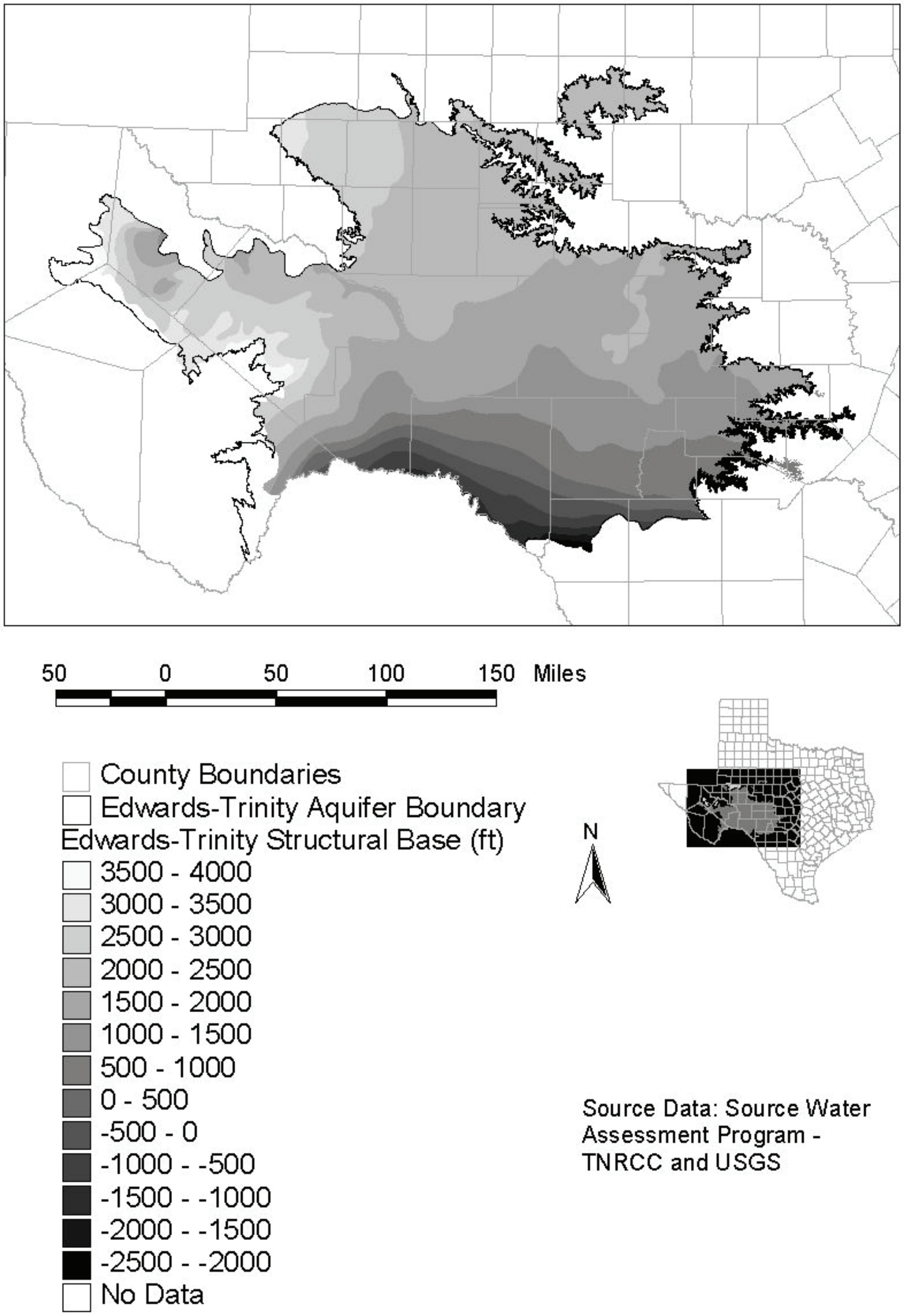



Figure 8-8: Structural base of the Edwards-Trinity aquifer.

## Water Levels

Although water levels are influenced by climate, they have remained fairly constant, except in areas of the northern and western plateau where a general trend of declining water levels is a result of increased irrigation pumpage (Ashworth and Hopkins, 1995). Long-term water levels of the Edwards-Trinity indicate the regional-flow groundwater within the aquifer system (fig. 8-9). There is a regional groundwater divide that trends from the northwest in Ector County to the southeast near the common boundary between Real, Kerr, and Edwards Counties that separates flow toward the Colorado River from flow toward the Rio Grande. The Pecos River valley has a significant influence on the groundwater flow in the western half of the plateau.

## Hydraulic Properties

The Edwards-Trinity aquifer is hydraulically connected to four major aquifers: (1) the Cenozoic Pecos Alluvium, (2) the Ogallala, (3) the Trinity (Hill Country), and (4) the Edwards (Balcones Fault Zone). The Edwards-Trinity aquifer is also hydraulically connected to several minor aquifers: (1) the Dockum, (2) the Capitan, (3) the Rustler, (4) the Hickory, (5) the Ellenburger-San Saba, (6) the Lipan, and, to a very small degree, (7) the Marble Falls.

The saturated thickness of less than 100 ft to greater than 800 ft for the Edwards-Trinity aquifer system generally increases from north to south and varies the greatest along the western margins of the aquifer (fig. 8-10). Gentle north-south-trending ridges and troughs in the pre-Cretaceous base depositional surface combined with the topographic influence on the water table control the variability in saturated thickness (Barker and Ardis, 1996). The aquifer is mostly under water-table or unconfined conditions, although the Trinity unit of the aquifer may be semiconfined locally where relatively impermeable sediments of the overlying Edwards Group exist (Ashworth and Hopkins, 1995).

Transmissivity is a function of the conductivity of the aquifer sediments and the saturated thickness. The Edwards-Trinity aquifer generally has transmissivity values of less than 5,000 $ft^2$/d in the north and eastern portions of the aquifer and values between 5,000 and 50,000 $ft^2$/d in the southern and western portions of the aquifer with an average of less than 10,000 $ft^2$/d (Barker and Ardis, 1996). Except for areas of significant karst-induced permeability, the average hydraulic conductivity of the Edwards-Trinity sediments is about 10 ft/d, judging from transmissivity and saturated thickness distributions (Barker and Ardis, 1996).

## Recharge

Most recharge occurs from the infiltration of precipitation over Edwards-Trinity outcrops and sinkholes and from stream losses of intermittent streams. Rees and Buckner (1980) estimated recharge over the Trans-Pecos region of the plateau to be between about 0.3 and 0.4 inches per year. Kuniansky (1989) estimated recharge over the eastern portion of the plateau to range between 0.12 and 2.24 inches per year. In general, recharge rates

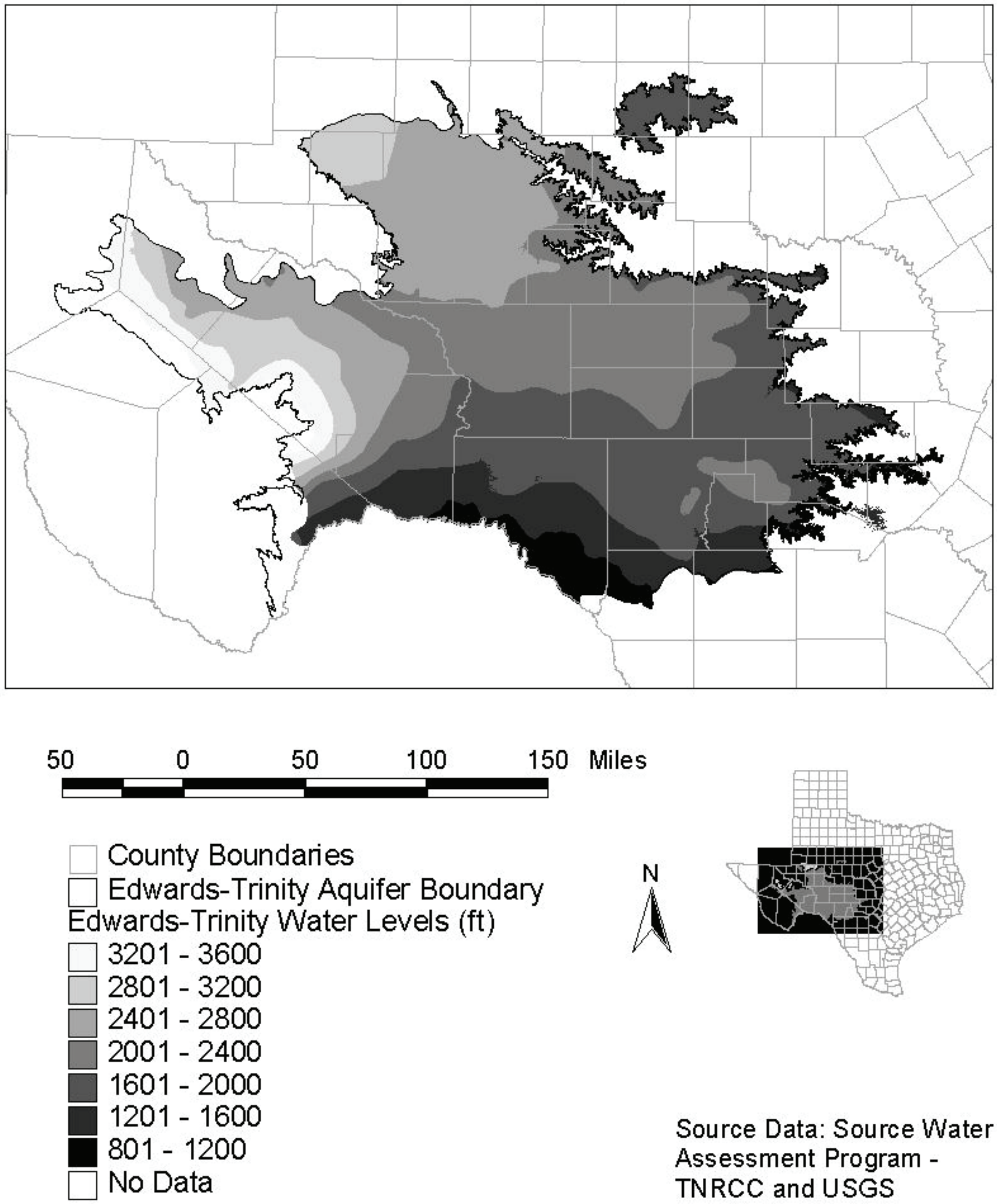



Figure 8-9: Historical water levels of the Edwards-Trinity aquifer.

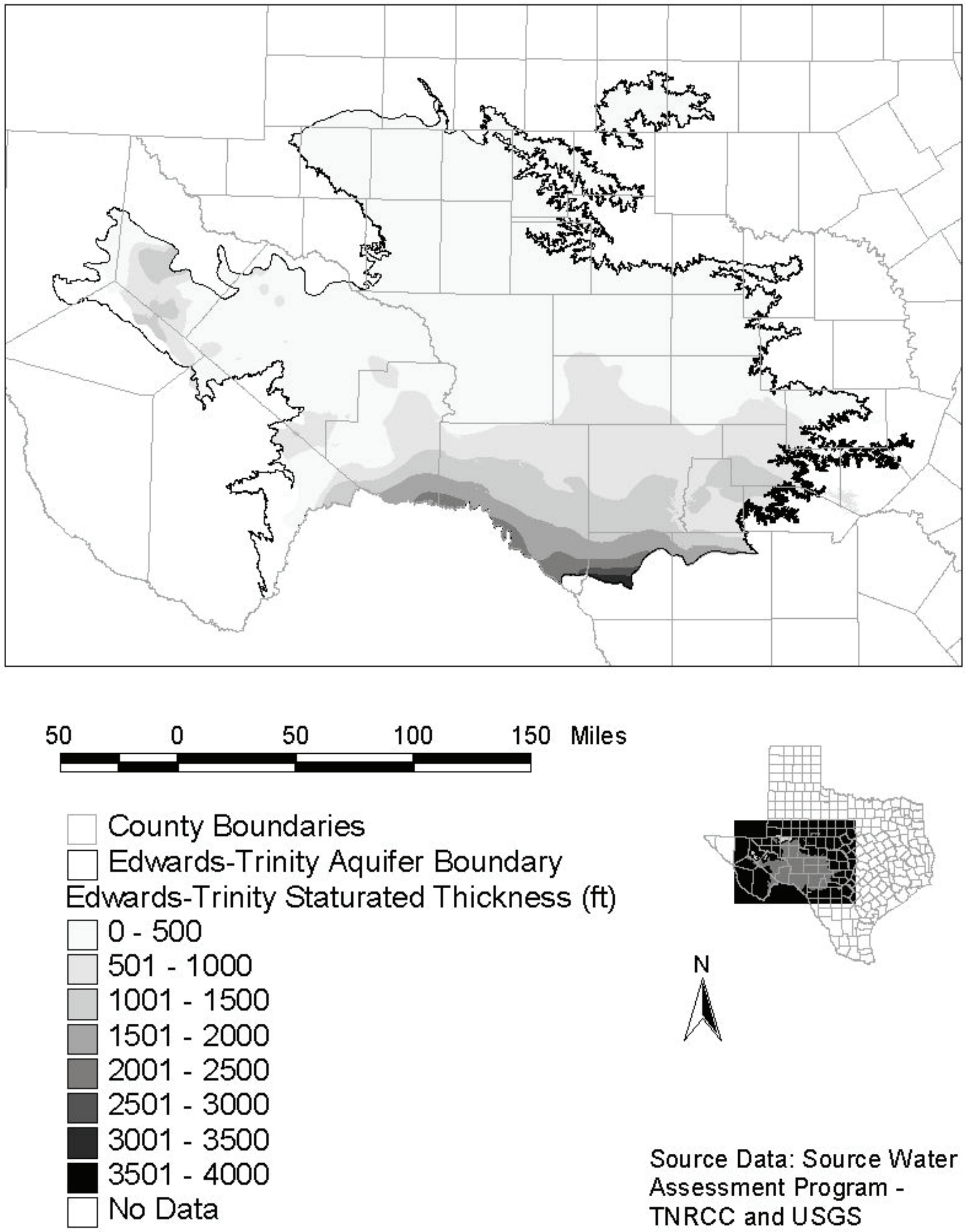



Figure 8-10: Saturated thickness of the Edwards-Trinity aquifer.

vary with climate conditions, surface topography, soils, and land cover/land use. Cross-formational flow from hydraulically contiguous major and minor aquifers also provides recharge to the Edwards-Trinity aquifer system, primarily in the northern and western portions of the aquifer. Induced recharge occurs in Pecos and Reeves Counties as a result of water-level declines due to irrigation pumpage (Barker and Ardis, 1996).

### Natural Discharge

Natural discharge from the Edwards-Trinity aquifer occurs mostly from springs where the water table intersects canyons or surface topography to provide base flow to streams. In addition, phreatophytes along major stream valleys discharge the aquifer naturally through evapotranspiration where the water table is shallow enough for the root networks. Cross-formational flow from hydraulically contiguous major and minor aquifers also provides natural discharge from the Edwards-Trinity aquifer system, primarily in the southern and eastern portions of the aquifer. As water levels have declined in the northern and western portions of the aquifer due to increased irrigation pumpage, spring flow within those areas has also declined.

## Groundwater Use

The Trinity Group sediments provide much of the water for the northern and western areas of the plateau, while the Fredericksburg Group sediments provide most of the water in the central, southern, and eastern portions of the plateau (Barker and Ardis, 1996). Over three-fourths of the total groundwater pumpage from the Edwards-Trinity is used for irrigation, primarily in the northern and western portions of the aquifer (fig. 8-11). Municipal water suppliers account for the second-most-common groundwater use, followed by minimal industrial, mining, livestock, and rural domestic uses. Climate has a significant effect on the amount of groundwater pumpage from the Edwards-Trinity aquifer because of the high percentage of irrigation use (compare fig. 8-5 with fig. 8-11).

## Water Quality

Although water quality is typically hard, it is generally fresh, except for areas in the Trans-Pecos where groundwater from Permian evaporite sediments and/or oil-field brines are able to mix with groundwater from the Edwards-Trinity aquifer (Rees and Buckner, 1980). Water quality is also affected by induced recharge from Pecos River stream losses (Barker and Ardis, 1996). East of the Pecos River, oil-field brines and agricultural runoff have a significant effect on the groundwater quality of the northern portion of the Edwards-Trinity aquifer (Walker, 1979).

## Past and Present Studies

Previous studies on the Edwards-Trinity aquifer began with countywide studies by the Texas Board of Water Engineers, Texas Water Commission, Texas Department of Water

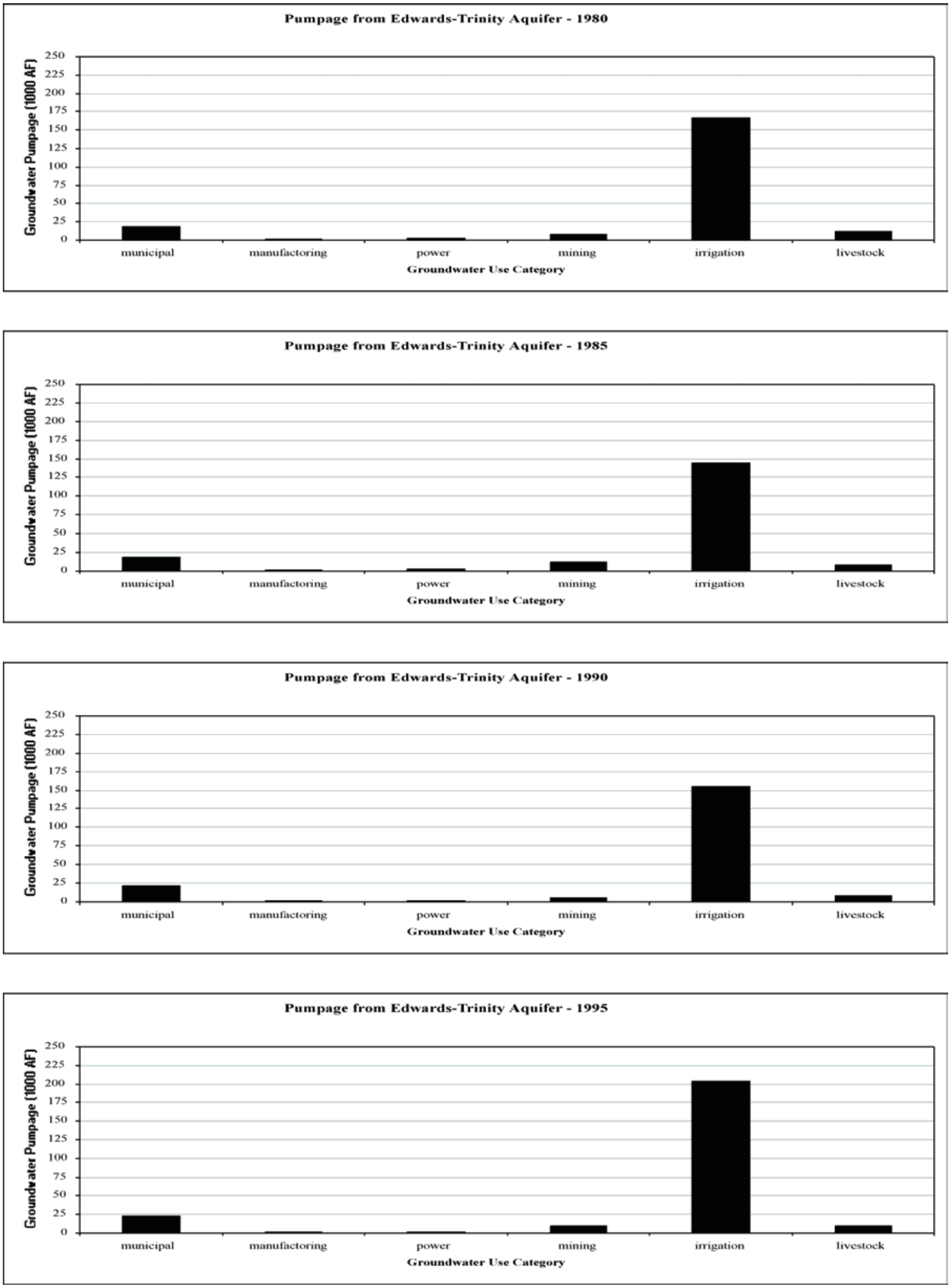



Figure 8-11: Groundwater pumpage from the Edwards-Trinity aquifer.

Resources, Texas Water Development Board, and the U.S. Geological Survey. The Texas Department of Water Resources (Walker, 1979; Rees and Buckner, 1980) was first to publish regional study reports on the Trans-Pecos and Plateau portions of the

Edward-Trinity aquifer. During the 1990's, the U.S. Geological Survey began a Regional Aquifer Systems Analysis (RASA) program for the Edwards-Trinity aquifer system, which resulted in the publication of the most recent and comprehensive reports on the Edwards-Trinity aquifer system, as well as a single-layer, finite-element, steady-state numerical model. Currently the Texas Water Development Board is conducting a comprehensive study to develop a state-of-the-art, multilayer, finite-difference numerical model of the Edwards-Trinity aquifer system, with a final report due for publication in late 2002. Information on this most recent study is updated and maintained at the following Internet Web address: http://www.twdb.state.tx.us/gam/.

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# Chapter 9

# Cenozoic Pecos Alluvium Aquifer

Ian C. Jones[1]

## Abstract

The Cenozoic Pecos Alluvium aquifer is an unconfined alluvium aquifer located in West Texas. This aquifer is composed of two main basins: the Pecos Trough and Monument Draw Trough. These basins form separate groundwater-flow systems because there is little or no interbasin flow. The Cenozoic Pecos Alluvium aquifer is most important as a source of irrigation water in parts of West Texas. The aquifer is naturally recharged by infiltration of precipitation and interaquifer flow, and natural discharge takes the form of base flow in the Pecos River, as well as evapotranspiration. Groundwater in this aquifer is generally slightly to moderately saline, exceeding drinking-water standards, with dissolved solids generally less than 5,000 mg/L. Groundwater quality is generally better in the Monument Draw Trough than in the Pecos Trough. Explanations for this are related to possibly higher recharge and lower irrigation pumpage rates in the Monument Draw Trough. The Cenozoic Pecos Alluvium aquifer displays effects of pumpage, mainly for irrigation. This pumpage has historically resulted in water-level declines of up to 200 ft, starting in 1940's, and has produced cones of depression in Reeves and Pecos Counties. Since the mid-1970's there has been some recovery owing to declining irrigation. A recent survey indicates that water levels in the Cenozoic Pecos Alluvium aquifer continue to recover in some areas previously impacted by irrigation pumpage. However, there still are areas, especially in Pecos and Ward Counties, where water levels are declining because of irrigation, public supply, and industrial pumpage.

## Introduction

The Cenozoic Pecos Alluvium aquifer is located in the upper Pecos River Valley of western Texas (fig. 9-1). This alluvium aquifer underlies parts of Crane, Ector, Loving, Pecos, Reeves, Upton, Ward, and Winkler Counties and extends north into New Mexico. The Cenozoic Pecos Alluvium occurs in a region with an arid climate characterized by average annual precipitation of 10 to 20 inches and high average annual evaporation rates approaching 70 inches (Boghici, 1999). These climatic conditions play an important role in determining the amount of water available for recharge to the aquifer. Climate and crop selection also play a role in determining the water demand for irrigation pumpage. In arid areas and especially during drought periods, irrigation pumpage increases to compensate for the absence of precipitation. This aquifer is of primary importance as a

[1] Texas Water Development Board

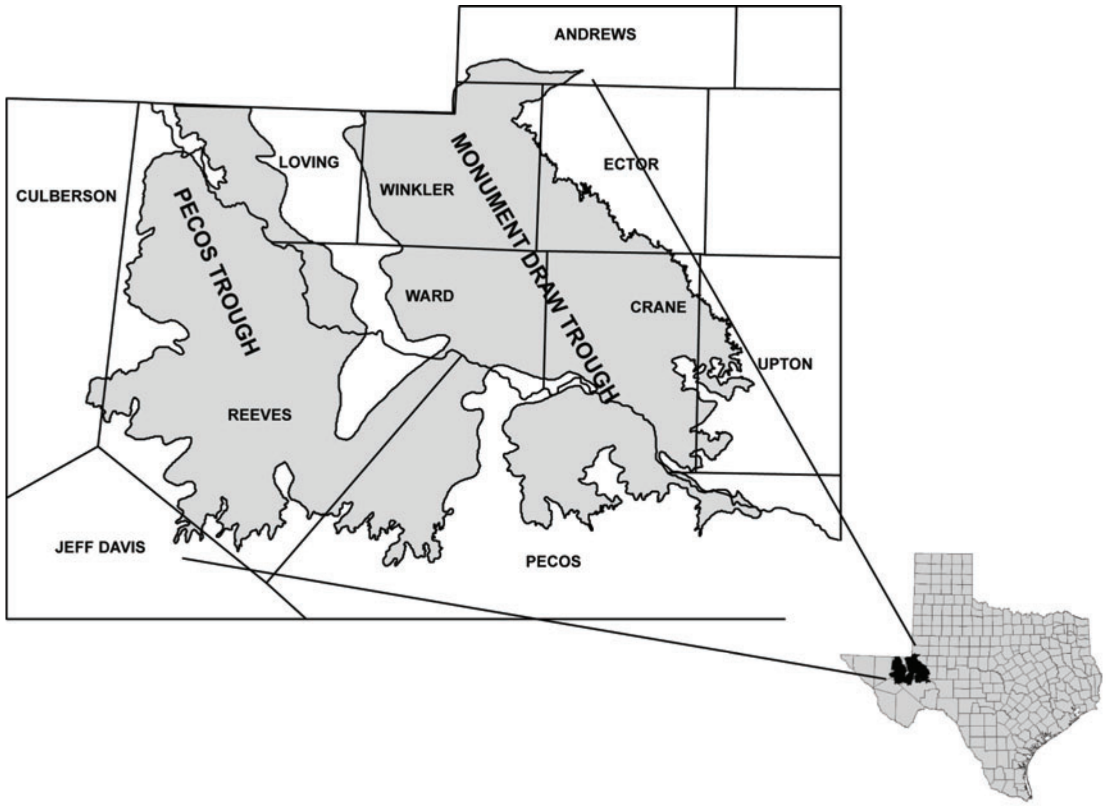


Figure 9-1: The Cenozoic Pecos Alluvium aquifer is located in the upper Pecos River valley of western Texas. This alluvium aquifer underlies parts of Crane, Ector, Loving, Pecos, Reeves, Upton, Ward, and Winkler Counties and extends north into New Mexico.

source of irrigation water, especially in Reeves and northwestern Pecos Counties (Ashworth and Hopkins, 1995; TWDB, 1997). Some groundwater from this aquifer is also exported to the City of Odessa by the Colorado River Municipal Water District (TWDB, 1997).

# Geology and Hydrogeology

The Cenozoic Pecos Alluvium aquifer is composed of Tertiary- and Quaternary-age alluvium up to 1,500 ft thick. This alluvium unconformably overlies older Permian, Triassic- and Cretaceous-age rocks (fig. 9-2; White, 1971). The alluvium is mostly composed of unconsolidated or poorly cemented clay, sand, gravel and caliche (White, 1971). North of the Pecos River, the alluvium is overlain in places by windblown sand deposited in dunes. This windblown sand was derived from the Pleistocene Blackwater Draw Formation, an older, extensive eolian deposit that crops out east of the region (White, 1971; Muhs and Holliday, 2001). The sand dunes are composed of fine quartz

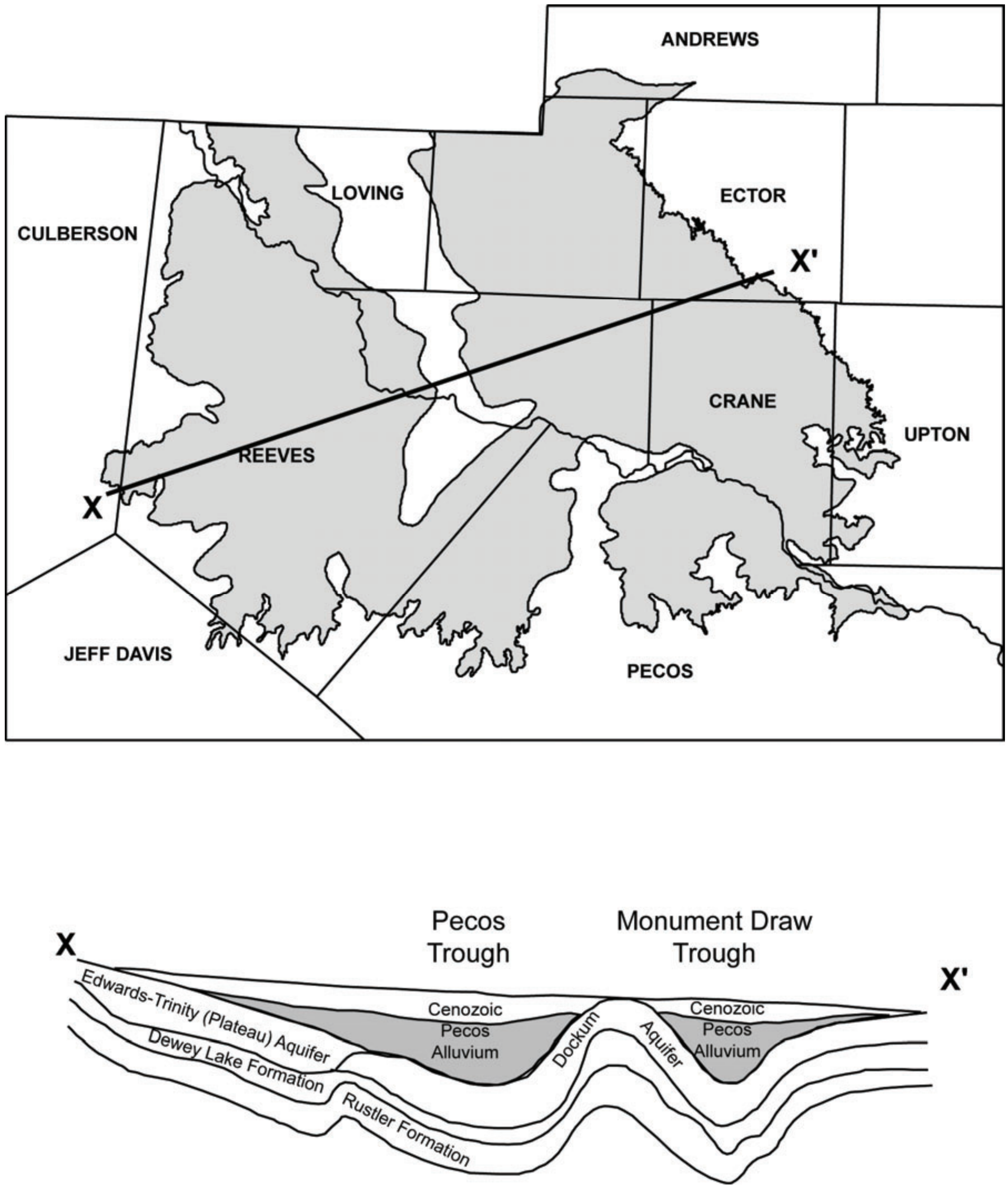


Figure 9-2: The Cenozoic Pecos Alluvium aquifer is composed of Tertiary and Quaternary age alluvium, up to 1,500 feet thick, that unconformably overlies older Permian, Triassic and Cretaceous age rocks. Modified from Ashworth and Hopkins (1995).

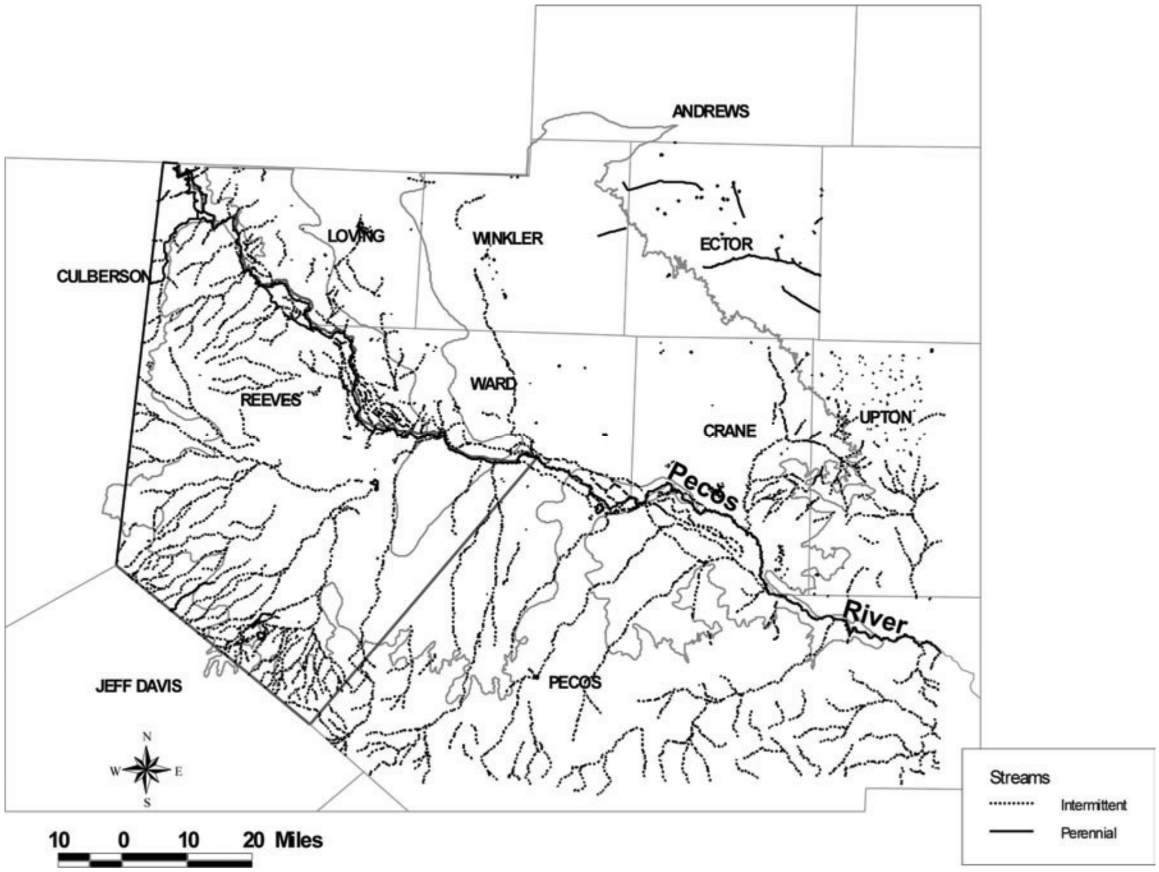


Figure 9-3: Excluding the Pecos River, there are few perennial streams flowing over the Cenozoic Pecos Alluvium aquifer. The high infiltration rates over the eastern part of the aquifer are responsible for the lack of either perennial or intermittent streams over that part of the aquifer.

sand, up to 250 ft thick (Garza and Wesselman, 1959; White, 1971). These dunes are potentially important sites for recharge (White, 1971). This is indicated by the fact that, excluding the Pecos River, there are few perennial streams north of the Pecos River because storm water quickly infiltrates into the dune sand (fig. 9-3; Garza and Wesselman, 1959; Ogilbee and others, 1962).

The Cenozoic Pecos Alluvium aquifer is unconfined, although clay beds may locally produce artesian conditions (Ashworth and Hopkins, 1995). This alluvium aquifer overlies, and in some places is hydrologically connected to, underlying aquifers. These aquifers include (1) the Edwards-Trinity (Plateau) aquifer in Pecos and Reeves Counties; (2) the Dockum Group in Ward and Winkler Counties; and (3) the Tertiary volcanics in Reeves County (Ashworth and Hopkins, 1995). Areas where groundwater is perched on clay beds that occur above the main water table have been identified near the City of

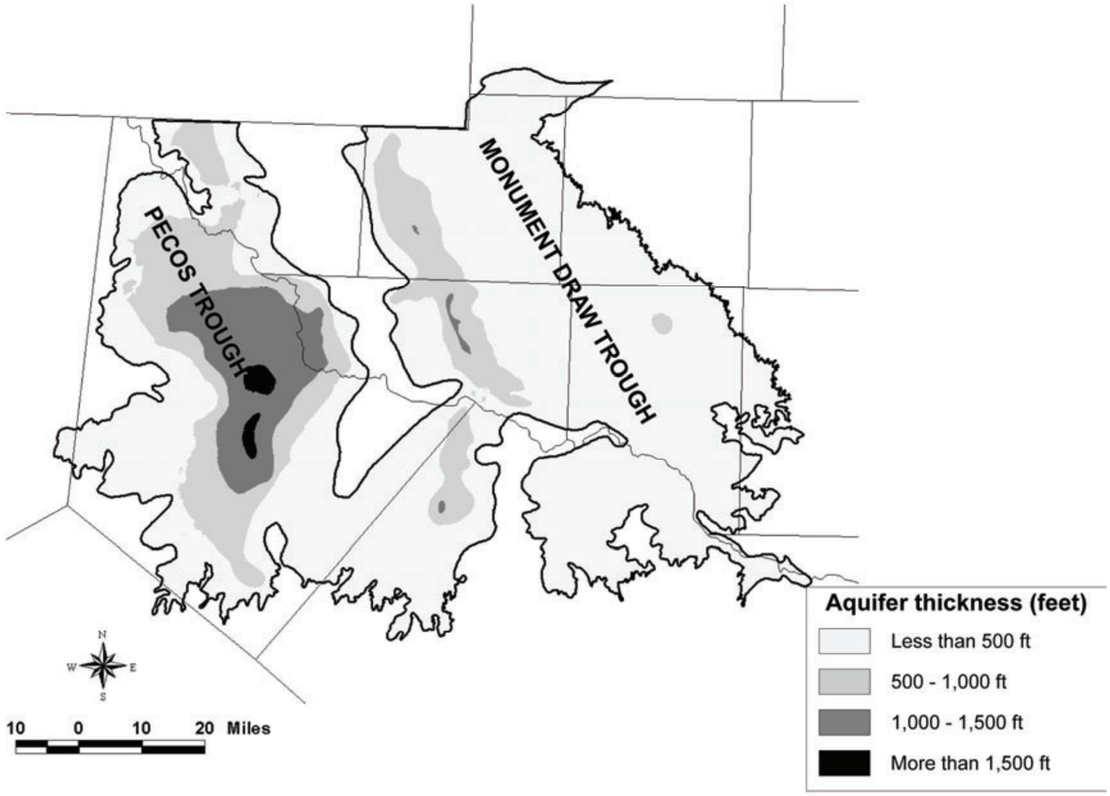


Figure 9-4: The Cenozoic Pecos Alluvium aquifer is composed of two main alluvium-filled troughs. These troughs were formed by subsidence that took place due to dissolution of underlying evaporites.

Pecos (Boghici, 1998; 1999). Well yields in the Cenozoic Pecos Alluvium aquifer are generally moderate to high (Ashworth and Hopkins, 1995). In the Pecos River Valley, depths to groundwater are 10 to 20 ft, increasing to about 50 ft away from the river (Boghici, 1998; 1999). Depths to groundwater are much greater in cones of depression adjacent to wells.

The Cenozoic Pecos Alluvium aquifer consists of two main basins or troughs: the Pecos Trough to west and the Monument Draw Trough in east (fig. 9-4). These are composed of alluvial sediments deposited in two major depressions during the Cenozoic Era (Ashworth, 1990). These troughs formed because of dissolution of underlying evaporites (rock salt, anhydrite, gypsum), especially but not exclusively in the Salado and Castile Formations (table 9-1). This dissolution resulted in the formation of the troughs due to subsidence of overlying rocks of the Rustler Formation, Dockum Group, and younger rocks (Ashworth, 1990).

Table 9-1: Stratigraphic units that comprise the aquifers of Loving, Pecos, Reeves, Ward and Winkler Counties.

| Era | System | Series/Group | | Stratigraphic Unit |
|---|---|---|---|---|
| Cenozoic | Quaternary | | | Cenozoic Pecos Alluvium |
| Mesozoic | Tertiary | | | Volcanic Rocks |
| | Cretaceous | Gulf | | undifferentiated |
| | | Comanche | Washita | undifferentiated |
| | | | Fredericksburg | |
| | | | Trinity | undifferentiated |
| | Triassic | Dockum | | undifferentiated |
| Paleozoic | Permian | Ochoan | | Dewey Lake Red Beds |
| | | | | Rustler Formation |
| | | | | Salado Formation |
| | | | | Castile Formation |
| | | Guadalupian | | Capitan Reef Complex |

Recharge to the Cenozoic Pecos Alluvium aquifer takes the form of infiltration of precipitation, seepage from ephemeral streams, and interaquifer flow, as well as irrigation return-flow (Ashworth, 1990). Most natural recharge is episodic, associated with heavy rainfall (Ashworth, 1990). Recharge is only likely to occur during long-duration rainfall events or periods of frequent smaller rainfall events; otherwise the water is lost to evapotranspiration (Ashworth, 1990). Recharge only occurs after soil moisture is high enough to overcome the effects of surface tension that would otherwise adhere the water to sand grains. High soil moisture allows water to infiltrate through to the water table (Ashworth, 1990). The most favorable sites for natural recharge of precipitation are the dune sands that overlie the Monument Draw Trough (fig. 9-5; Richey and others, 1985). These sand dunes are highly permeable and in some places sparsely vegetated (White, 1971). High permeability and sparse vegetation result in rapid infiltration of precipitation because together they minimize losses to evapotranspiration (White, 1971). It is possible that due to the occurrence of these highly permeable sand dunes, recharge rates may be

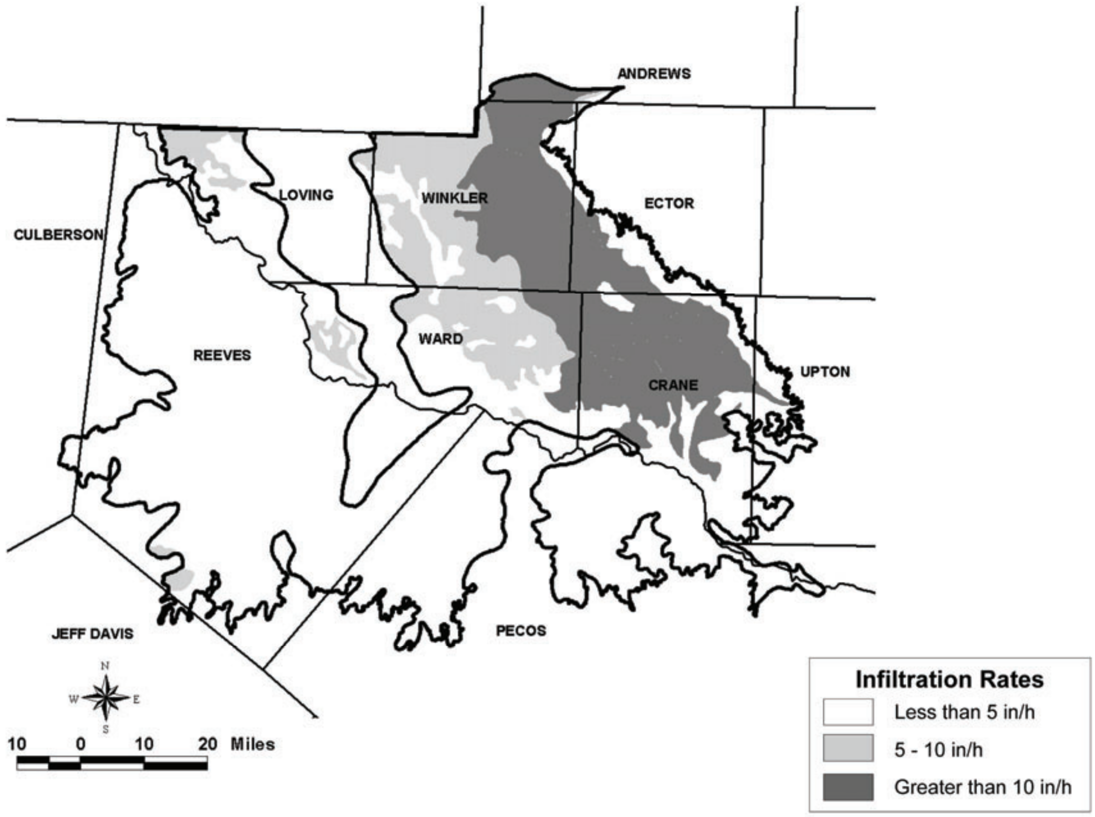


Figure 9-5: Average infiltration rates for soils overlying the Cenozoic Pecos Alluvium aquifer. The highest infiltration rates are associated with sand dunes that overlie the eastern part of the aquifer.

higher to the Monument Draw Trough than to the Pecos Trough. Recharge due to infiltration from ephemeral streams is also episodic, requiring sufficient precipitation to generate runoff through these streams.

Interaquifer flow primarily enters the Cenozoic Pecos Alluvium aquifer in the south and west, where the aquifer is hydrologically connected to Permian (Rustler Formation), Cretaceous (Edwards-Trinity aquifer), and Tertiary volcanics aquifers (Ashworth, 1990). Seepage from irrigation canals and irrigation return-flow also contributes water to the aquifer. Estimates of losses due to seepage from irrigation canals range from 30 to 72 percent (Ashworth, 1990). These high loss rates can be attributed to the high-permeability sandy soils that overlie parts of the aquifer. Overall, irrigation return-flow is estimated to be 20 percent of applied irrigation water (Ashworth, 1990).

Natural discharge from the Cenozoic Pecos Alluvium aquifer takes the form of evapotranspiration adjacent to the Pecos River and discharge into the Pecos River (White, 1971). Evapotranspiration losses are greatest in lowlands adjacent to the river and other

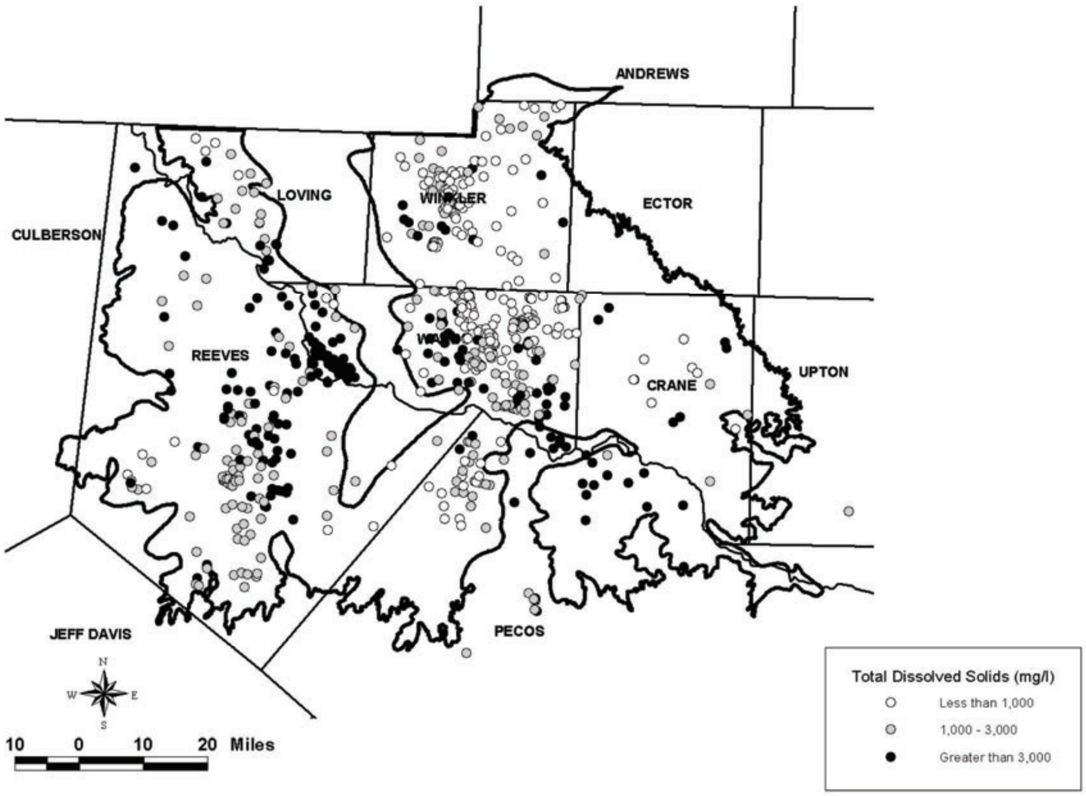


Figure 9-6: Total dissolved solids in Cenozoic Pecos Alluvium aquifer groundwater. Groundwater salinity tends to be greatest west of, and along the Pecos River. The lowest groundwater salinities are associated with the sand dunes that occur in the eastern part of the aquifer.

areas where the water table is close to land surface. These losses primarily take the form of uptake by vegetation (e.g., saltcedar and mesquite) that are abundant in these areas (White, 1971). Water uptake by vegetation can be substantial. For example, estimated transpiration rates for saltcedar, juniper, mesquite, cattail, and shrubs are 2 to 20, about 2, 1 to 2, 4 to 10, and 1 to 2 acre-ft/acre/yr, respectively (Gatewood and others, 1950; McDonald and Hughes, 1968; Van Hylckama, 1970; Weeks and others, 1987; Devitt and others, 1997; Ansley and others, 1998; Dugas and others, 1998).

# Water Quality

Groundwater quality in the Cenozoic Pecos Alluvium aquifer is variable. Dissolved solids in Cenozoic Pecos Alluvium groundwater range from 300 mg/L to more than 5,000 mg/L (Ashworth and Hopkins, 1995). Groundwater quality is generally better in the

Monument Draw Trough portion of the aquifer than in the Pecos Trough (fig. 9-6). Groundwater in the Pecos Trough is generally slightly to moderately saline, while groundwater in Monument Draw Trough varies from fresh to moderately saline. In the Monument Draw Trough, more saline groundwater tends to occur on the western side of the trough adjacent to the Pecos River in parts of Winkler, Ward, and Pecos Counties. The lowest dissolved solids (< 500 mg/L) in the aquifer are generally associated with dune sands. Groundwater quality generally deteriorates with depth, although the most saline groundwater in the Cenozoic Pecos Alluvium aquifer actually occurs in shallow wells (fig. 9-7).

Saline groundwater that occurs in this aquifer is mostly the result of natural processes. However, poor water quality may result from anthropogenic activity in some areas (Ashworth and Hopkins, 1995). Groundwater quality in the Cenozoic Pecos Alluvium is influenced by several factors: (1) the presence of evaporite beds, (2) evaporation, (3) recharge of irrigation return-flow, (4) pumpage, and (5) past oil-field practices. The presence of evaporite beds in the Rustler Formation, especially underlying northern and western parts of Pecos Trough, produces elevated sulfate in the groundwater owing to interaquifer flow (Ashworth, 1990). Shallow saline groundwater occurs in the Pecos

River Valley. This salinity can be attributed partially to evaporation in areas where the water table is shallow. Saline groundwater may also result from activities related to agriculture. Salinity may result from recharge of irrigation return-flow, especially in Reeves County, or encroachment of saline groundwater related to heavy pumpage (Ashworth, 1990). Irrigation return-flow may become saline because of evaporation at land surface or dissolution of salts accumulated in the soil. In some areas, nitrate derived from fertilizers may impact groundwater quality (Ashworth, 1990). In these areas, fertilizer nitrogen is leached from the soil by infiltrating precipitation or irrigation return-flow. Groundwater salinity may increase because heavy pumpage draws in more saline groundwater that occurs at depth. Locally, saline groundwater occurs because of oil-field brine, especially in Winkler and Loving Counties (Ashworth, 1990). Most of this contamination is related to past disposal of large quantities of brine in unlined pits or improperly cased oil wells (Ashworth, 1990).

# Water levels

Groundwater in the Cenozoic Pecos Alluvium aquifer generally flows toward the Pecos River, except where pumpage forms cones of depression (fig. 9-8; Boghici, 1998; 1999). This situation suggests that there is probably no groundwater flow between the two main troughs. Therefore, it can be concluded that they are separate groundwater flow systems.

This aquifer has experienced historic water-level declines of more than 200 ft in parts of south-central Reeves and northwest Pecos Counties. One of the results of this water-level decline has been reduced base flow to the Pecos River. The water-level variations over time in the aquifer have been associated with varying intensity of irrigation pumpage (Ashworth and Hopkins, 1995). Irrigation farming developed in Reeves and Pecos

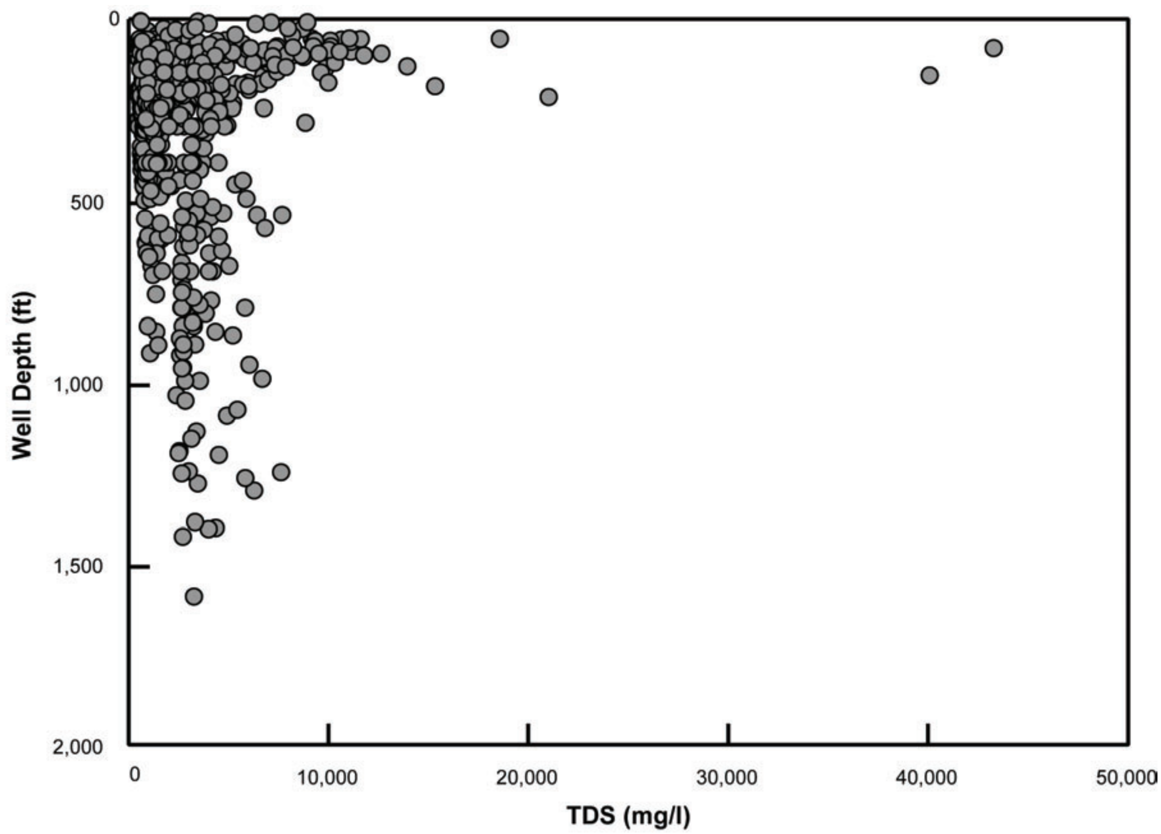


Figure 9-7: Total dissolved solids (TDS) in Cenozoic Pecos Alluvium aquifer groundwater generally increase with depth. In some areas, saline occurs in shallow groundwater due to the effects of evaporation or influxes of saline irrigation return-flow.

Counties in the 1940's peaked in the 1950's and began declining in the mid-1970's (TWDB, 1997). In Reeves County, the number of irrigation wells increased tenfold, from 35 to 355, between 1940 and 1950 (Hood and Knowles, 1952).

Water levels in the aquifer have responded to changes in irrigation pumpage rates (fig. 9-9). Water levels dropped sharply in the late 1940's and early 1950's in response to the development of irrigation farming and leveled off in the 1960's. Water levels began to recover in the mid-1970's owing to decreased irrigation pumpage (Ashworth and Hopkins, 1995). In the main irrigated areas, water levels also exhibit seasonal fluctuations related to seasonal irrigation cycles (Ashworth, 1990). Groundwater levels drop during summer months when irrigation demand is greatest and recover slightly during the winter when little or no irrigation is taking place. Water-level declines have been greatest in the major irrigation areas of Reeves and northern Pecos Counties (fig. 9-8 and 10). Two major cones of depression have formed in irrigated areas along State Highway 17 in Reeves County and the Coyanosa irrigation area of Reeves and Pecos Counties (fig. 9-8; Boghici, 1998; 1999). Irrigation pumpage has been less intense in the

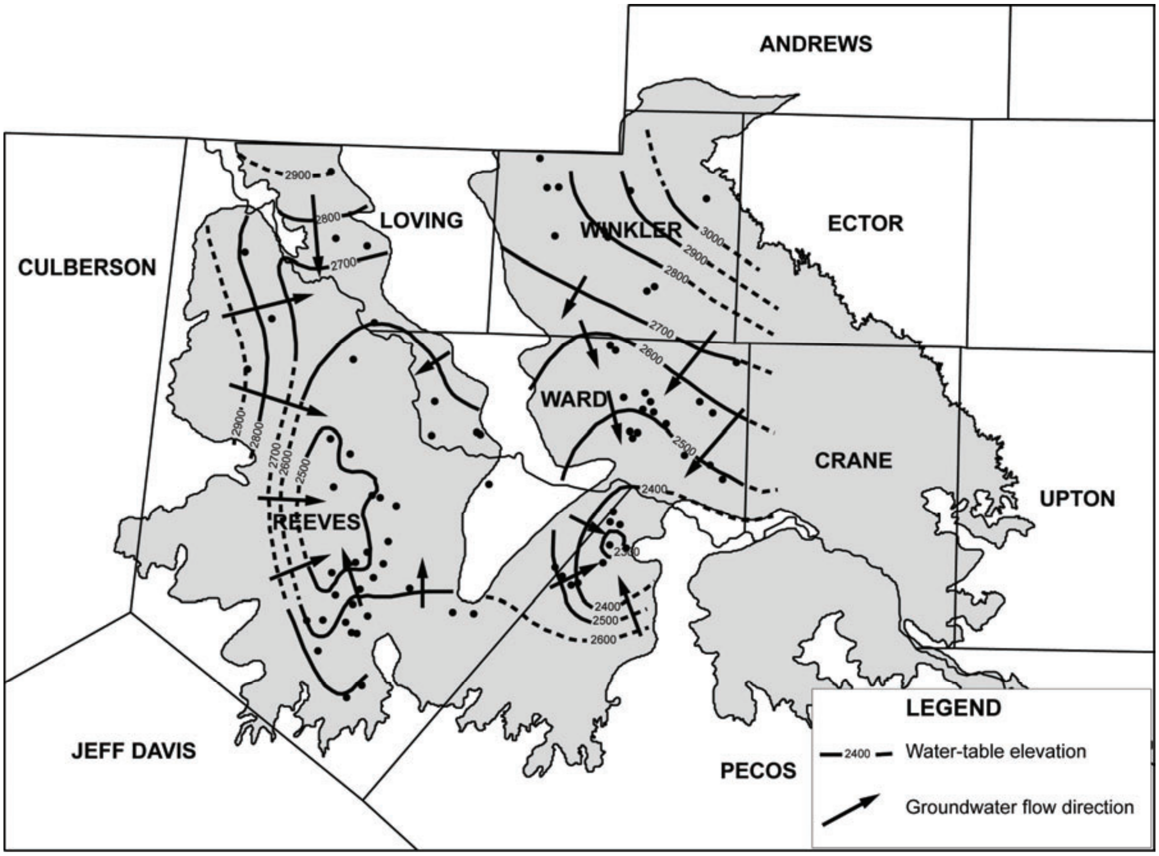


Figure 9-8: Water-level elevations in the Cenozoic Pecos Alluvium aquifer, 1998 (modified from Boghici, 1999).

Monument Draw Trough than the Pecos Trough. Therefore, water-level declines have been less of a problem north of the Pecos River (Ashworth, 1990).

In the 1990's, groundwater levels rose in parts of Reeves County that had previously been heavily impacted by irrigation pumpage. However, water-level declines have been

observed in other parts of the aquifer (fig. 9-11; Boghici, 1999). Rising water levels have been observed along State Highway 17, the main irrigation area in Reeves County, while water-level declines have been observed in the Coyanosa area of Reeves and Pecos Counties and in eastern Ward County south of Monahans (Boghici, 1998; 1999). Unlike the water-level declines in Reeves and Pecos Counties that are attributable to continued irrigation, the water-level declines in Ward County are associated with pumpage related to public supply and industrial uses (Boghici, 1998, 1999).

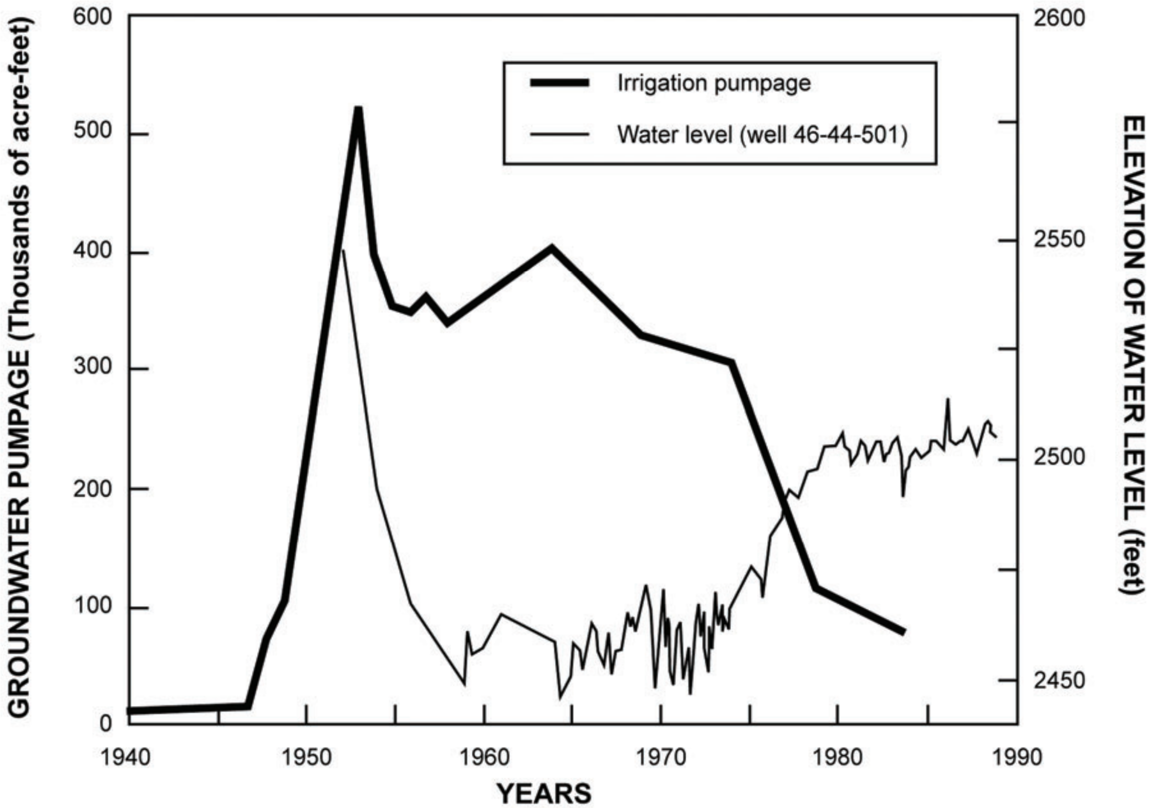


Figure 9-9: Irrigation pumpage from the Cenozoic Pecos Alluvium aquifer in Reeves County and associated groundwater-level responses (modified from Ashworth, 1990)

# Summary

The Cenozoic Pecos Alluvium aquifer is an unconfined alluvial aquifer. This aquifer is naturally recharged by infiltration of precipitation, seepage from ephemeral streams, and interaquifer flow from underlying aquifers. Discharge from the aquifer primarily takes the form of evapotranspiration where the water table is shallow, base flow to the Pecos River, and by pumpage primarily related to irrigation.

Cenozoic Pecos Alluvium groundwater is characterized by dissolved solids concentrations that are generally less than 5,000 mg/L. Groundwater salinity is generally lower east of the Pecos River than to the west. Groundwater salinity is mainly related to natural or pumpage-related inflows of saline groundwater, evaporation from the aquifer, saline irrigation return-flow, and local oil-field brine contamination. The lowest salinity is associated with sand dunes and may thus be recharge related. Recharge of precipitation, characterized by low dissolved solids, will potentially reduce groundwater dissolved solids by dilution.

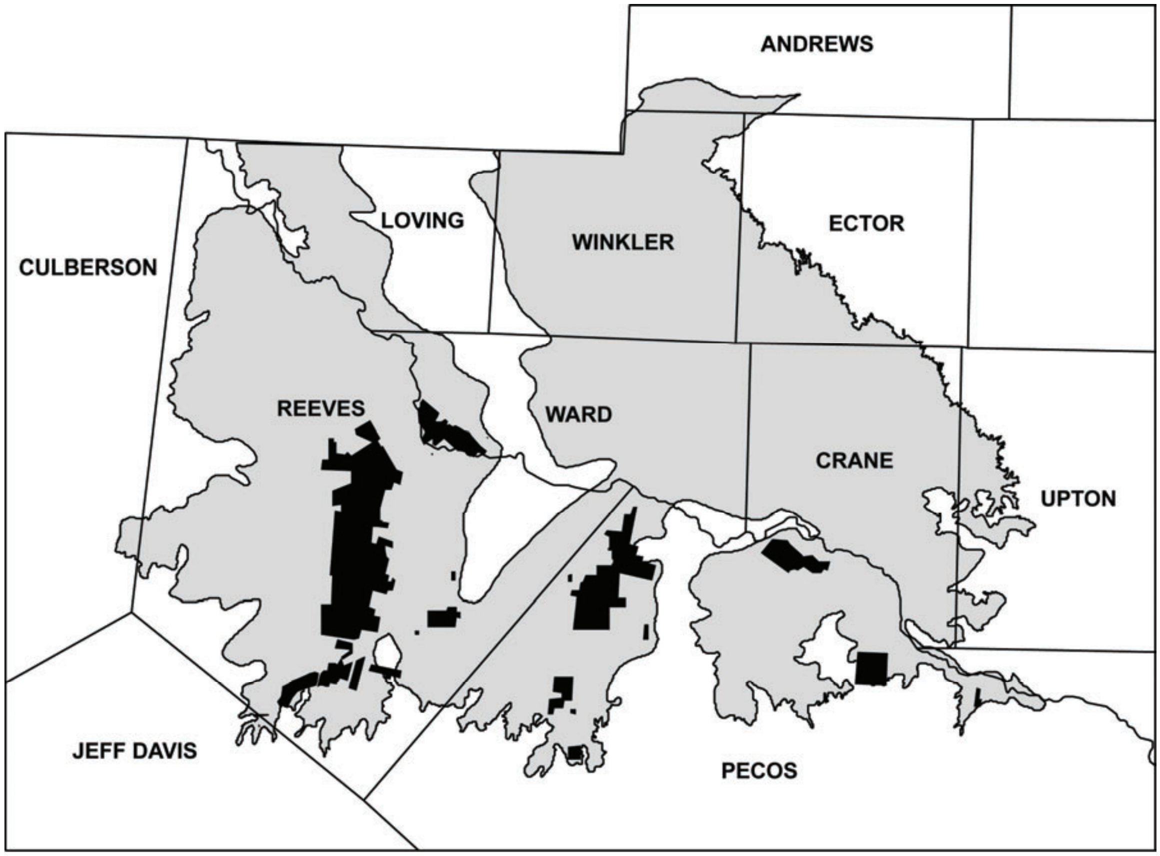


Figure 9-10: Irrigated farmland overlying the Cenozoic Pecos Alluvium aquifer. Based on 1994 survey of irrigation in Texas.

The Cenozoic Pecos Alluvium aquifer is divided into two parts: the Pecos Trough and the Monument Draw Trough. These troughs form separate groundwater flow systems. The Monument Draw Trough displays the potential for higher recharge rates than the Pecos Trough because of the presence of permeable dune sands. The better groundwater quality in the Monument Trough can be attributable to many factors, such as higher recharge of precipitation, less irrigation, and less inflow of saline groundwater from underlying rock units.

Starting in the 1940's, irrigation pumpage resulted in water-level declines of up to 200 ft in parts of the Cenozoic Pecos Alluvium aquifer. This water-level decline has primarily taken place in the major agricultural areas of Reeves and Pecos Counties. Decreased irrigation pumpage starting in the mid-1970's has resulted in water-level recovery in some parts of the aquifer. A recent survey indicates rising water levels in some parts of

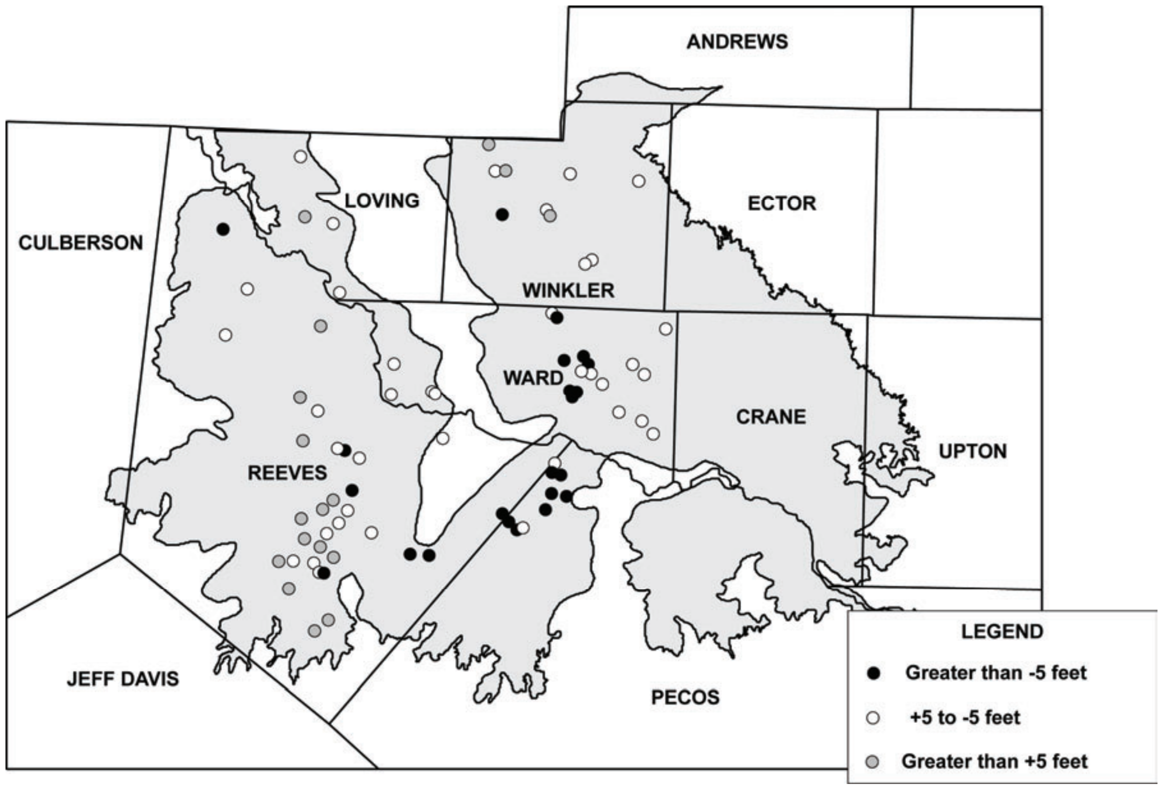


Figure 9-11: Cenozoic Pecos Alluvium aquifer groundwater-level changes between 1989 and 1998 (modified from Boghici, 1999).

Reeves County that had previously been heavily impacted by irrigation pumpage. However, water levels continue to decline in other areas, especially in Pecos and Ward Counties because continued irrigation pumpage, as well as public supply and industrial pumpage.

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# Chapter 10

# Bone Spring-Victorio Peak Aquifer of the Dell Valley Region of Texas

John B. Ashworth[1]

## Introduction

The Bone Spring-Victorio Peak aquifer produces groundwater in an irrigated region commonly referred to as Dell Valley. Because of its importance to the local agricultural economy, the Texas Water Development Board (TWDB) has designated the Bone Spring-Victorio Peak as a minor aquifer and has delineated its extent in Texas on the basis of its occurrence underlying irrigable land. This paper, which is a modification and update of TWDB Report 344 by the same author, describes the groundwater resource underlying the valley in terms of its geological and hydrological characteristics, quantity, quality, historical use, and changing conditions.

## Location

Dell Valley is located 75 mi east of El Paso and 20 mi west of the Guadalupe Mountains in northeastern Hudspeth County (fig. 10-1). Dell City, with a population of approximately 500, is located in the center of the irrigation district. The valley consists of approximately 40,000 acres of irrigable land in Texas and extends northward into Otero County, New Mexico, where it is referred to as Crow Flats. Low rainfall, averaging 8 to 10 inches annually, and a high rate of evaporation, which averages nine times the precipitation rate, characterize the arid climate in the region.

Dell Valley is a broad, alluvial, outwash plain that is bordered on the east by the Salt Basin and gently rises to the west and south to limestone uplands of the Diablo Plateau. The land-surface elevation of the valley rises gradually from approximately 3,640 ft above sea level on the eastern edge to approximately 4,200 ft on the western edge.

Although infrequent, a major problem in the watershed is its susceptibility to flooding. Surface drainage originates in the Cornudas Mountains, Sixteen Mountains, and the Sierra Tinaja Pinta in the far western extent of the watershed. Floodwaters intermittently traverse from the highlands onto the Dell Valley alluvial plain through Eightmile, Hitson, C&L, and Washburn draws in Texas, and Cornudas and North draws in New Mexico (fig. 10-2). Total watershed area is approximately 600 mi$^2$. Runoff from a storm in 1966 resulted in the largest flood in Dell Valley's recorded history and caused approximately

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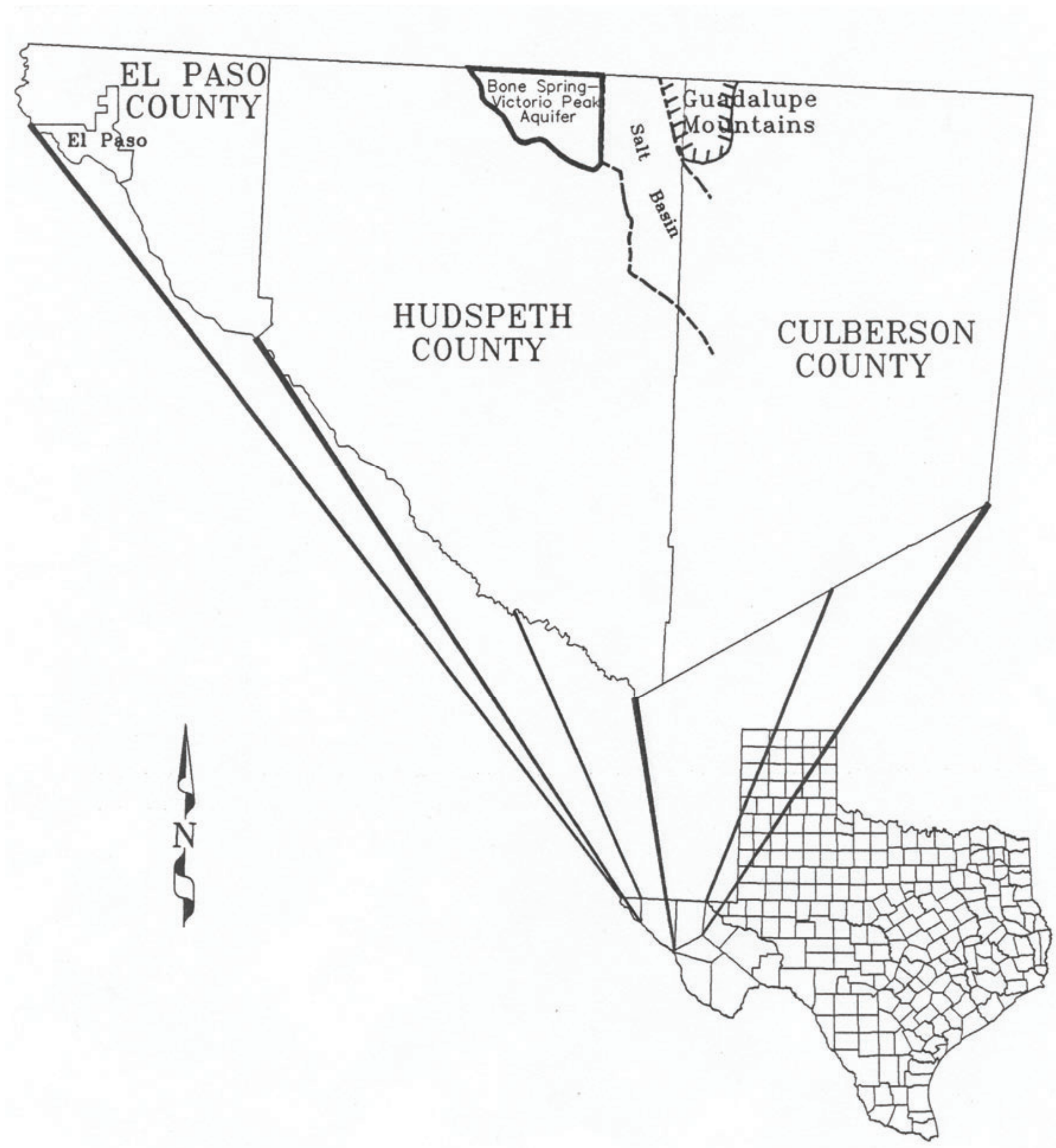


Figure 10-1: Location of the Bone Spring-Victorio Peak aquifer.

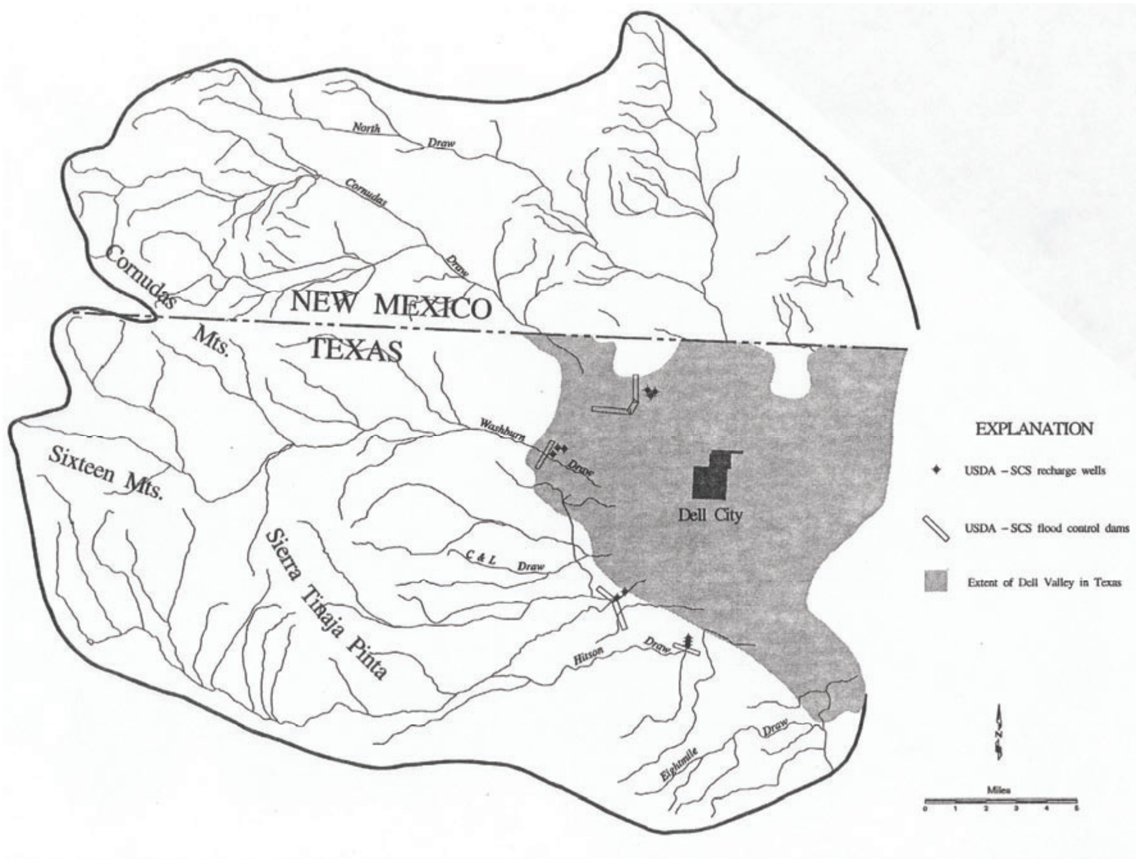


Figure 10-2: Watersheds contributing to Dell Valley.

$3 million in damages (El Paso-Hudspeth Soil and Water Conservation District and others, 1969).

Floodwaters reaching the valley floor typically fan out in an overland or sheet-type flow causing extensive damage. However, the U.S. Department of Agriculture, Soil Conservation Service, has constructed four flood-control structures on the west side of the valley (fig. 10-2) to capture floodwaters draining through Cornudas, Hitson, C&L, and Washburn draws.

# History of Water Use

## Springs

The first mention of water supplies in the area is recorded in scientific journals and military travel logs from the middle 1800's (e.g., Marcy, 1851; Pope, 1954). Travelers involved in exploration and survey trips, wagon trains, and military expeditions frequently stopped at Crow Springs (also known as Ojos del Cuervo) to replenish their water supplies (Brune, 1981). The springs were also a stage stop on an early Butterfield Overland mail route. Located northeast of Dell City and near the State line, Crow Springs issued brackish water that had a strong sulfur odor and filled two shallow lakes that covered 4 or 5 acres. Shallow wells dug in the vicinity of the springs provided more potable water. As late as 1948, the springs still trickled approximately 3 gallons per minute (gpm). However, by 1950, pumping of irrigation wells drilled near the springs lowered the water table and brought an end to the discharge.

## Irrigation

Prior to the introduction of the first irrigation wells in 1947, Dell Valley was primarily the site of cattle ranching, and the only use of groundwater was for domestic and livestock needs (Scalapino, 1950). By 1949, 78 wells had been drilled; however, only 32 wells had sufficient yields to be used for irrigation purposes. About 2,500 acres were irrigated in 1948. A year later, about 6,000 acres of feed crops and cotton were irrigated with approximately 18,000 acre-ft of groundwater.

Across the State line in the Crow Flats area of New Mexico, a few wells were completed with windmills for domestic and livestock use as early as 1905 and 1906 (Bjorklund, 1957). The first irrigation wells were drilled in 1949 shortly following their introduction in Dell Valley. In 1956, 17 out of 23 wells drilled to supply water for irrigation in Crow Flats were in use to irrigate approximately 3,000 acres of cotton and alfalfa. In the combined Dell Valley–Crow Flats region, 228 irrigation wells were in use in 1956 to irrigate approximately 32,000 acres.

Irrigated agriculture in the valley continued to expand through the late 1970's as approximately 39,000 acres of cropland were irrigated with 144,000 acre-ft of groundwater (TWDB, 1996). Irrigation diminished through the 1980's as a result of declining market conditions, increased labor expense, and government conservation programs. Irrigation in the 1990's once again increased. A 1994 survey indicated approximately 28,000 acres was irrigated with 165,000 acre-ft of groundwater. Figure 10-3 shows the amount of water pumped for irrigation use and the corresponding number of acres irrigated for specified time periods.

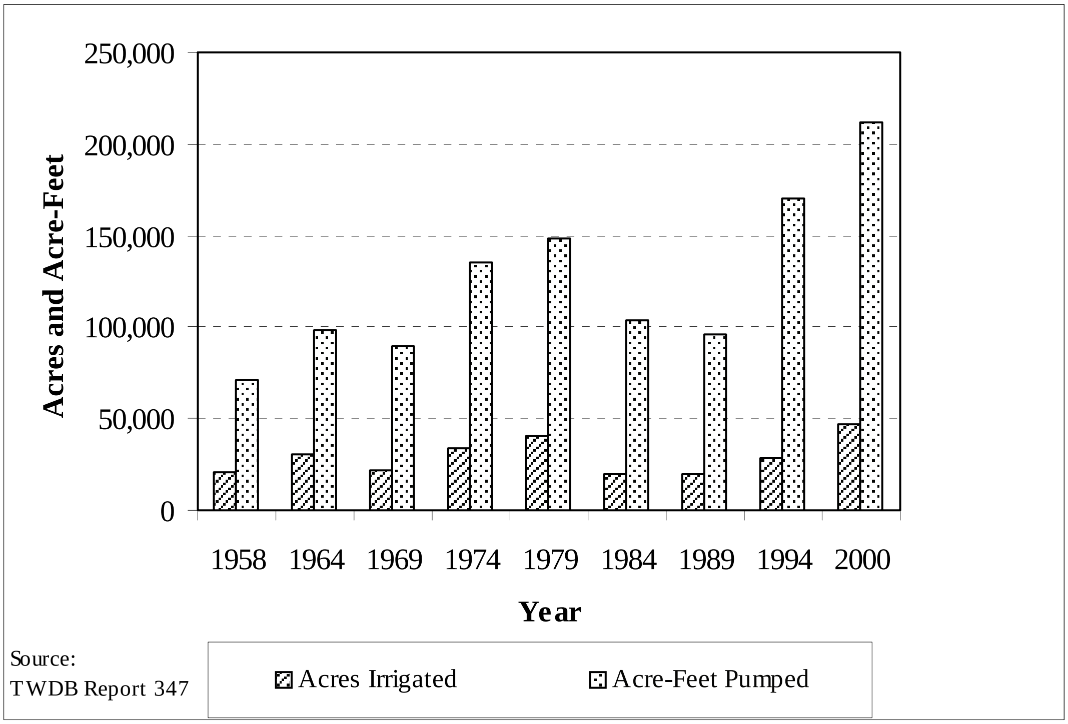


Figure 10-3: Irrigation pumpage and acres irrigated.

## Public Supply

With a population of more than 500, Dell City became incorporated in 1961 and began plans for a public water-supply system (Young, 1975, 1976). Domestic water supply had previously been provided by several private water companies; however, by 1964, it was evident that the area groundwater supply was becoming increasingly saline and a water treatment system would be necessary.

In 1967, the city installed an electrodialysis treatment plant with a capacity of 50,000 gallons per day (gpd). The plant was designed to mitigate as much as 2,450 parts per million (ppm) dissolved solids to potable standards. Water was supplied from a single well (North Well, 48-07-522). Dell City was the first community in the United States to

incorporate saline water conversion equipment in a system financed by the Farmers Home Administration, U.S. Department of Agriculture. The plant was increased to 69,000 gpd in 1968 in order to treat enough water to satisfy peak demands. Also, an additional well (Elias or South Well, 48-07-523) was brought into the system.

By 1974, the plant had become ineffective. Chemical quality of the source water from the aquifer had deteriorated beyond the design specifications for the plant. In addition, the plant system had not been adequately maintained. The old plant was replaced in 1976 with a modern, reverse-polarity-type electrodialysis plant with a 100,000-gpd capacity. The plant is currently in use and operates at a rate of 50,000 to 70,000 gpd.

In 1986, the Prather Well (48-07-219), located 3 mi north of town, was drilled to provide the primary source of water to the plant. The North Well is still connected to the system as a backup but is rarely used for that purpose.

The city also operates a separate water delivery system for irrigation use. Water for this purpose is pumped from the North Well and is supplemented with the by-product water from the electrodialysis treatment plant. The plant by-product water is actually of better quality than the water from the North Well. The Elias Well is used occasionally as a backup or supplement to the irrigation system.

# Geology

The principal water-bearing rocks that underlie Dell Valley are limestones and dolomites of Permian age. These rocks of marine origin were deposited in the early development of the Delaware Basin. The Victorio Peak Limestone occupies much of the surface area immediately west of the Salt Basin on the Diablo Plateau and, along with the underlying Bone Spring Limestone, is prominently exposed on the eastern escarpment of the Sierra Diablo south of Dell Valley (King, 1965). The Bone Spring Limestone is predominantly a black to dark-gray, cherty limestone with thin interbedded black or brown layers of siliceous shale. The Bone Spring grades upward into the Victorio Peak Limestone, a light-gray, thick-bedded, mainly calcitic but slightly dolomitic limestone.

Two significant faults are of particular interest in the Dell Valley area. A north-south-trending fault is the west boundary of the Salt Basin and represents the approximate eastern extent of the aquifer. Displacement along the fault has not been precisely determined; however, sediments on the eastern side of the fault have probably dropped several hundred feet (King, 1948; Gates and others, 1980; Goetz, 1977). A second fault, trending northwest-southeast, forms the southern topographic edge of the valley and is also the designated aquifer boundary. Downward displacement of approximately 100 ft occurs on the north side of the fault.

An igneous (volcanic) intrusive body of Tertiary age, known locally as Round Mountain, crops out 3 mi east of Dell City and rises about 175 ft above the valley floor (fig. 10-4). Other prominent igneous peaks comprise the Sierra Pinta and Cornudas Mountains approximately 10 to 15 mi west of Dell City.

Overlying much of the Permian limestone formations in the delineated aquifer area is a mantle of up to 150 ft of Quaternary and recent alluvial sediments ranging in size from boulders to silt and clay. The sediments were eroded from highlands to the west and northwest, transported by flooded streams, and deposited on the relatively flat valley floor. Surface soils overlying the alluvium are largely gray silts and silt loams, underlain at depths of 1 to 3 ft by a soft marl or caliche. The high natural salinity of the soil suggests that at one time, the salt lake that currently exists in the Salt Basin to the east

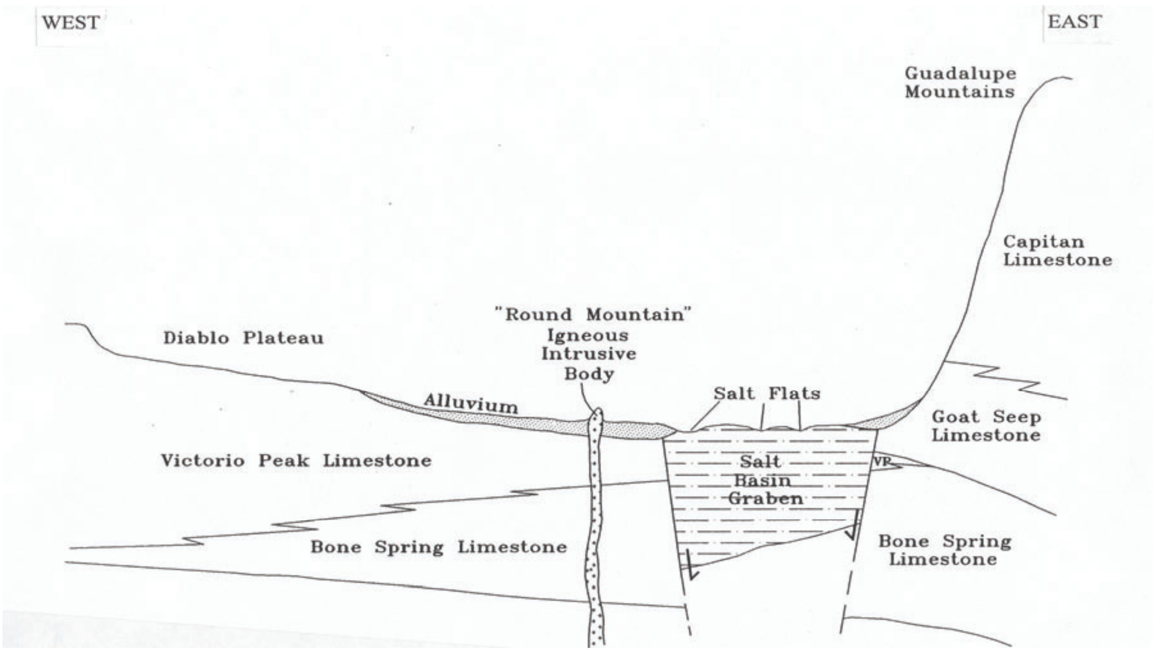


Figure 10-4: Generalized geologic section across Dell Valley and the Salt Basin.

may have covered the entire valley. Years of irrigation water application have actually improved the chemical condition of the soil by lowering the pH and total salt and sodium content (Longenecker and Lyerly, 1959). Unfortunately, these minerals have not been eliminated but, instead, have been transported downward to the underlying aquifer.

# Hydrology

## Occurrence

Groundwater occurs in the Permian limestones throughout the Diablo Plateau region. However, unlike elsewhere on the plateau, the aquifer in the Dell Valley area has been developed because of the relatively shallow water table and the presence of soils capable of growing crops. Groundwater in the aquifer is concentrated in interconnected solution cavities that have developed in joints, fractures, and bedding planes that vary in size and dimension. Water-bearing zones have been encountered in wells drilled in excess of 2,000 ft. Well production is thus linked to the number and size of cavities intercepted by the well bore.

## Recharge

Recharge to the regional Diablo Plateau aquifer system is derived from the infiltration of precipitation on the entire plateau area of approximately 2,900 $mi^2$ and the downward seepage of water in the Sacramento River. Recharge on the Diablo Plateau primarily occurs as infiltration of runoff in beds of ephemeral streams, or arroyos, during occasional flash floods. Only during intense rainstorms is the rate of precipitation greater than evaporation. Much of the groundwater in Dell Valley originates as precipitation that infiltrates into the regional aquifer system within the drainage area (fig. 10-2). Karst features, such as vertical fractures and sinkholes, permit rapid access of infiltrating surface water. The presence of tritium in most well samples collected from the plateau aquifer indicates recent recharge (Kreitler and others, 1987).

The Sacramento River, which drains the Sacramento Mountains in New Mexico, is a major source of recharge in the northern segment of the plateau (Scalapino, 1950; Mayer, 1995). Water drains rapidly into the subsurface as the river leaves higher elevations and encounters the flatter surface of the plateau. Mayer (1995) showed that groundwater in the northern part of the plateau is chemically similar to the water in the river but differs from groundwater elsewhere in the plateau. Mayer speculates that the river source may influence the quality of the aquifer in the northern and eastern parts of Dell Valley, where fresher conditions occur.

A change in the chemical quality of groundwater in the valley over time is a possible indication that some water pumped for irrigation use has returned to the aquifer. Logan (1984) suggested that 35 percent of groundwater pumped returns to the aquifer. Davis and Gordon (1970) estimated a return-flow of as much as 50 percent.

A continuous water-level record in well 48-07-516 and annual irrigation pumpage in the valley for specified years are compared in figure 10-5. Since 1984, irrigation pumpage has varied from approximately 40,000 to 100,000 acre-ft annually. At the lower range of annual pumpage (40,000 to 60,000), water levels have risen, while at a higher range of pumpage (90,000 to 100,000), water levels have remained relatively constant. Therefore, 90,000 to 100,000 acre-ft appears to be a reasonable estimate of total annual recharge to the aquifer, which includes both lateral inflow and irrigation return-flow.

Construction of four flood-control structures on the western side of Dell Valley is capable of providing as much as 3,300 acre-ft of recharge annually by seepage through the highly permeable pool area (El Paso-Hudspeth Soil and Water Conservation District and others, 1969). However, there has not been enough significant rainfall to fill the reservoirs since the completion of the dams. Included in the project are 11 wells for recharging water captured by the dams (fig. 10-2). Each well is designed with the intention of recharging water by gravitational flow at a rate of at least 2,000 gpm (Logan, 1984).

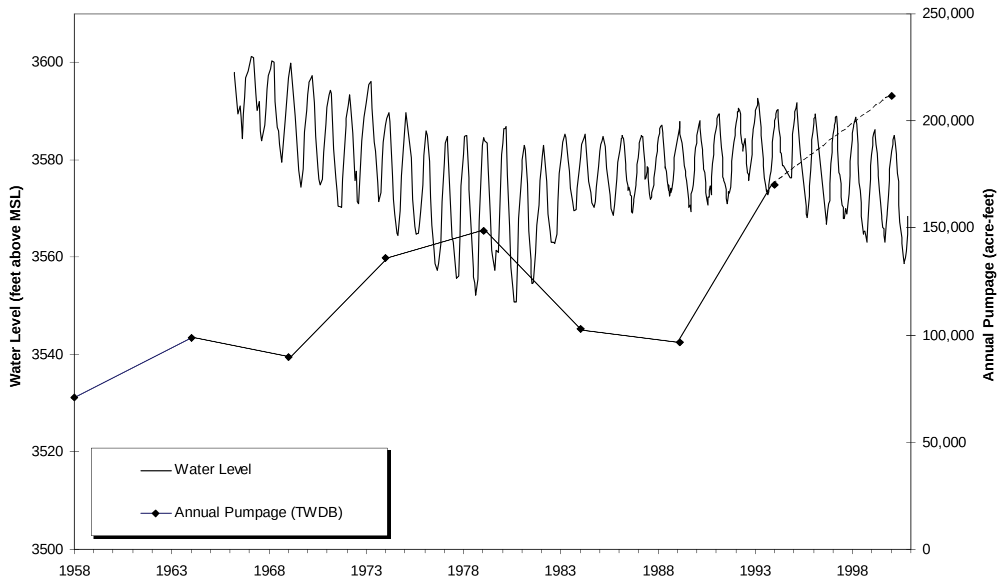


Figure 10-5: Hydrograph of water level in well 48-07-516 and annual pumpage during designated years.

## Movement

Regionally, groundwater moves in an east-to-northeasterly direction from the Diablo Plateau in Texas toward the Salt Basin, where it discharges naturally by evaporation from the salt flats. Across the State line in New Mexico, groundwater flow moves in a southeasterly direction toward the basin (Mayer, 1995). A regional potentiometric surface map prepared by Kreitler and others (1987) illustrates a relatively low hydraulic gradient of 2.5 to 5 ft/mi. Within the Salt Basin, groundwater percolates upward to the surface, drawn by evaporation through the capillary fringe in the flats (Boyd and Kreitler, 1986).

The orientation and concentration of solution cavities developed along prominent fractures and bedding planes control water movement on a local scale. During the irrigation season, movement is altered in the direction of pumping wells.

Declining water levels caused by pumpage may reverse the groundwater flow direction on the eastern side of the valley and allow highly saline water to move westward into the irrigated region. Current water-level elevations in the central part of Dell Valley are, in fact, lower than levels in the adjacent Salt Basin, which suggests that the potential for such movement does exist. However, chemical-quality analyses of water samples from wells located along the eastern side of the valley do not indicate a significant influx of saline water. The less-permeable sediments that fill the Salt Basin may hinder rapid migration of the saline water.

## Discharge

Large quantities of water are discharged from the aquifer annually. Discharge occurs naturally through springs, seeps, and evaporation from the salt flats and artificially by pumpage.

### *Natural Discharge*

Eastward migrating groundwater underlying the Diablo Plateau moves into the Salt Basin, where it partially discharges by evaporation, especially from the salt flats where the water table is 3 to 10 ft below the surface (Boyd and Kreitler, 1986). Prior to irrigation development, the aquifer was at a quasi-steady state, and the amount of water discharged through evaporation from the salt flats was approximately equal to the recharge to the Bone Spring-Victorio Peak aquifer in the Dell Valley and Crow Flats areas.

Bjorklund (1957) estimated that less than 100,000 acre-ft annually were originally discharged by way of evaporation.  Davis and Leggat (1965) estimated that approximately 40,000 acre-ft evaporate annually in the Texas portion of the Salt Basin. Boyd and Kreitler (1986) suggested that evaporation rates on the salt flats could theoretically range from 15.7 to 78.7 inches/yr or more. At this rate 49,000 to 243,000 acre-ft of groundwater could evaporate annually.

### *Pumpage*

With the advent of irrigated agriculture in the 1940's, pumpage has become the principal means of discharge from the aquifer. Except for a scattering of wells throughout the Diablo Plateau, almost all of the pumpage occurs in the Dell Valley and Crow Flats areas. Pumpage in the Dell Valley area reached a peak in the late 1970,s, with more than 140,000 acre-ft being pumped annually. Annual withdrawals for irrigation use since 1984 range from approximately 40,000 to 100,000 acre-ft. During the 1970's, pumpage exceeded recharge, resulting in a decline in the elevation of the water table. The historical development of groundwater use in this area is discussed more thoroughly in the section titled "History of Water Use."

Regionally, the aquifer is highly transmissive. However, at any particular location, well yields can vary significantly. Highly productive wells, producing up to 3,000 gpm, are those that intersect numerous fractures and solution zones. Fractures are not, however, equally distributed throughout the aquifer, as is evidenced by the number of lower capacity wells (e.g., 300 gpm) that have been drilled in the near vicinity of highly productive wells.

# Water Levels

## Water Table

Depth to water was measured in 72 wells in February 1994 at a time when the aquifer water level should have reached its maximum recovery just prior to the start of the spring pumping season. Altitude of the water table above mean sea level was calculated and contoured as shown on figure 10-6. Ninety-three percent of the measurements vary between altitudes of 3,587 and 3,602 ft and average 3,594 ft. The lowest water levels occur near the center of the valley in the vicinity of Dell City and north of town near the location of the primary municipal supply well. The fault that forms the southern boundary of the valley does not appear to affect water levels in its vicinity. South of the delineated valley, low water levels occur in the vicinity of Highway 62-180.

The relative flatness (low gradient) of the water table results in a westerly increase in depth to water as the land surface altitude increases. Depth to water ranges from a few feet below the surface in the salt flats to more than 800 ft in higher elevations of the

Diablo Plateau (Kreitler and others, 1987). Within the irrigated region of the valley, depths to water range from 33 ft along the eastern side to 323 ft on the west.

## Seasonal Fluctuation

Water levels in the valley exhibit a seasonal fluctuation. During the irrigation season, the large quantity of water pumped from the aquifer results in a depressed water-table surface as more water is being withdrawn than can be immediately replaced. However, during the winter (nonpumping) season, the water table rebounds as additional water recharges the aquifer system, and cones of depression recover. The seasonal water-level fluctuation can be observed on the hydrograph of well 48-07-516 (fig. 10-5). The hydrograph shows that the aquifer response is in the range of 15 to 35 ft, depending on the amount of annual pumpage.

## Water-Level Change

Drawdown on the aquifer occurred immediately after irrigation wells began pumping in the late 1940's. The water level dropped at an average rate of 1.3 ft/yr for the next 30 yr as pumpage exceeded recharge to the aquifer. By the late 1970's, water-level declines of 25 to 45 ft had occurred throughout the valley. During the 1980's irrigation pumpage diminished somewhat, and water levels remained relatively constant or, in some

locations, rose slightly. Since the mid-1990's, water levels once again are on a downward trend, averaging 1 to 2 ft of decline per year.

# Water Quality

## Chemical Quality Characteristics

The groundwater underlying Dell Valley is generally brackish, very hard, and dominated by elevated levels of calcium, sodium, sulfate, and chloride. Water in the Dell Valley area can be classified as slightly to moderately saline, with TDS ranging from approximately 1,000 to more than 6,500 milligrams per liter (mg/L, and averaging about 3,500 mg/L (fig. 10-7). TDS is greatest along a north-south strip east to southeast of Dell City, where concentrations exceed 5,000 mg/L.

The prominence of calcium, sodium, sulfate, and chloride minerals in the groundwater can be traced to two dominant processes: (1) water flowing through the aquifer system and dissolving minerals along the flow path and (2) irrigation water percolating downward through the soil zone. Calcium and sulfate minerals are readily dissolved by groundwater that comes in contact with evaporite deposits in the Bone Spring and Victorio Peak Limestones. The very high hardness (as $CaCO_3$) value is also indicative of groundwater in a limestone/dolomite environment.

Irrigation water percolates with relative ease through the naturally saline soils and underlying gypsiferous caliche of the valley. However, some of the water applied to the

land surface is partially evaporated, which leaves behind a slightly more concentrated dissolved mineral solution. In order to leach salt minerals from the root zone of crops, relatively large amounts of water are applied annually to the porous land surface. Thus, each application of water delivers additional dissolved minerals, especially sulfates and chlorides, downward to the aquifer.

## 1992 Quality Survey

A water-quality survey was conducted in 1992 in which samples were collected from 30 wells and were analyzed for primary and trace inorganic minerals, nutrients, pesticides, and radionuclides. Sulfate was found to be the most prominent constituent, with concentrations ranging from 631 to 2,448 mg/L. Calcium, sodium, and chloride also attain high levels of concentration.

Water samples were analyzed for the following minor or trace inorganic constituents: arsenic, barium, copper, iodide, iron, manganese, selenium, and zinc. Concentrations were below detection limits in all samples except for one iron and five zinc analyses. However, even these did not exceed Federal Safe Drinking Water Standards.

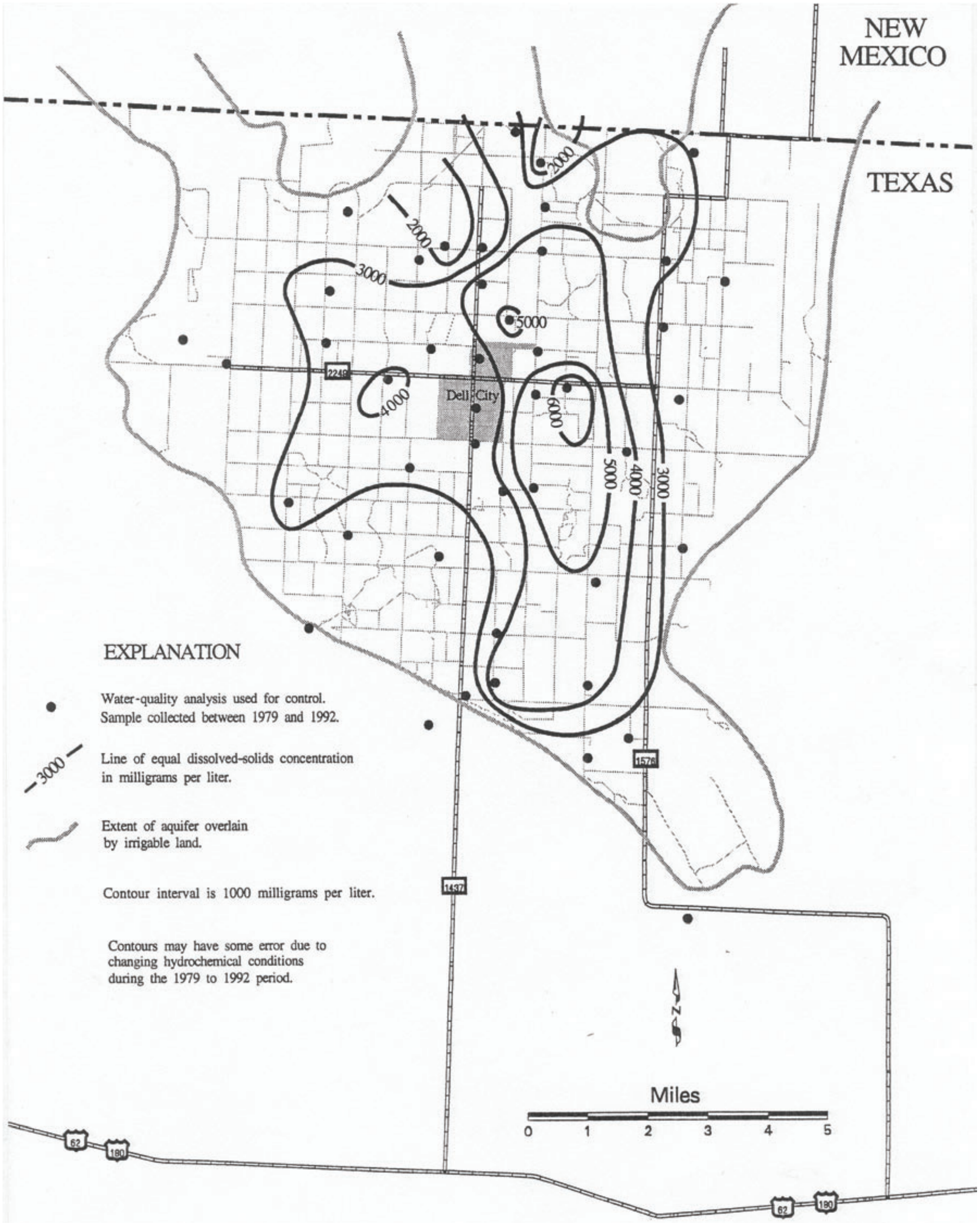


Figure 10-7: Dissolved-solids content, 1979 through 1992.

Nutrients in groundwater are various derivatives of nitrogen. When found dissolved in groundwater, nutrients are an indicator of contamination from, most commonly, decaying organic matter, human and animal waste, and fertilizers. Samples from the 30 wells were analyzed for ammonia, nitrite, nitrate, and Kjeldahl nitrogen. All ammonia and nitrite analyses were below detection limits, and Kjeldahl nitrogen values ranged from 0.2 to 1.0 mg/L. Eight of the samples had nitrate (as $NO_3$) concentrations in excess of the recommended limit for drinking water of 44.3 mg/L. The elevated nitrate concentrations are most likely derived from fertilizers transported rapidly by irrigation water return-flow.

Dissolved radionuclide activity above recommended safe levels was detected in sampled water. Fourteen of thirty well samples had measured gross alpha activity in excess of the recommended maximum safe level of 15 picocuries per liter (piC/L). Only two samples exceeded the recommended safe level for gross beta activity of 50 piC/L. Radioactive particles, or radionuclides, are found as trace elements in most rocks and soils. The source of most of the radioactive elements in the groundwater underlying Dell Valley is probably derived from the disintegration of volcanic rocks that occur in the near vicinity.

Because of the high permeability of the unsaturated zone above the aquifer, it is reasonable to expect contaminants from the surface to travel rapidly downward to the water table. Potential contaminants to the aquifer that pose a health hazard include various pesticides used in agriculture. A pesticide scan analysis, which included 48 organic compounds, was run on 5 well samples. No organic compounds were found above detection limits in any of the 5 wells.

## Suitability for Drinking Water

The quality of water for human consumption is always of concern. In 1974, the Federal Safe Drinking Water Act was adopted, and standards were set for drinking-water quality. Twenty-four of thirty chloride samples and all of the thirty sulfate samples from the 1992 water-quality survey exceeded set limits. Also, all 30 samples exceed set limits for total dissolved solids. Other constituents and quality characteristics that exceeded recommended standards in a lesser percentage of the samples include nitrate, gross alpha and beta, fluoride, and pH. Groundwater in Dell Valley is, therefore, not recommended for human drinking purposes without prior treatment, such as the desalination process now employed for the Dell City community system.

## Suitability for Irrigation

The suitability of groundwater for irrigation purposes is largely dependent on the chemical composition of the water. The extent to which the chemical quality will affect the growth of crops is determined in part by the climate, soil, management practices, crops grown, drainage, and quantity of water applied. Primary characteristics that determine the suitability of groundwater for irrigation are total concentration of soluble salts, relative proportion of sodium to other cations (calcium and magnesium), and concentration of boron or other toxic elements. These are termed the salinity hazard (specific conductance), sodium hazard (SAR), and boron hazard, respectively.

The specific conductance of water is used as an index of its salinity hazard. Specific conductance measured in 1992 in samples from 22 wells ranged from 1,438 to 8,810 micromhos per centimeter and averaged 4,720. All samples but one fell within the category of having a very high salinity hazard. Dissolved solids are approximately 78 percent of specific conductance.

High concentrations of sodium relative to calcium and magnesium in irrigation water adversely affect soil structure by forming a hard, impermeable crust that results in cultivation and drainage problems. An index used for predicting the sodium hazard is the sodium-adsorption ratio (SAR). SAR values computed from the analyses of the 30 well samples range from 0.3 to 7.9 and average 4.2.

Boron is necessary for good plant growth but rapidly becomes toxic at higher concentrations. Permissible limits of boron for various crops range from 0.67 to 3.00 mg/L. The concentration of boron in 30 well samples collected in 1992 range from 0.12 to 2.36 mg/L, and average 0.81 mg/L. Nineteen of the thirty samples exceed the lower limit; however, none exceed the upper limit. Water from the aquifer, therefore, appears to be acceptable for the irrigation of most semi-boron-tolerant crops.

Although the water is high in salinity, irrigated agriculture has been successful in Dell Valley owing to the high permeability of the soil, the balance of the dissolved minerals, and the low sodium percent. A study by Longenecker and Lyerly (1959) shows that 6 to 8 yr of water application definitely improved the chemical conditions in the irrigated soils versus uncultivated soils. With the application of sufficient quantities of water, resident salts in the soil profile are easily leached downward beyond the root zone of crops. Soil salinity, however, cannot be reduced below the salinity of the water used for leaching. Although the leaching process has been beneficial to crop growth, it has unfortunately caused a degradation of the quality of the groundwater because of irrigation return-flow.

## Water-Quality Change

Groundwater quality changes have been occurring since the 1940’s, when return-flow of water from the first irrigation wells began altering the natural chemical composition of the aquifer. Water applied to agricultural land has percolated down to the water table, leaching additional minerals on its way. Also, the drilling and open completion of hundreds of wells in the valley has created a condition in which zones containing poor-quality water can mix with all other water-bearing zones.

Over time, the concentration of individual dissolved constituents in the groundwater has increased. Typical water-quality change in the valley is illustrated in figure 10-8, which shows the increasing concentration of sulfate, chloride, sodium, and dissolved solids in samples collected from well 48-07-205 between 1948 and 1992. During this period, dissolved solids increased from 1,119 mg/L to 4,395 mg/L. The increase in sulfate

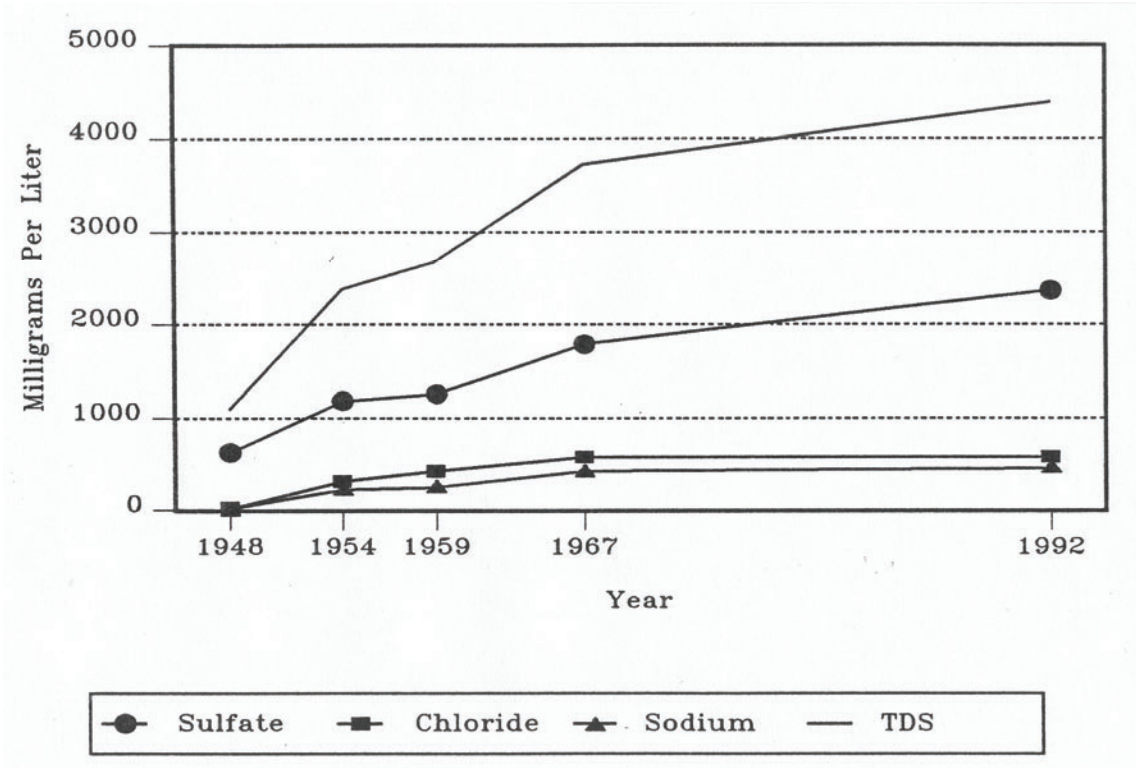


Figure 10-8: Water-quality change in well 48-07-205 from 1948 through 1992.

concentration alone represents approximately half of the total increase. The disproportionate increase in sulfate is primarily the result of dissolution of gypsum as irrigation return-flow water percolates downward through the soil zone.

# Conclusions

The amount of groundwater annually available on a sustainable basis in the Dell Valley region is contingent on rates of water-level decline and water-quality deterioration. A comparison of water-level and pumpage trends indicates that an annual pumpage of approximately 90,000 to 100,000 acre-feet can be maintained without continuously lowering the water table. At this rate, the seasonal water-level fluctuation remains at about 15 ft. An increase in annual pumpage to approximately 140,000 acre-ft, such as was common in the late 1970's and early 1980's, results in a noticeably declining water level and increases the seasonal water-level fluctuation to about 30 ft. The significance of a greater seasonal water-level fluctuation is that it steepens the hydraulic gradient, which increases the likelihood of the migration of highly saline water from the salt flats to the east.

The economy of the Dell Valley region is supported almost entirely by the agricultural industry, which in turn is dependent on the availability of groundwater. Today, the Bone Spring-Victorio Peak aquifer displays the effects of almost a half-century of intense use. A continuous 5 yr of water-level declines in the valley indicate that the groundwater resource is being depleted at a rate in excess of recharge. Local management decisions that will impact the viability of this aquifer and those that depend on it are currently being debated.

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# Chapter 11

# The Geology and Hydrogeology of the Capitan Aquifer: A Brief Overview

Matthew M. Uliana[1]

## Introduction

The Capitan aquifer occurs in the Capitan Reef Complex, an ancient reef that formed around the margins of the Delaware Basin (fig. 11-1) (Ashworth and Hopkins, 1995). The Delaware Basin was an embayment covered by a shallow sea that persisted throughout most of the Permian. Most of the reef complex was buried by tectonism and subsequent sedimentation; however, relatively undeformed remains of the reef are exposed in West Texas and New Mexico, with exceptional exposures in Guadalupe Mountains National Park (Bebout and Kerans, 1993). Remnants of the reef are also exposed in the Apache and Glass Mountains of West Texas.

This paper provides a brief overview of the structural history and stratigraphy of the Delaware Basin. The hydrostratigraphic relations between the Capitan Reef Complex and the other basin facies are also discussed. Groundwater occurrence and flow within the Capitan Reef Complex, as well as water-quality trends, are also addressed.

## Delaware Basin Structural History

During the Early Pennsylvanian Period, the North American and South American plates slammed into each other in a tectonic train wreck known as the Ouachita collision (Muehlberger and Dickerson, 1989). This event was responsible for the formation of a regional structure called the Ouachita-Marathon fold-thrust belt (fig. 11-2), which formed the southern shore of the Delaware Basin. Structurally high areas called the Diablo and Central Basin Platforms flanked the east and southwest edges of the basin. After convergence of the North and South American plates ceased in the Early Permian, extensive deposition of carbonates and siliciclastics occurred throughout the remainder of the Permian. In the later part of the Permian, the basin was cut off from the ocean, and evaporite deposition filled in the basin. By the latest Permian, the shallow seas that

[1] Terra Dynamics, Inc.

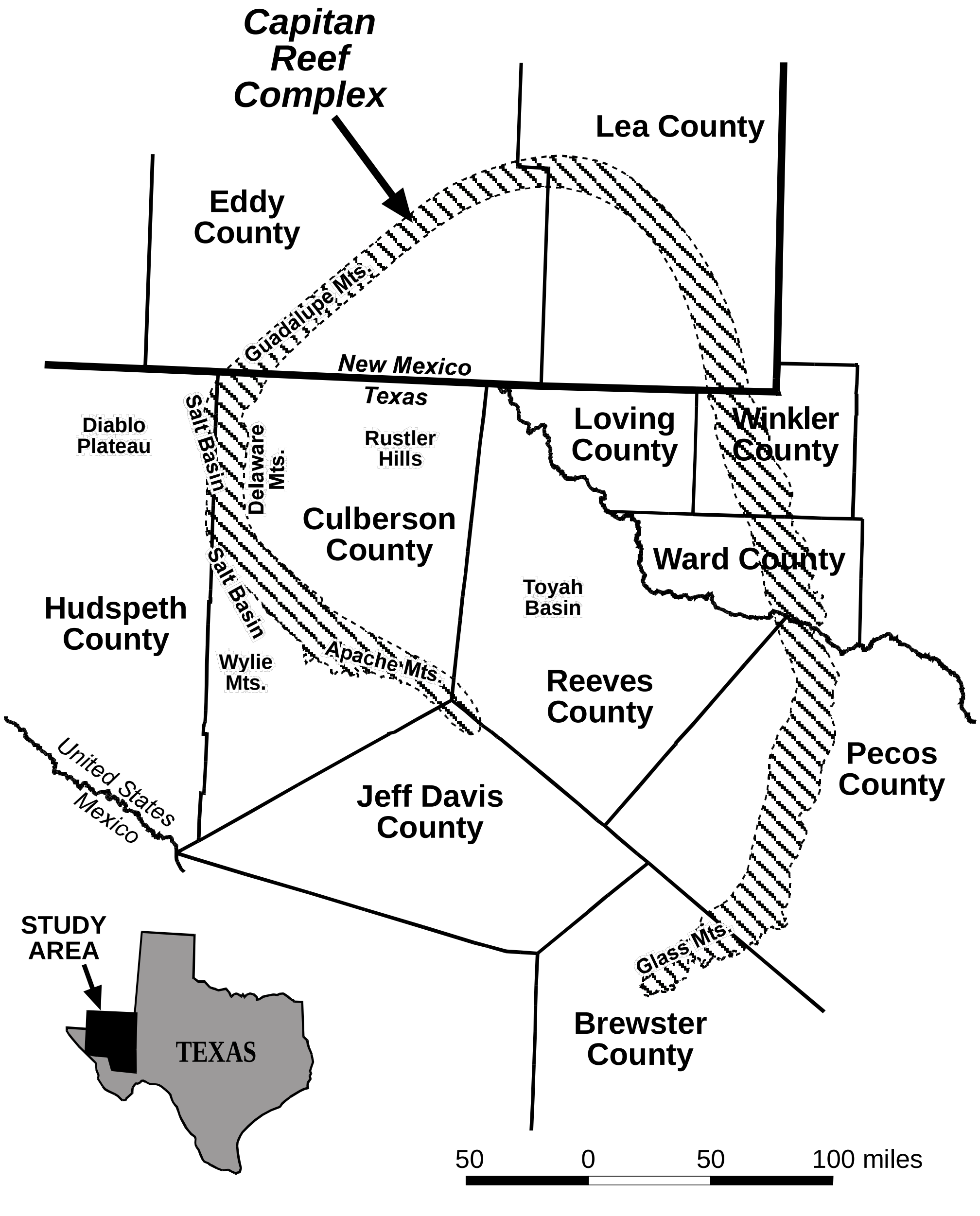


Figure 11-1: Location of the Capitan Reef Complex in western Texas and New Mexico (modified from Ashworth and Hopkins, 1995, and Dutton and others, 1999).

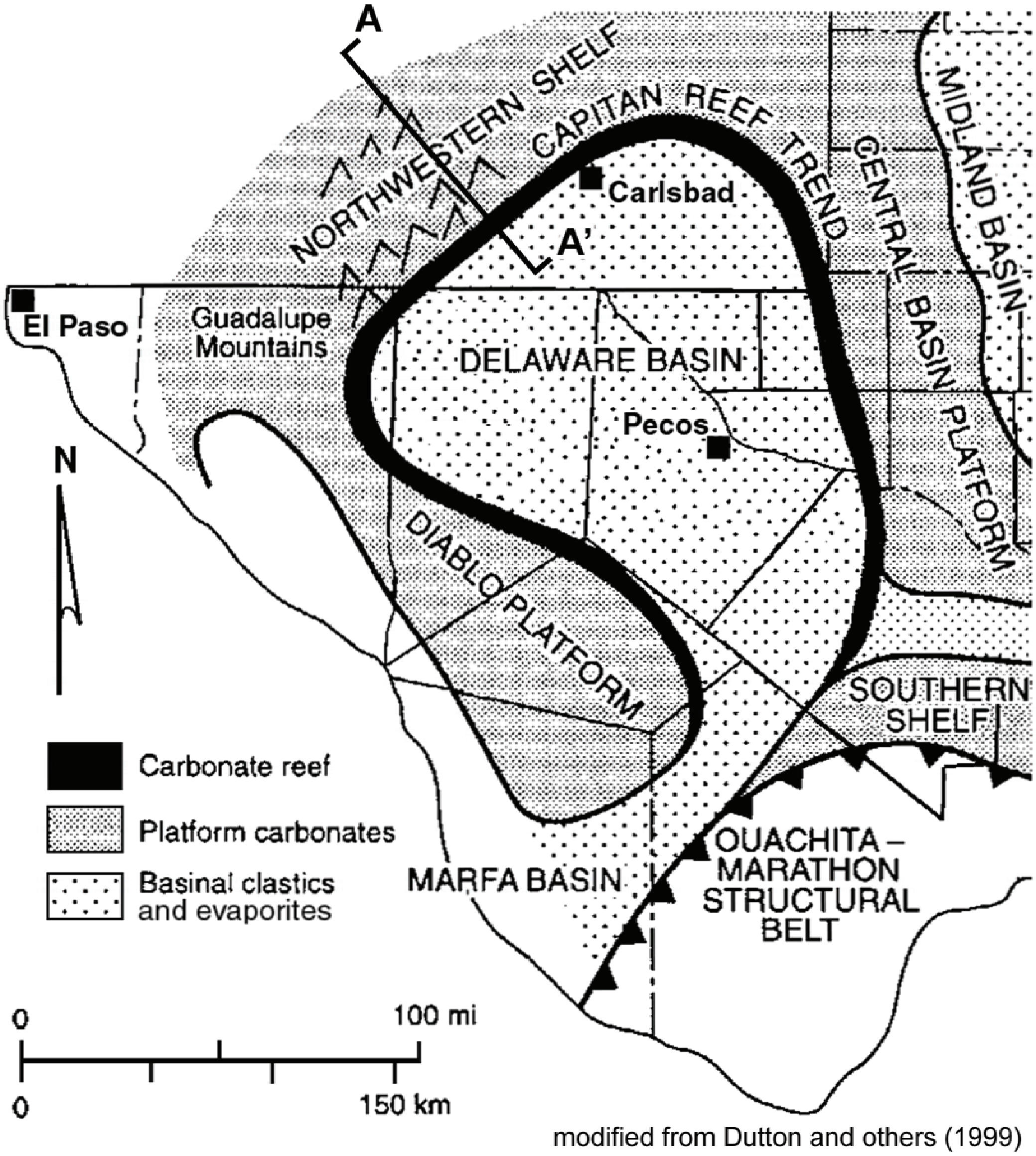


Figure 11-2: Paleogeographic setting of the Delaware Basin during the Late Permian.covered the region had completely withdrawn, and significant deposition of sedimentary strata would not occur in this area again until the Cretaceous.

# Delaware Basin Hydrostratigraphy

The Permian strata in the study area are divided into four series—the Wolfcampian, Leonardian, Guadalupian, and Ochoan (fig. 11-3). These series can be subdivided into three hydrostratigraphic facies on the basis of location relative to the center of the basin (fig. 11-4) (Hiss, 1975). The Capitan aquifer is composed of Guadalupian shelf-margin reef facies that include the Capitan Formation, parts of the Goat Seep Formation, and the Carlsbad Formation (Hiss, 1980). High primary porosity, high permeability, and extensive karst mark the reef facies of the Guadalupian.

The Leonardian and Wolfcampian shelf units of the Victorio Peak Limestone, Goat Seep Formation, and the Carlsbad Formation constitute the shelf aquifers (Hiss, 1980). The shelf facies are generally characterized by highly variable fracture-dependent permeability.

The Guadalupian and Ochoan basin facies of the Brushy Canyon, Cherry Canyon, and Bell Canyon Formations of the Delaware Mountain Group make up the basin aquifers (Hiss, 1980). These units are primarily siliciclastic fill deposited in the Delaware Basin and generally have much lower well yields and poorer water quality than the Capitan. Evaporites (anhydrite and gypsum) and some carbonates associated with the Castile and Rustler Formations and the Dewey Lake Redbeds of the Ochoan Series overlie the Guadalupian rocks in the Rustler Hills.

# Hydraulic Parameters and Water Occurrence

Transmissivities averaging 0.0624 $ft^2/s$ (0.0058 $m^2/s$) (Gates and others, 1980) and as high as 0.1872 $ft^2/s$ (0.0174 $m^2/s$) (Reed, 1965) have been measured in the Capitan aquifer. The high primary porosities and permeabilities of the reef facies are most likely augmented by extensive karstification, as exemplified in Carlsbad Caverns in southeastern New Mexico. In the Guadalupe Mountains of New Mexico, the aquifer is capable of providing large quantities of fresh water and is a significant water source for the City of Carlsbad (Ashworth and Hopkins, 1995). However, water quality throughout the reef facies in Texas is generally poor (Armstrong and McMillion, 1961; White, 1971; Richey and others, 1985).

Wells drilled into the Permian shelf facies exhibit highly variable yields, suggesting that permeability in the shelf aquifer units is dependent on fracture and karst porosity (Nielson and Sharp, 1990; Mayer and Sharp, 1998). High well yields and good water quality in the shelf facies are associated with regional fracture trends (Mayer and Sharp, 1998). Average transmissivities of wells drilled into cavernous zones in the shelf facies have been reported at 0.247 $ft^2/s$ (0.023 $m^2/s$) (Davis and Leggat, 1965) and 0.387 $ft^2/s$ (0.036 $m^2/s$) (Scalapino, 1950). However, the shelf facies generally are lower permeability and tend to yield lower quality water than do the reef facies.

| | | Diablo Plateau | Guadalupe Mts. | Rustler Hills/ Delaware Mts. | Wylie Mts. | Apache Mts. |
|---|---|---|---|---|---|---|
| PERMIAN | Ochoa | | | Castile Fm | | |
| | | | | Rustler Fm. | | |
| | | | | Castile Fm. | | |
| | | | | Dewey Lake Redbeds | | |
| | Guadalupe | | Carlsbad Fm. | undivided; Bell Canyon Fm. | | Tansill Fm. |
| | | | | | | Yates Fm. |
| | | | Capitan Fm. | | | Capitan Fm.; Seven Rivers Fm. |
| | | Goat Seep Fm. | Goat Seep Fm. | Cherry Canyon Fm. | | Munn Fm. |
| | | | Brushy Canyon Fm. | Brushy Canyon Fm. | | |
| | Leonard | Cutoff Shale | Cutoff Shale | | | |
| | | Victorio Peak Ls. | Victorio Peak Ls. | | Victorio Peak Ls. | Victorio Peak Ls. |
| | | Bone Springs Fm. | Bone Springs Fm. | Bone Springs Fm. | ? | |
| | Wolfcamp | Hueco Ls. | | | | |

Figure 11-3: Stratigraphy of the Permian Delaware Basin.

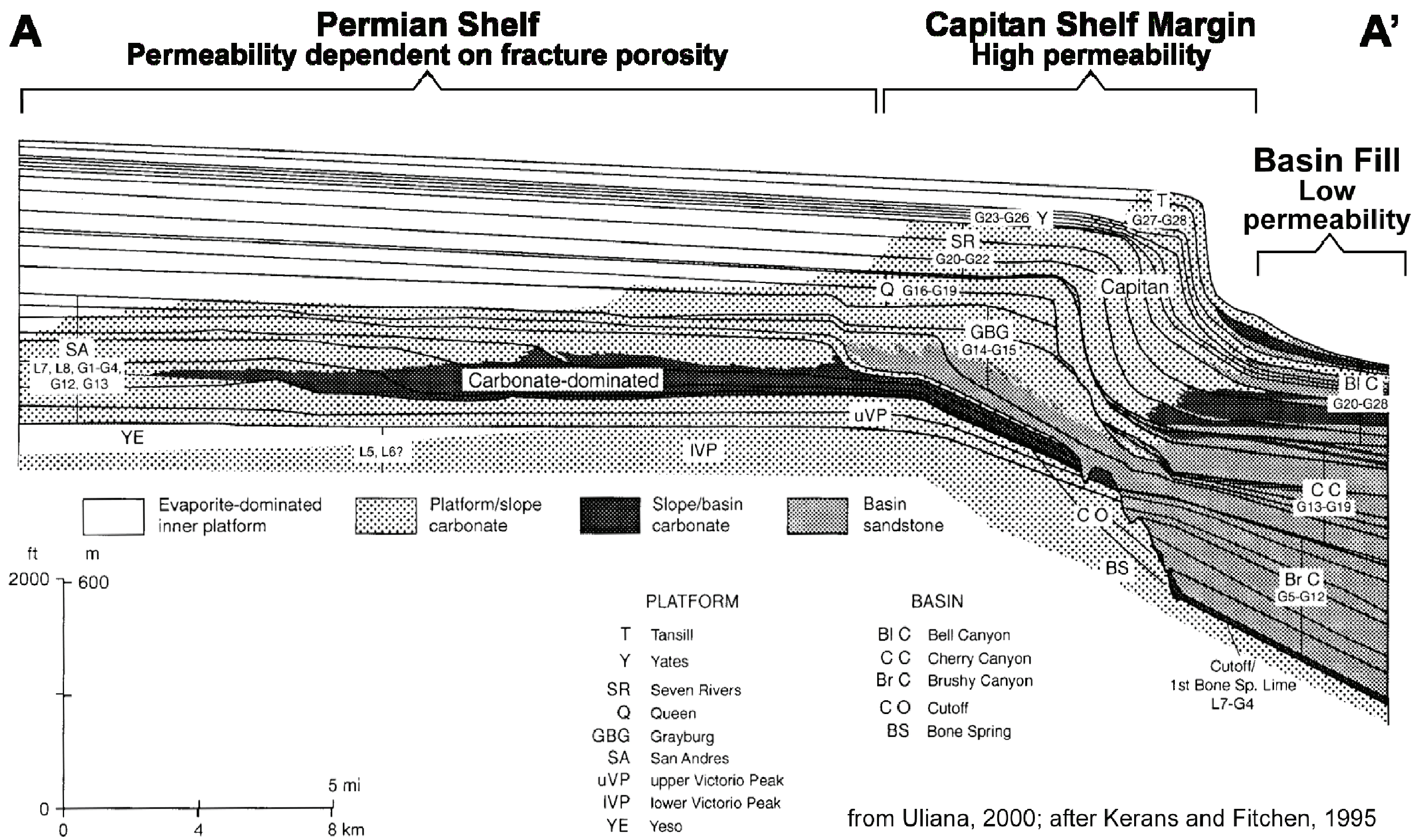


Figure 11-4: Generalized cross section of Delaware Basin showing the stratigraphic relationship between the three hydrostratigraphic facies (cross-section location shown on figure 11-2).

Basin sediments deposited during the Permian form aquifer units with low permeabilities, poor-quality water, and low well yields that range from 5 to 20 gpm (0.003 to 0.0012 $m^3$/s). The average hydraulic conductivity of the basin-fill facies is generally one to two orders of magnitude lower than the reef facies (Hiss, 1980; Nielson and Sharp, 1985), and the quality of water in the basin facies is generally much lower than the reef facies (Hiss, 1980).

Published transmissivity values from wells in the Delaware Basin are presented in table 11-1.

# Hydraulic Gradients and Groundwater Flow Paths

Hiss (1975, 1980) looked at the movement of groundwater in the Delaware Basin strata and examined the relationship between flow in the Capitan aquifer and in the surrounding basin and shelf facies. In general, groundwater flow in the basin and shelf facies is primarily toward the east. The high permeability of the Capitan aquifer results in concentrated flow along the trend of the reef, generally toward the north and northeast. Following uplift of the Guadalupe and Glass Mountains and before the excavation of the Pecos River Valley, flow in the Capitan aquifer was north and east to a main discharge point near present-day Hobbs, New Mexico (fig. 11-5a). Water exiting the Capitan aquifer discharged into the San Andres Limestone, where it then moved eastward to eventually discharge into streams draining to the Gulf of Mexico.

Following the deposition of the Ogallala Formation, the Pecos River Valley began to form across the Capitan Reef trend. The river valley eroded into the Permian and developed a hydraulic connection with the aquifer and eventually incised deep enough to drain water from the aquifer and reverse flow paths in the aquifer between Hobbs and Carlsbad, New Mexico (fig. 11-5b). Draining of the Capitan aquifer by the river also changed the hydraulic gradients and flow paths in the shelf and basin facies surrounding the aquifer. Development of the petroleum and groundwater resources in the area during the last 70 yr has drained additional water from the Capitan aquifer and has affected the gradients so much that the original terminal discharge area is now bypassed (fig. 11-5c).

West of the Apache and Guadalupe Mountains, the Capitan aquifer has been displaced into the subsurface by faulting and covered by up to 750 m of alluvial sediment in the Salt Basin. Kreitler and others (1990) speculated that groundwater in the Diablo Plateau of Hudspeth County may be flowing toward the southeast through the reef facies at depth.

Neilson and Sharp (1985) suggested that the high-permeability reef facies of the Capitan aquifer may provide a conduit for regional water flow through the Apache Mountains into the Toyah Basin. Although the geochemical and isotopic evidence presented by Uliana (2000) and Uliana and Sharp (2001) supports the regional-flow hypothesis, cross sections of the Apache Mountains published by Wood (1965) (fig. 11-6) indicate that the exposed reef facies are above the water table and that flow is most likely in fractures in the underlying basin facies.

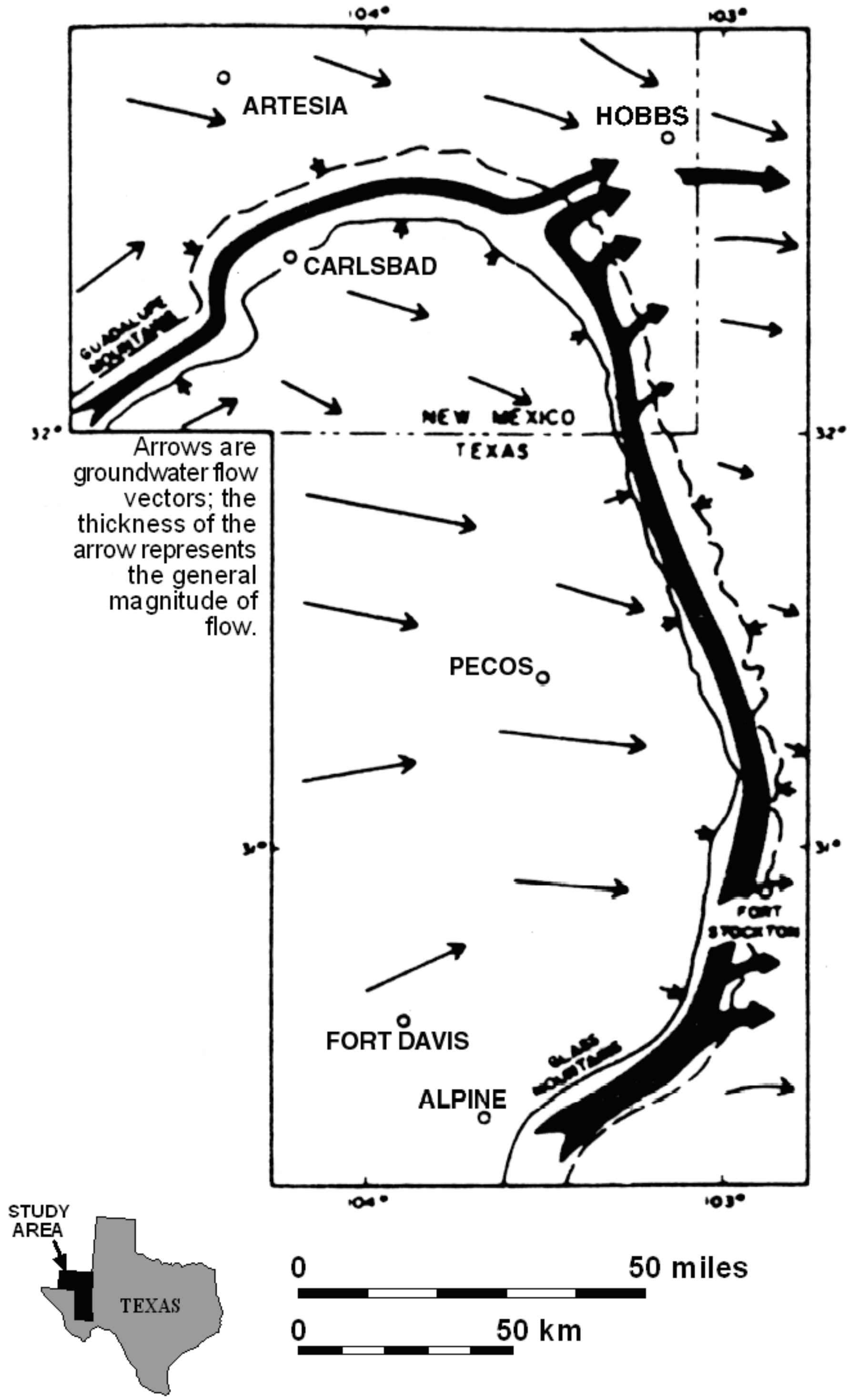


Figure 11-5a: Groundwater flow regimen prior to incision of the Pecos River and development of the oil and groundwater in the Delaware Basin (from Hiss, 1980).

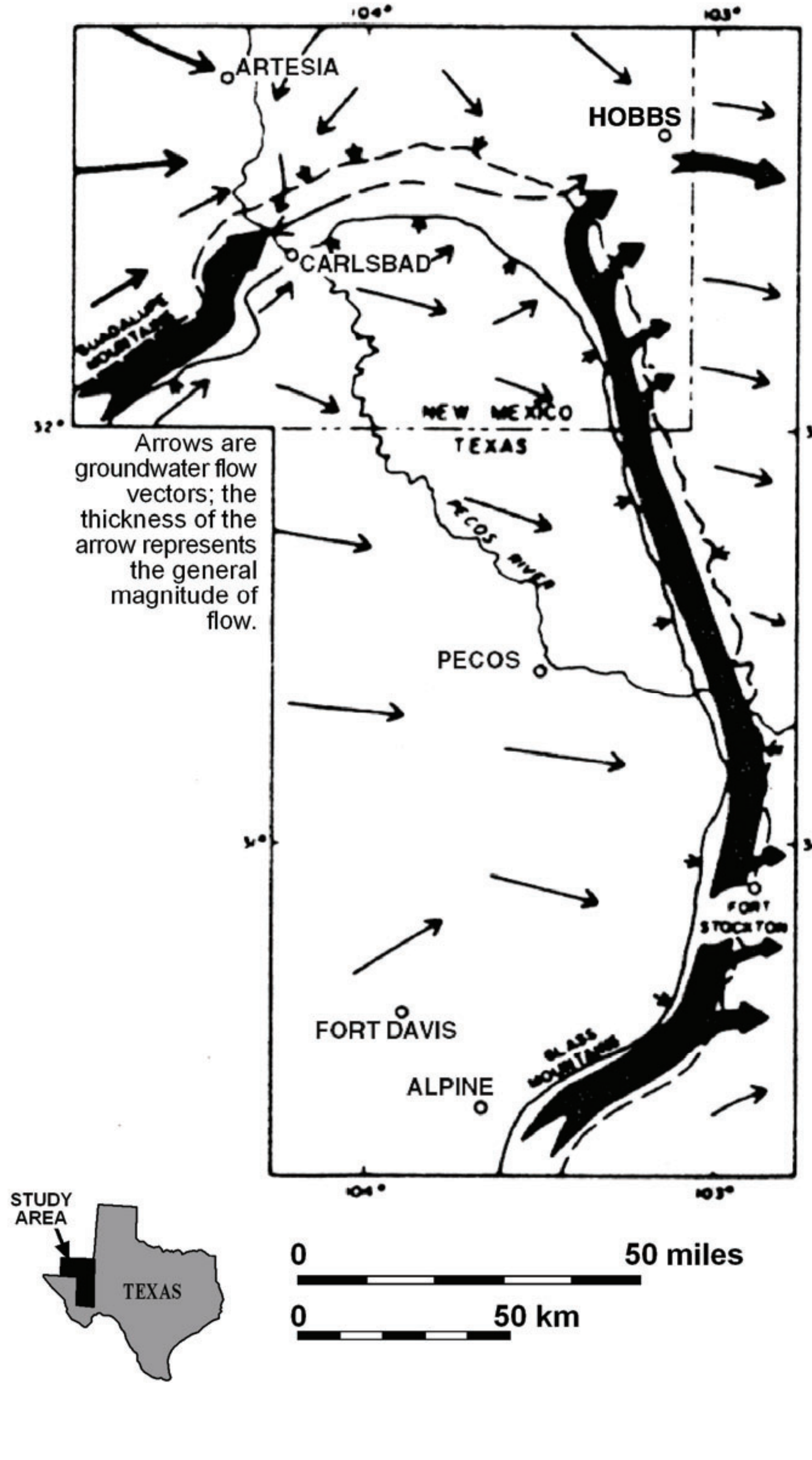


Figure 11-5b: Groundwater flow regimen influenced by incision of the Pecos River at Carlsbad into hydraulic communication with the Capitan aquifer (from Hiss, 1980).

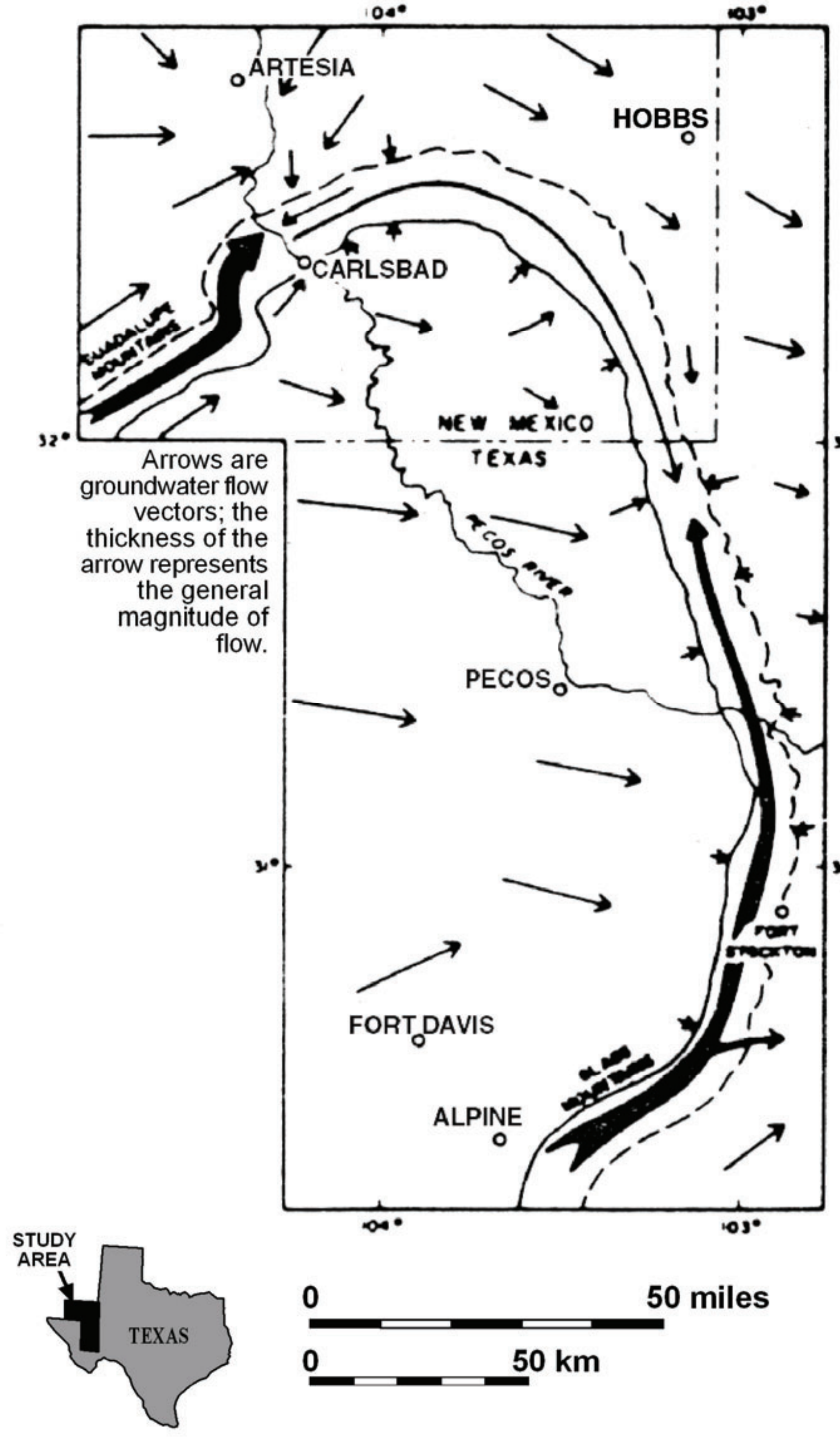


Figure 11-5c: Groundwater flow regimen influenced by both incision of the Pecos River and exploitation of oil and groundwater in the Delaware Basin (from Hiss, 1980).

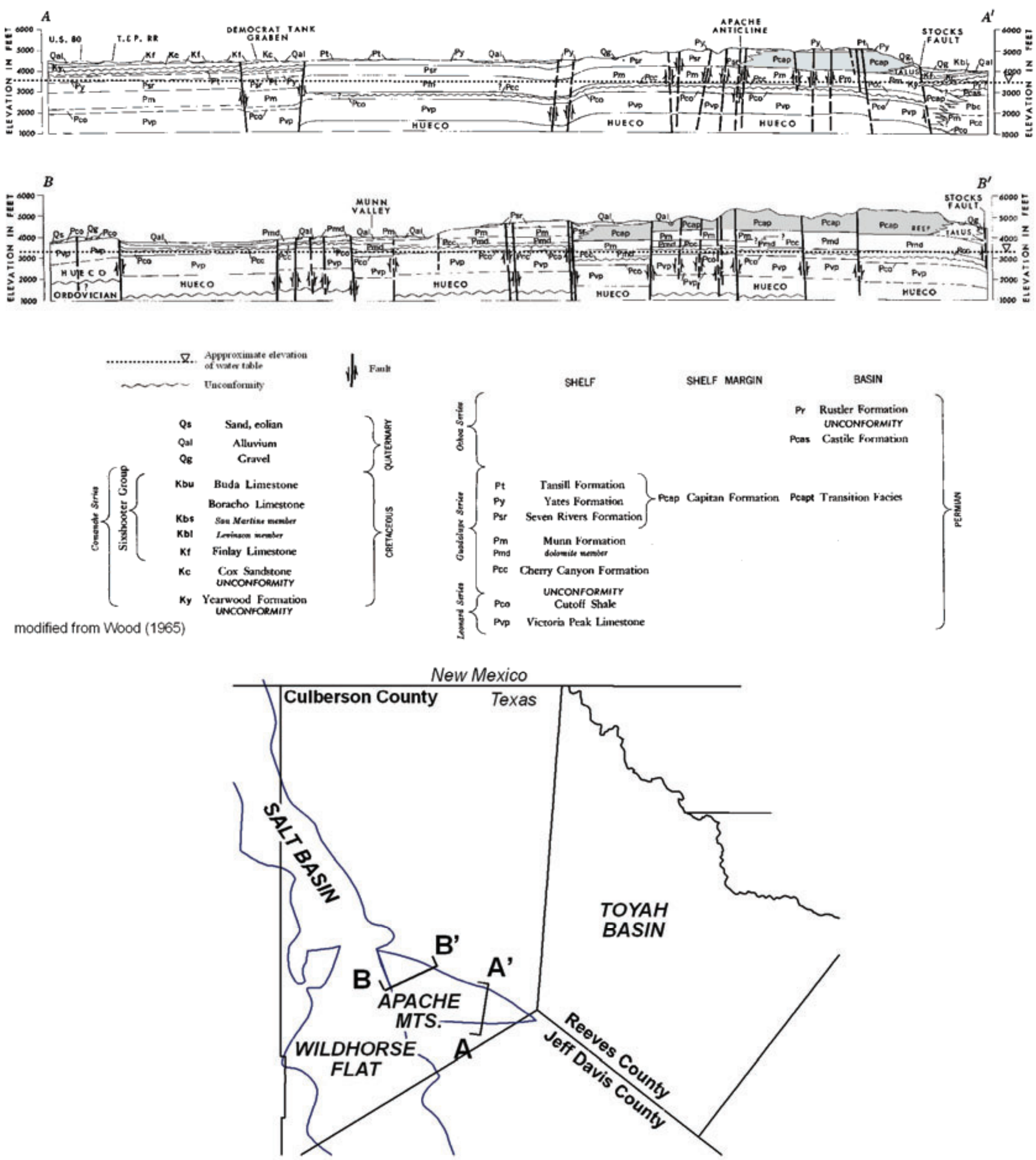


Figure 11-6: Apache Mountains cross sections showing the Capitan Reef facies and groundwater levels (water levels from Sharp, 1989).

## Water Quality

The aquifer generally contains water of poor quality and yields small to large quantities of moderately saline to brine water. Analysis of water samples from 17 reef facies wells in Texas indicates an average total dissolved solids concentration (TDS) of 3,059 mg/L and an average chloride concentration of 881 mg/L (Brown, 1997). These samples also indicate that the primary constituents are sodium, chloride, and sulfate. Because of the low quality, water pumped from the Capitan aquifer in Texas is primarily used for oil-reservoir waterflooding operations in Ward and Winkler Counties, with a small amount used for irrigation of salt-tolerant crops in Pecos and Culberson Counties (Ashworth and Hopkins, 1995). Water of the freshest quality is located on and near areas of recharge where the reef is exposed at the surface in the Guadalupe and Glass Mountains. The city of Carlsbad, New Mexico, uses Capitan water for a municipal supply.

## Conclusions

The Capitan aquifer is the remains of a vast reef that surrounded the Delaware Basin during the Permian. Permeability and well yields are generally high, but water quality tends to be too poor for municipal or irrigation use. The exception is in the areas where the reef is exposed, such as the Guadalupe Mountains of New Mexico. The regional flow paths in the aquifer have been affected by incision of the Pecos River and by development of the groundwater and petroleum resources in the area. The water is primarily sodium-chloride-sulfate water with an average TDS greater than 3,000 mg/L.

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# Chapter 12

# The Dockum Aquifer in West Texas

Robert G. Bradley[1] and Sanjeev Kalaswad, Ph.D.[11]

## Introduction

The Dockum aquifer, classified as a minor aquifer by the Texas Water Development Board (TWDB), extends over approximately 42,000 $mi^2$ primarily in the Panhandle region of north Texas (figs. 12-1, 12-2). A portion of the southern tip of the aquifer extends into Crane, Ector, Loving, Pecos, Reeves, Ward and Winkler Counties in West Texas. Although the Dockum aquifer can be an important source of groundwater for irrigation, public supply, oil-field activity, livestock and manufacturing purposes, deep pumping depths, poor water quality, low yields, and declining water levels have generally discouraged its use except locally.

The purpose of this article is to present a summary of the characteristics of the Dockum aquifer in West Texas. Much of the information presented in the article was obtained from previous literature and from TWDB records.

## Physiography and Climate

The area overlying the Dockum aquifer in West Texas is generally flat with a gentle slope toward the southeast-flowing Pecos River, which drains much of the region. Drainage north and east of the Pecos River typically is closed, with runoff collecting in swales, sinks and playas (Ashworth, 1990). The climate of the region is semiarid, with hot summers and mild winters (Larkin and Bomar, 1983). Mean annual precipitation in the Pecos River Valley is approximately 10 inches, and lake surface evaporation about 80 inches/yr. (Larkin and Bomar, 1983).

## Geologic Setting

The approximately 2,000-ft-thick Triassic sediments of the Dockum Group that form the Dockum aquifer consist of a series of alternating sandstones and shales (Cazeau, 1962). Individual sandstone units are light to dark or greenish-gray, buff, and red, and range in thickness from a few feet to about 50 ft. The red and maroon sandy shale units that separate the sandstones range in thickness from about 50 to 100 ft.

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[1] Texas Water Development Board

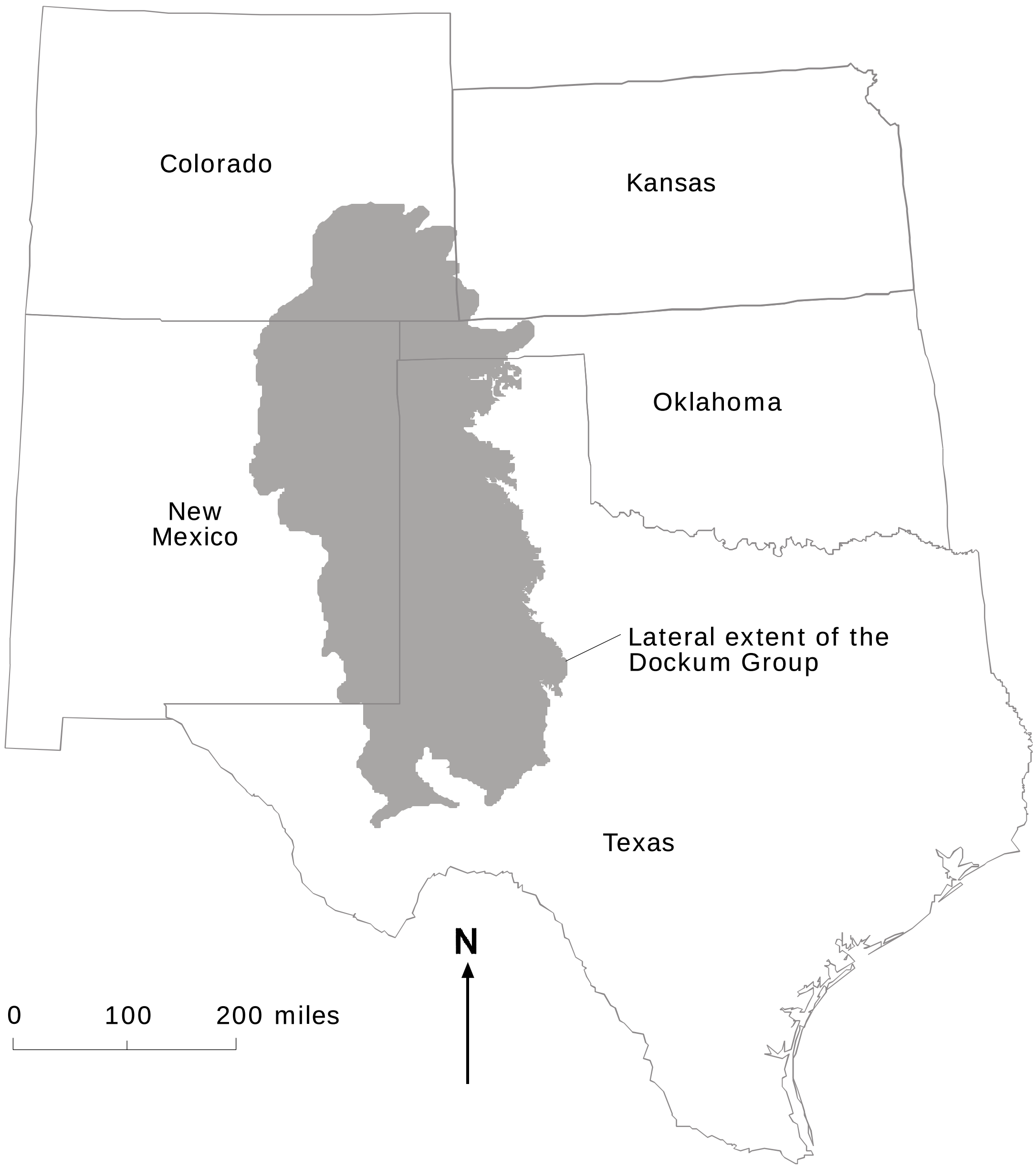


Figure 12-1: Location of the Dockum Group in Texas, New Mexico, Colorado, Kansas, and Oklahoma.

The formations within the Dockum Group (in ascending stratigraphic order) are: Santa Rosa Formation, Tecovas Formation, Trujillo Sandstone, and Cooper Canyon Formation. Locally the term *Santa Rosa* has been applied to the lower sandstone zones in the Dockum Group that may include all units of the Dockum Group except the upper mudstone.

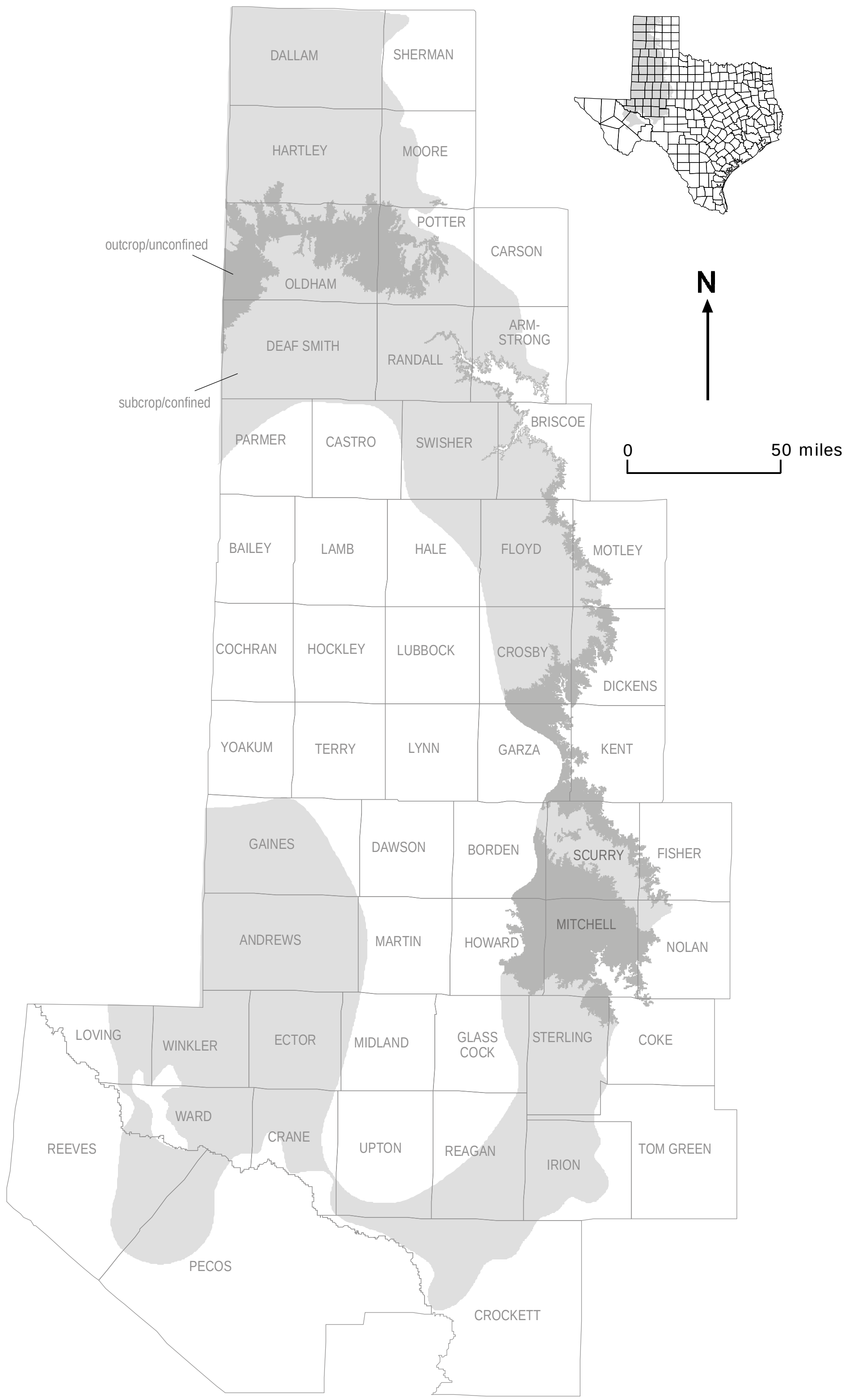


Figure 12-2: Location of the Dockum aquifer in Texas.

The basal unit, called the Santa Rosa Formation, rests unconformably on Upper Permian red beds and can be up to 130 ft thick (Lehman and others, 1992; Lehman, 1994a, b; Riggs and others, 1996). The Santa Rosa Formation is overlain by variegated mudstones and siltstones of the Tecovas Formation (Gould, 1907), which in turn is disconformably overlain by the 250-ft-thick Trujillo Formation composed of massive, crossbedded sandstones and conglomerates (Lehman, 1994a, b). The Cooper Canyon Formation consists of reddish-brown to orange mudstone, with some siltstone, sandstone and conglomerate (Lehman and others, 1992).

The Dockum Group is generally considered to represent sediments deposited in fluvial, deltaic, and lacustrine environments within a closed continental basin (McGowen and others, 1977, 1979; Granata 1981). The basin apparently received sediments from all directions, although in West Texas the source areas were primarily to the south and southwest (Fallin, 1989).

The beds of the Dockum Group are essentially horizontal, with very gentle dips toward the center of the main basin, whose axis trends approximately north-south. The dip varies considerably from location to location but is approximately 30 ft/mi (Rayner, 1963). In West Texas, the primary structural features are the Central Basin Platform in the east and the Delaware Basin in the west (Fallin, 1989).

The top of the Dockum Group is relatively flat and reflects the final filling of the Dockum Basin and the effects of postdepositional erosion. The opening of the Gulf of Mexico in the Cenozoic Period tilted the entire region toward the southeast.

# Hydrogeology

Recoverable groundwater in the Dockum aquifer is contained within the many sandstone and conglomerate beds that are present throughout the sedimentary sequence. The coarse-grained deposits form the more porous and permeable water-bearing units, whereas the fine-grained sediments form impermeable aquitards (Fallin, 1989). Consequently, the better groundwater flow zones are developed in the lower and middle sections of the stratigraphic sequence, where the coarse-grained sediments predominate. Locally, any water-bearing sandstone within the Dockum Group is typically referred to as the Santa Rosa aquifer. In the Pecos River Valley, the Dockum aquifer is usually known as the Allurosa aquifer (White, 1971).

In West Texas, the Dockum aquifer overlies Permian-age beds and is overlain by the Cenozoic Pecos Alluvium. The aquifer typically is under confined or partially confined conditions where Dockum Group sandstones are in contact with the Cenozoic Pecos Alluvium.

## Water Levels and Groundwater Flow

Potentiometric maps drawn from water levels measured by the TWDB between 1981 and 1996 indicate that groundwater flow in the Dockum aquifer in West Texas is generally to the southeast. Hydrographs of wells located in Crane, Ector, Loving, Reeves, Ward and Winkler Counties show a variety of water-level fluctuations. In Loving, Ector and Reeves Counties, the water table appears to have declined markedly whereas in Ward and Winkler Counties, it has remained relatively stable or has declined only slightly. The most significant water-level decline (almost 85 ft) was recorded in well 28-39-401 in Ector County. The decline presumably was the result of pumping in a nearby municipal water-supply well.

## Recharge

The Dockum aquifer is recharged by precipitation over areas where Dockum Group sediments are exposed at the land surface. Groundwater in the confined portions of the Dockum aquifer most likely originated as precipitation that fell on outcrops in eastern New Mexico. This recharge ceased when the Pecos and Canadian River Valleys were incised during the Pleistocene between the present-day Dockum aquifer in Texas and the paleo-recharge areas to the west (Dutton and Simpkins, 1986).

The Dockum aquifer is also recharged by upward leakage from the underlying Permian aquifer (Bassett and others, 1981; Bentley, 1981; Wirojanagud and others, 1984; Orr and others, 1985). Downward leakage into the Dockum aquifer occurs from the overlying Cenozoic Pecos Alluvium as a result of hydraulic-head differences between the aquifers (Dutton and Simpkins, 1986; Nativ and Gutierrez, 1988). Estimated annual recharge for outcrop areas and other areas in contact with overlying aquifers for the entire Dockum aquifer in Texas is approximately 31,000 acre-ft.

## Aquifer Properties

The hydraulic properties of the Dockum aquifer vary considerably from location to location. In West Texas, well yields measured by the TWDB ranged from approximately 23 gallons per minute (gpm) in Crane County to 353 gpm in Reeves County. Similarly, specific capacity ranged from 5.3 (Wink County) to 25 (Reeves County).

An aquifer test conducted on City of Kermit wells (Winkler County) by the TWDB in 1957 yielded an average transmissivity of 4,600 $ft^2$/day. These wells are completed in the Santa Rosa Sandstone that was described by Garza and Wesselman (1959) as a massive sandstone unit of limited areal extent. The storage coefficient was approximately $2.5 \times 10^{-4}$, which suggests that the aquifer in the test area is confined to partially confined.

## Groundwater Quality

Groundwater in the Dockum aquifer generally is of poor quality. It is characterized by decreasing quality with depth, mixed types of water, concentrations of total dissolved solids (TDS) and other constituents that exceed secondary drinking water standards over most of the area, and high sodium levels that may be damaging to irrigated land.

The chemical quality of water in the Dockum aquifer in West Texas ranges from fresh (TDS <1,000 milligrams per liter [mg/L]) in outcrop areas to moderately saline (TDS between 3,000 and 10,000 mg/L). Fresh water generally is present only at the edges of the Dockum basin, especially in outcrop areas where the aquifer is recharged. TDS ranges from 473 mg/L (Winkler County) to 4,040 mg/L (Reeves County). Water from the Dockum aquifer is typically hard, with $CaCO_3$ concentrations ranging from 203 mg/L (Ector County) to 1,394 mg/L (Crane County).

Where overlain by the Cenozoic Pecos Alluvium, groundwater in the Dockum aquifer is characterized by Ca-$SO_4$-mixed-anion-type waters. Groundwater samples collected from Ector County had gross alpha particle concentrations of 6 to 23 picocuries per liter (piC/L). The MCL established by the Texas Natural Resource Conservation Commission for gross alpha particle activity limit is 15 piC/L. Groundwater samples from Crane County had maximum radium-226 and radium-228 concentrations of 6.8 piC/L and 5 piC/L, respectively. The MCL for combined radium-226 and radium-228 is 5 piC/L. The occurrence of uranium in the Dockum Group has been known for years (McGowen and others, 1977) and is the source of the high concentrations of radium-226 and radium-228 detected in the groundwater samples.

Sodium in groundwater is a constituent that has neither an MCL nor a secondary standard but is still a concern where the water is used for irrigation purposes. Sodium adsorption ratios higher than 18 (which typically result in excess sodium in the soils) were detected only in groundwater samples from Ector County. These same samples also had residual sodium carbonate (RSC) values greater than 2.5 meq/L, suggesting that the water was not suitable for irrigation.

## Discharge

Discharge of groundwater from the Dockum aquifer occurs at pumping wells, small springs that contribute to stream base flow in the outcrop, evapotranspiration, and cross-formational flow. The greatest amount of discharge occurs from the pumping of wells installed in the aquifer.

Irrigation and public supply use is limited to areas of the Dockum aquifer where the water quality is acceptable, depth to water is shallow, and a sufficient thickness of sandstone exists to make the aquifer productive. Municipal users of Dockum aquifer water include the cities of Barstow, Kermit and Pecos. The Colorado River Municipal Water Authority also uses water from the Dockum aquifer.

Springs occur in areas where the Dockum sediments intersect the water table. Brune (1981) described springs issuing from the Dockum aquifer along the Pecos River Valley. Many of these springs are now dry or have lower flows than they did in the past.

# Conclusions

The Dockum aquifer in West Texas occupies a relatively small area and is only locally important where sufficient sandstone thickness and acceptable water quality are present. High TDS concentrations and salinity limit its use for many purposes.

Recharge of the Dockum aquifer only occurs in areas where the sandstone units are exposed at the surface or are in contact with overlying aquifers. However, since much of the Dockum aquifer in West Texas is confined, it receives little recharge so any water withdrawn from it is not immediately replenished.

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# Chapter 13

# Igneous Aquifers of Far West Texas

Andrew Chastain-Howley[1]

## Introduction

The igneous aquifers of far west Texas are currently under review for a study financed by the Texas Water Development Board (TWDB). This study is analyzing the igneous aquifers in the tri-county area of Brewster, Jeff Davis, and Presidio Counties (fig 13-1). The igneous units also extend into southern Culberson County and southern Reeves County, but their extent is small compared with that of the other three counties.

The overall area covered by the igneous aquifers exceeds 5,000 mi$^2$, and the greatest measured depth of these units is 6,032 ft just north of Valentine. The average thickness of the igneous units is probably over 1,000 ft.

## Igneous Aquifers Geology

The Igneous aquifer is not a single aquifer like the Ogallala or Edwards aquifers. The Fort Stockton, Marfa, and Emory Peak-Presidio sheets of the Geologic Atlas of Texas show over 40 named volcanic units (table 13-1), not counting those in Big Bend National Park. Many of the units have been subdivided by more detailed mapping. "Igneous aquifers" would be a better name and should include the entire area where volcanic rocks crop out or are present beneath the alluvial cover—approximately 5,000 mi$^2$.

These volcanic rocks were formed mainly within the Tertiary Period between 39 and 31 million years ago (Ma). The approximate extent of these volcanic eruptive units and their respective chronology are shown in figure 13-2a through 13-2c. The volcanic rocks consist of a complex layering of vents, flows, and interbedded volcanic-sedimentary units, which were deposited in the many intervals between eruptions. This layering has led to the very complex interrelationships between the rock units. Figures 13-2a through 13-2c show the locations of the volcanic centers, which were most active in each of the main phases of volcanic activity. The most obvious trends are the main-center shifts from the south in the early phase (48 to 39 Ma), to the north in the middle phase (39 to 35 Ma), and back to the south again in the late phase (35 to 27 Ma). The overall geological map showing the surface outcrops related to these volcanic centers is depicted on figure 13-3.

[1] Water Prospecting, LLC

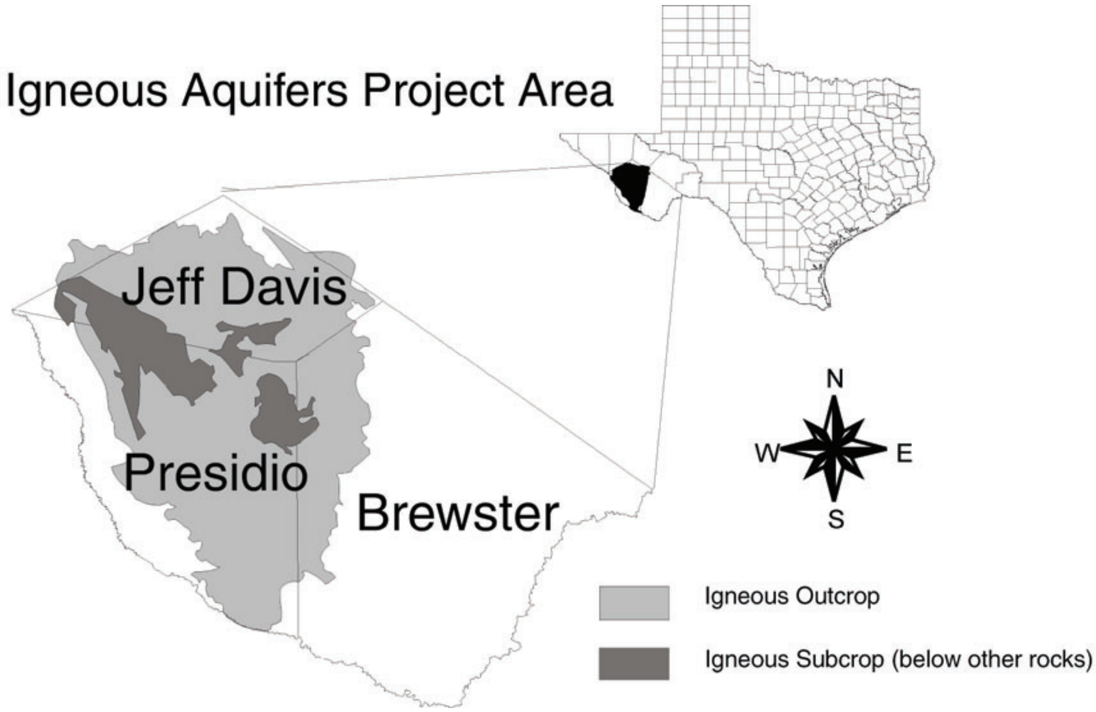


Figure 13-1. Igneous Aquifers Project Area within Brewster, Jeff Davis and Presidio Counties

The aquifers within this study area are found within three distinct geological-type aquifer units:

- **Igneous extrusive aquifers** (basalts, trachytes, rhyolites, tuffs)—generally referred to as volcanic.
- **Igneous sedimentary aquifers** (sandstone and conglomerate )—formed by the erosion of volcanic rocks and may be interbedded with volcanic units (e.g., Tascotal Formation).
- **Structurally controlled aquifers** (fault and fracture zones)—water-bearing capacity of the extrusive aquifers is generally structurally controlled. This unit refers to the improvement of the water-bearing capacity of all other units where faults and fractures occur.

The igneous geology also includes igneous intrusive rocks (these consist of the volcanic plugs and other slow-cooling igneous units). These intrusive rocks are important for

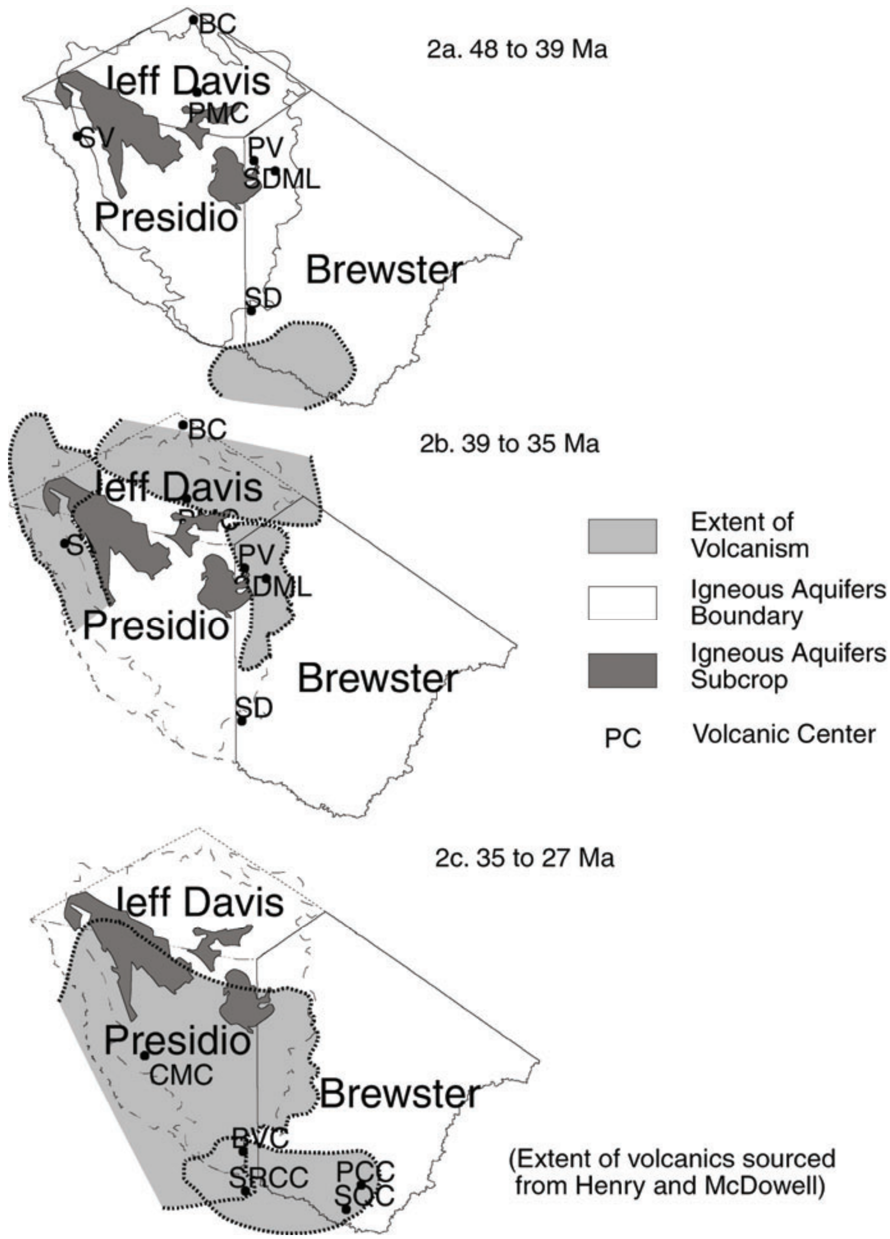


Figure 2a to 2c. Generalized history of the Trans-Pecos volcanic field (stipled pattern) separated into three phases, following Henry and McDowell (1986). 2a). Early phase consisting of Christmas Mountains Intrusions (XM) and the Alamo Creek Basalt (ACB). 2b). Middle phase consisting of Buckhorn Caldera (BC), Paradise Mountain Caldera (PMC), Paisano Volcano (PV), Sierra Vieja (SV), the Southern Davis Mountains Mafic Lavas (SDML), and Solitario Dome (SD). 2c). Late phase and early tensional phase consisting of the Chinati Mountains Caldera (CMC), the Siera Quemada Caldera (SQC), Pine Canyon Caldera (PCC), the Bofecillos Volcanic Complex (BVC), and the Sierra Rica Caldera Complex.

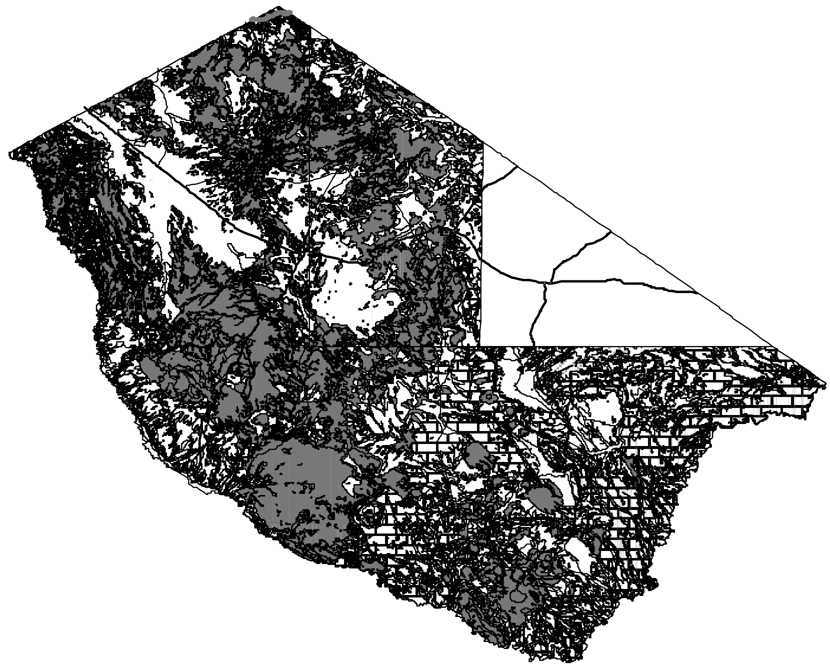

Figure 13-3: Surface geology of study area.

spring flow in certain areas but are not classed as an aquifer type because they do not have extensive storage.

The groundwater resources of the Igneous aquifers have been only cursorily studied but have tremendous potential. The entire municipal supplies of Alpine, Marfa, and Fort Davis and the supplies of the three commercial farms in the vicinity of those towns are

produced from an area of about 5 $mi^2$ and from at least five different volcanic aquifers. Recent studies by Brown and Caldwell (2001) suggest that much of the water in the Ryan Flat Bolson is also coming from the Igneous aquifers. Therefore, the six main groundwater supply fields are without exception pumping large volumes of water from comparatively small areas within the Igneous aquifers. This fact alone should show that there is great potential for withdrawal from these aquifers.

## Igneous Extrusive Aquifers

Igneous extrusive aquifers are mainly located in interflow zones. These often include vesicular zones near the tops of flows and rubble at the base of the overlying flows. Water is also found in the cooling fractures, such as those seen in many of the outcrops in the Davis Mountains. Because the porous zones are generally separated by dense flow rocks (mainly basalt and trachyte in this area), the aquifers are usually poorly connected except in the vicinity of faults or fracture zones. The effect is similar to sand layers separated by shale layers. Even within a single formation there may be multiple interflow zones/aquifers.

Examples of this phenomenon are available from studies around the world. A British Geological Survey study from the Deccan Plateau in India evaluated the hydrogeological capacity of basalt flow layers using down-hole geophysics and pumping-test interpretation. This same methodology could be used in future studies to outline the possibilities and complexity of these Igneous aquifer systems. Studies in the Snake River Plain in Idaho also show the complexity of these systems that are from a similar time period as the Far West Texas Igneous units.

The Igneous extrusive rocks in the tri-county area are varied in content and extent (fig. 13-4). There are extrusive rocks that include lava flows, ash-flow tuffs, and detrital rocks, which include erosional rubble. Studies conducted by Woodward (1954) and Wightman (1953) outline the complexity of the igneous rocks around Valentine. Woodward recorded over 40 different lava flow or tuff units within the 6,032 ft of volcanics from the Killam oil test well.

## Igneous Sedimentary Aquifers

These aquifers include both the Tertiary and Quaternary sedimentary units that have formed from erosion of local volcanic rocks. The Perdiz and Tascotal are the main aquifers that were deposited at around the time of volcanism. Quaternary bolson deposits in the Valentine area (Ryan Flats, Lobo Flats) are also made up of volcanic sediments and are therefore included in this unit as shown in figure 13-5.

Review of oil-well logs and recent studies (Brown and Caldwell, 2001) suggests that the bolson and underlying Igneous extrusive aquifers may be interconnected. Additional

work is needed, but sufficient data may be available to begin evaluating these aquifers. If, however, the bolsons are being recharged from the underlying volcanic aquifers, recharge rates may be more complicated than has been assumed. The extent and hydrologic attributes of the volcanic aquifers are likely to be the controlling mechanism. Other Quaternary alluvial deposits in general have too much clay to make good aquifers. There are reports of strong water flows from the alluvium (Sunny Glen area and southern Jeff Davis County near Point-of-Rocks) being possibly related to stream-channel deposits. However, these are generally of small areal extent.

## Structurally Controlled Aquifers

The structure of the Igneous units is very complicated. The basin-and-range faulting that created the Rio Grande valley has created a number of northwest-southeast-trending structures that are highlighted in the bolson valleys and in the McCutcheon Fault Zone on the northern edge of the Davis Mountains. figure 13-6 shows the regional trends of the fault and fracture systems within the Igneous aquifers area. There are a number of springs

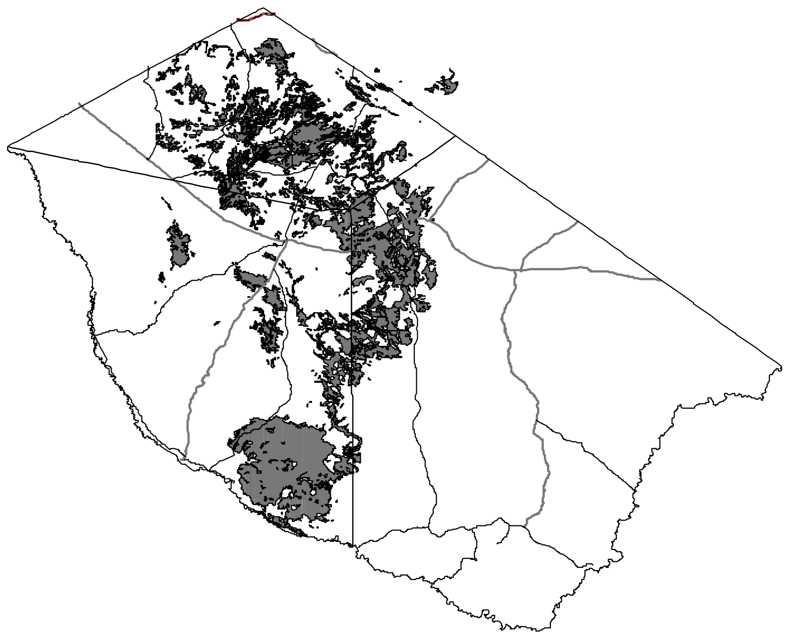

Figure 13-4 Igneous extrusive aquifer units---surface utcrop withing main study area.

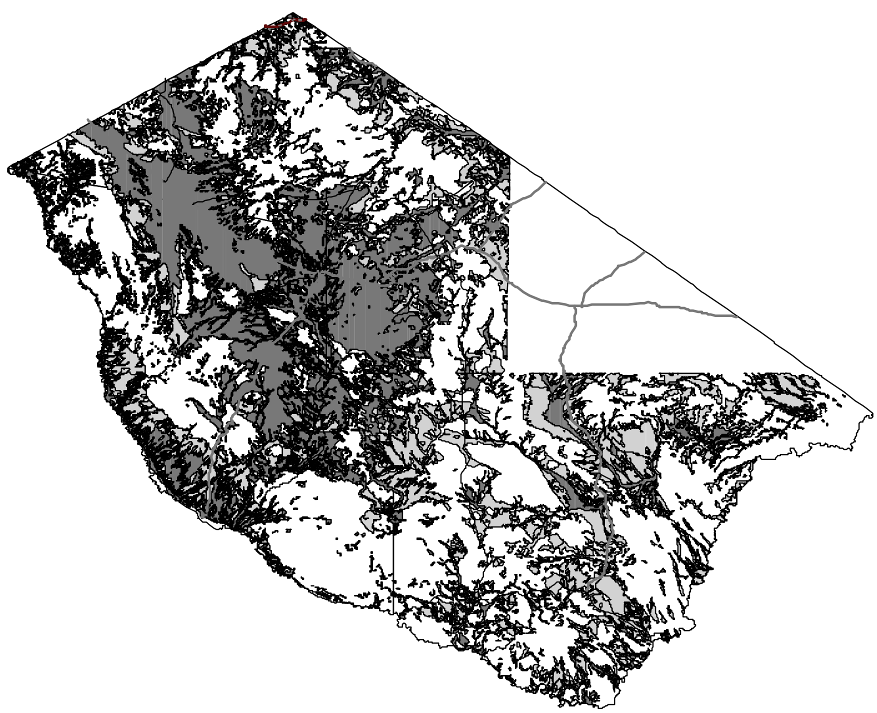

Figure 13-5 Igneous sedimentary aquifers units---surface outcrop within main study area.

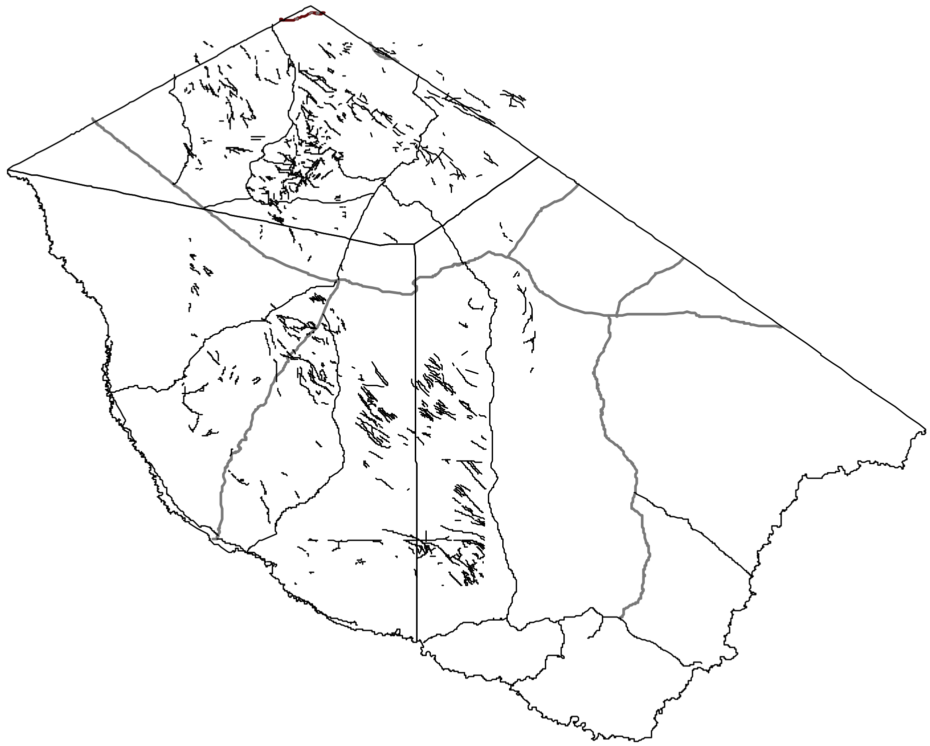

Figure 13-6: Structural features (faults) within main study area.

associated with faults and fractures, suggesting that there is hydraulic connection between units through faults.

Even today there is basin-and range-movement within the area. The area is seismically active and has produced numerous small earthquakes. The overall structural deformation sequence is well described in Henry (1998) and is not discussed in detail here. However, the main structural trends in this region are northwest to southeast, which includes the basin-and-range faulting in the Ryan Flat area and along the Rio Grande.

## Subsurface Stratigraphy

The most complex geological details are those found underground. A small number of oil-well tests have been drilled through the Igneous rocks in this area, and these wells give the basic interpretation of the thickness of Igneous rocks in the study area (fig. 13-7). However, because of the large amount of faulting in this area and the scarcity of drill holes, this data should be used carefully. The outside contour line equates to a 2,000-ft thickness. Most of these oil-well tests were drilled in valleys and close to the edge of

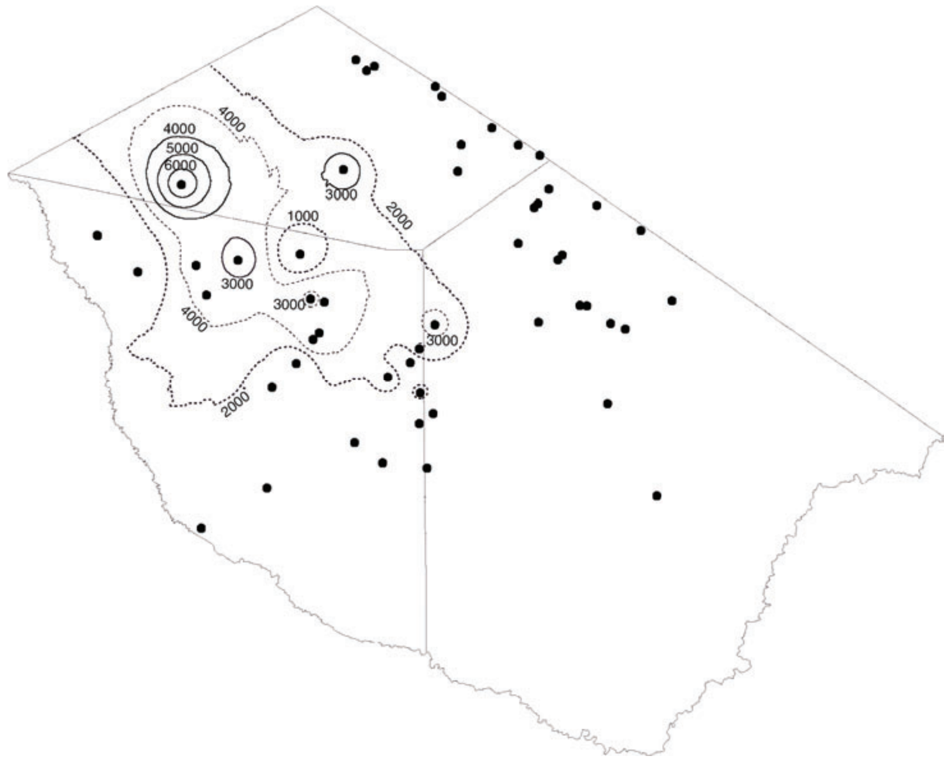


Figure 13-7: Approximate extent and depth of Igneous rocks derived from oil-well logs.

volcanic outcrops. Therefore, they will probably mask the true thickness of the igneous units (the depth is likely to be greater than that shown in figure 13-7).

Geophysical techniques are being used in this area to create a more detailed model of the subsurface. Gravity, magnetics, and remote-sensing data are currently being evaluated. In addition, all available downhole sample logs and geophysical logs are being studied to determine any gross changes in stratigraphy.

Previously published gravity data (Mraz and Keller, 1980) give an indication of the basic structural models these data can provide (fig. 13-8).

# Hydrogeology

The hydrogeology of the Igneous aquifers is very complex. There are many discrete and interconnected aquifers. The faulting and fracturing prevalent in the rocks of this region also increase the chances of connection between units. Even within aquifers that are commonly thought to be comparatively homogeneous (e.g., Ryan Flat Bolson), we have determined that the flow is very complex and consists of many different flow units (over 40 different units reported by Woodward [1954]).

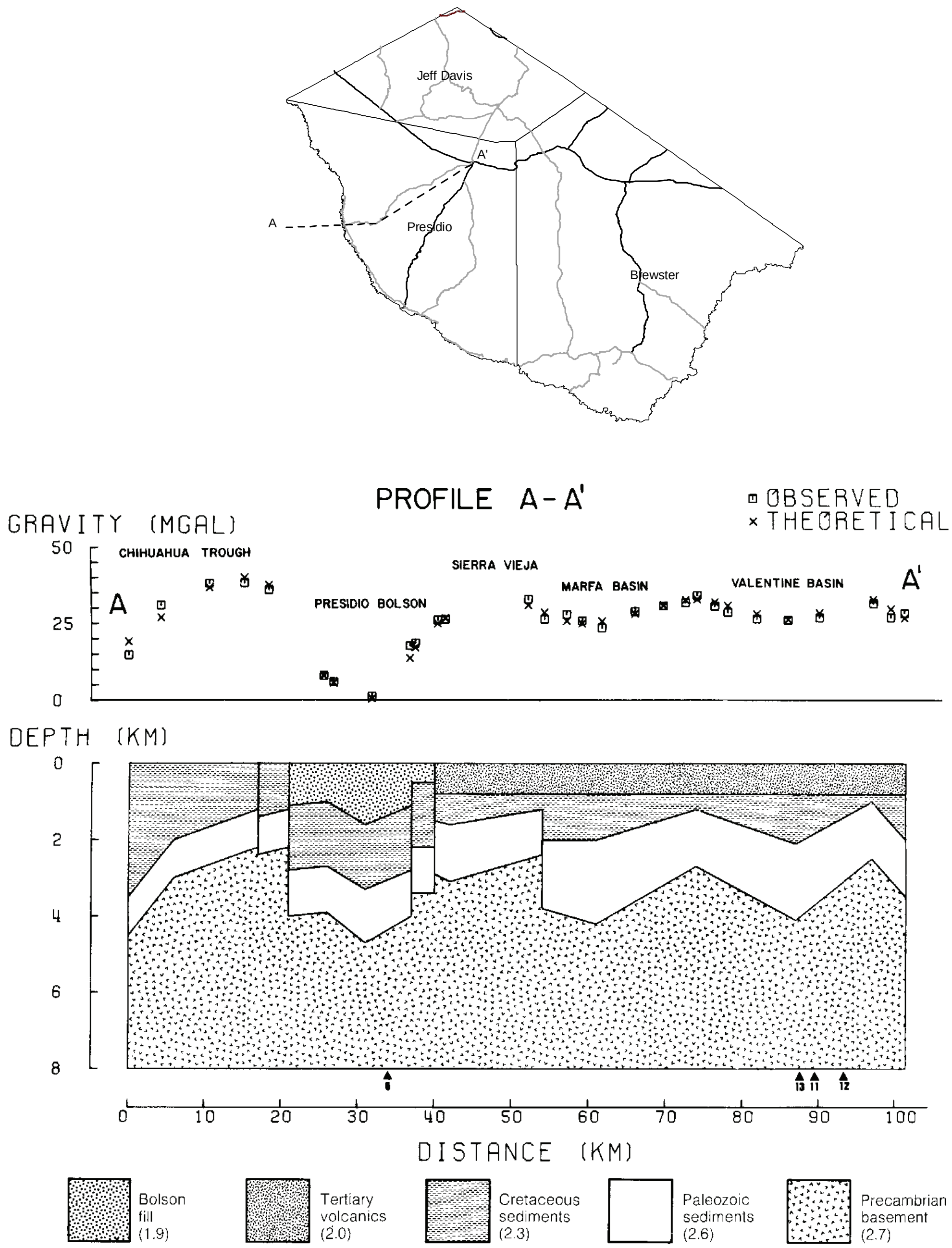


Figure 13-8: Gross estimate of the geological structure to a depth of 8 km (26,000 ft) (from Mraz and Keller, 1980).Groundwater Flow Paths

Table 13-2: Subbasin watershed areas for Brewster, Jeff Davis, and Presidio Counties.

| County Basin | Subbasin | Area ($mi^2$) |
|---|---|---|
| Brewster | Alpine | 273 |
| | | |
| Jeff Davis (Hart 1992) | Toyahvale | 299 |
| | Limpia | 351 |
| | Marfa | 153 |
| | Valentine/Ryan Flat | 227 |
| | Michigan Flat | 206 |
| | Kent | 287 |
| | | |
| Presidio | Alamito Creek | To be determined |
| | Valentine/Ryan Flat | To be determined |
| | | |
| | | |

The groundwater basin areas within Jeff Davis County were delineated by Hart (1992). The groundwater basin areas for Brewster and Presidio Counties were being analyzed at the time of publication. Current data is outlined in Table 13-2. These basins will be reevaluated after all the new water-level data are collated and organized.

All the streams in this area are losing streams and provide a recharge mechanism for the aquifers at certain locations. Many of the local landowners have reported that there are recharge zones on their properties. However, no organized delineation of these sites has yet been conducted.

## Springs

There are many springs in the area, and these have been mapped approximately to determine which formations provide most of the spring flow in the region. Hart (1994) examined the springs in Jeff Davis County. An initial review by the author of the known springs as determined by the U.S. Geological Survey (from within the Igneous aquifers) in Presidio County suggests that the units listed in Table 13-3 are the most likely to produce spring flow.

Some of the discharge rates have been estimated in the area around the Davis Mountains. The rates of flow vary from 0.5 gallons per minute (gpm) to 200 gpm. The total discharge from springs around the Davis Mountains is estimated to be 1.1 million gallons per day (Hart, 1992).

Table13-3. Summary of Spring data from study area in Jeff Davis, Brewster and Presidio Counties.

| Jeff Davis County (Hart 1992) | | Presidio and Brewster County (Chastain-Howley 2001) | |
|---|---|---|---|
| Geological Unit | No. of Springs | Geological Unit | No. of Springs |
| Intrusives | 7 | Intrusives | 19 |
| Merrill | 1 | Merrill | 20 |
| Wild Cherry | 3 | Petan | 19 |
| Mount Locke | 4 | Tascotal | 12 |
| Barrel Springs | 18 | Perdiz | 10 |
| Sheep Pasture | 3 | Chinati | 24 |
| Sleeping Lion | 2 | Rawls | 44 |
| Frazier Canyon | 13 | Alluvium | 29 |
| Adobe Canyon | 7 | Cottonwood Springs | 33 |
| Limpia | | Potato Hill | 13 |
| Gomez | 7 | Sheep Canyon | 16 |
| Star Mountain | 25 | Duff | 6 |
| Huelster | 47 | Crossen Trachyte | 4 |
| Alluvium | 4 | | |

The Presidio and Brewster County springs appear to be more evenly distributed across the different igneous extrusive geological units compared with springs within Jeff Davis County, where the Huelster, Star Mountain, and Barrel Springs account for the majority of flow. However, the majority of springs do appear to originate within the igneous intrusive and extrusive units rather than the alluvials.

## Igneous Aquifer Water-Well Locations

The groundwater systems in this area are poorly understood. Records from wells drilled are sparse, and water-level records are not common. Basic well data include location and depths of wells and are recorded in the TWDB Water Well Database. The extent of this database after additions from the ongoing Igneous aquifers project are shown in figure 13-9. These data will be used as a baseline for further analysis of the flow and near-surface storage in future years. Further data will be available after completion of this project in October 2001.

## Current known water use / Historical water use

The Far West Texas Regional Water Planning Group (which was set up in response to the requirements presented by Senate Bill 1 from the 1997 Texas Legislature) created a basic analysis of the current and projected use of each of the major water-user groups within the three counties. The data from that study is shown in table 13-4.

Table 13-4: Water use by county estimated for 2000 and 2050 (all values in acre-ft/yr).

| County | Location | 1996 | 2000 | 2050 |
|---|---|---|---|---|
| | | | | |
| Brewster | Alpine | 1,147 | 1,524 | 2,461 |
| | County Other | 2,427 | 2,895 | 3,611 |
| Jeff Davis | Fort Davis | 216 | 236 | 225 |
| | County Other | 870 | 3,928 | 3,611 |
| Presidio | Marfa | 722 | 977 | 1,189 |
| | Presidio | 646 | 768 | 1,652 |
| | County Other* | 23,924 | 26,451 | 24,102 |
| | | | | |
| **Total** | | **29,952** | **36,779** | **36,851** |

[Note * Presidio "County Other" includes up to 17,000 acre-ft/yr from Rio Grande surface-water allocations.]

The majority of the increase in projected water use is in Presidio and Alpine. Therefore, these are the locations within the counties that will probably need to be looked at in the greatest amount of detail.

## Recharge

Rainfall in the area varies from an average of 18.5 inches at Mount Locke to 11 inches at Kent. Most of the precipitation comes in the form of thunderstorms, which have their greatest frequencies between June and September. Most of the actual recharge will probably occur by direct infiltration through fractures within the rocks. This is also the case at the locations where the streams lose water to the aquifers. The amount of recharge is still a matter of great discussion. Recharge estimates range from a few thousand acre-feet per year to over 200,000 acre-ft/yr. The recharge is most likely to be somewhere between 50,000 and 100,000 acre-ft/yr over the Igneous aquifers area. However, further research is needed to verify these estimates.

## Storage

Calculations by LBG-Guyton for the Far West Texas Regional Water Plan (2001) suggest that there were a total of 9 million acre-ft of recoverable water in the tri-county area from the Igneous aquifers. Of this, 3.1 million acre-ft of recoverable groundwater was estimated to be in Brewster County, 1.3 million acre-ft is estimated to be in Jeff Davis County, and 4.6 million is estimated to be in Presidio County. This has to be a large underestimate because it was calculated from data from the Texas Water Development Board, which suggest that the extent of the igneous aquifers is only 785 $mi^2$. The boundaries as defined by this study suggest an area of approximately 5,000 $mi^2$.

The storage calculation determined for the Regional Water Plan used gross assumptions for the aquifer characteristics, and further study is urgently needed to better define the systems within this area. The scarcity of data makes the assumption of storage defined here highly uncertain.

The Igneous aquifers appear to be highly compartmentalized, and the units are often not laterally continuous. Therefore, this storage would only be available over a very large area, and major withdrawals would probably be cost-prohibitive.

## Water Quality

The overall quality of the water within the Igneous aquifers is excellent. The range of total dissolved solids (TDS) concentrations varies between 200 and 700 mg/L. These are all under the maximum concentration level for drinking water of 1,000 mg/L.

## Availability

Current supply is able to meet demand quite easily in this region of Far West Texas (FWTRWPG, 2001). Recharge is probably greater than the overall withdrawals, so the existing system appears to be sustainable on a regional scale, as well as through the 2050 planning period.

There is a large amount of water in storage in this area, and this is the subject of great discussion regarding export of this water to population centers in need, such as El Paso. The exportation of large amounts of water will probably mean that the overall aquifer units are locally being mined. However, the complexities of the Igneous aquifers may provide areas where recharge will be concentrated enough to allow larger sustainable withdrawals and where pumping will not significantly affect streamflow. However, further study would be necessary to determine the validity of this statement.

## Modeling

Conceptual and numerical modeling of the Igneous aquifers is very complicated and can currently only realistically be conducted on a gross, regional-conceptual scale. Numerical models of the Igneous aquifers have not yet been completed. Current water-level and pumping data within the area are very poor, and therefore it will be very difficult to build meaningful models with existing records. The only areas with enough data to consider modeling at this time would be the Ryan Flat area and possibly the Sunny Glen well field that feeds the City of Alpine.

The conceptual (geological and stratigraphical) models need to be created first before any regional modeling should be attempted. The models can be run, but the data are so sparse that any results may be difficult to prove.

Ongoing research will act as the first stepping stone to creating a usable database with which to determine groundwater availability through numerical modeling.

# Further Study and Ongoing Research

Test wells need to penetrate the entire volcanic sequence to adequately evaluate the potential of the full igneous sequence.

As we have seen in Sunny Glen, given the right stress regime (Morin and Savage, in prep.), production can get markedly better from deeper zones within igneous aquifers. Deepening the Roberts No. 3 and the Gardner wells within the Sunny Glen well field (which had been contributing almost nothing to Alpine's supply) resulted in the wells reportedly becoming the principal sources of water in Sunny Glen. The Lewis well found a small flow in the main trachyte aquifer but was deepened into the underlying basalt where a strong (350–500 gpm) flow was encountered. This well is currently not connected to the city supply because of the improved production from the other wells. Most water wells stop at the first water sufficient for the user's purpose (municipal wells may go past weak flows; private wells usually do not). Further exploration in this area would enhance the conceptual and stratigraphic knowledge to aid with studies in other igneous aquifer units in this region.

Initial analysis of well cuttings from the Killam oil-test well suggests that examination of the cuttings from the deep oil tests may be valuable in determining the stratigraphic and hydrogeological variations in the Igneous aquifers.

The University of Texas at El Paso (UTEP) in conjunction with WPR Consulting is currently conducting research of the area using geophysical methods to help determine the structure and extent of the Igneous aquifers. Data of satellite imagery and gravity and magnetics profiles should be available for interpretation and discussion in fall 2001.

There are many other areas of study that could be addressed in this section. The Igneous aquifers are the least-studied aquifers in this region and even just the collection of basic hydrogeological data such as water-level and production data will be valuable. In my opinion, these aquifers have the most potential to gain from more in-depth studies than any of the other aquifers in Far West Texas.

# Acknowledgments

Many thanks to John Olson for his valuable insight into the igneous geology of the area and for his review of this document. Thanks also to Jon Shub for his efforts in conducting fieldwork and GIS interpretations concurrently with this paper and for relaying his specific hydrogeological field observations.

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# Chapter 14

# Hydrogeology of the Marathon Basin Brewster County, Texas

Richard Smith[1]

## Introduction

The Marathon Basin lies in the northeastern part of Brewster County in western Texas (fig. 14-1). The region, referred to as Trans-Pecos Texas in the literature, is in the westward-projecting part of the state that lies along the Rio Grande west of the Pecos River. Physiographically, the region is closer to Mexico and New Mexico than to the rest of Texas. It is a region of high plateaus, broad cuestas, rugged mountains, and gently sloping intermontane plains. Very little vegetation is present except in sheltered valleys and on the higher summits. This factor allows an uncluttered view of the bedrock geology.

Ephemeral streams, which are little more than dry gravel beds for the vast majority of the year, gather runoff from the mountains and flow across the plains during the summer rainy season. These include Maravillas Creek, San Francisco Creek, Dugout Creek, Pena Blanca Creek, Pena Colorada Creek, and Woods Hollow Creek, to name a few. Several springs that flowed in historical times have now ceased and very little year round surface water is present in the area. The basin includes about 760 $mi^2$ centered on the town of Marathon. The basin is bounded on the north and west by the Glass Mountains and Del Norte Mountains, respectively. On the east, the boundary is recognized at Lemons Gap between Spencer and Housetop Mountains (fig. 14-2). The southern extent of the basin basically ends at Maravillas Gap, where Maravillas Creek cuts through the southwest end of the Dagger Flat anticlinorium.

## The Influence of Water on Area History

Water resources, particularly groundwater in the form of springs, have guided the history of the area. Most of Brewster County and specifically the Marathon Basin drain into the Rio Grande, although the northern part drains into the Pecos River. Soils are generally shallow and stony, with some loamy to sandy soils and clayey subsoils. Less than 1 percent of the land in the county is considered prime farmland. Vegetation at lower elevations in the county is drought resistant. Sparse grasses; desert shrubs such as ocotillo, lechuguilla, sotol, acacias, tarbrush, and creosote bush; some mesquite; and

[1] Texas Water Development Board

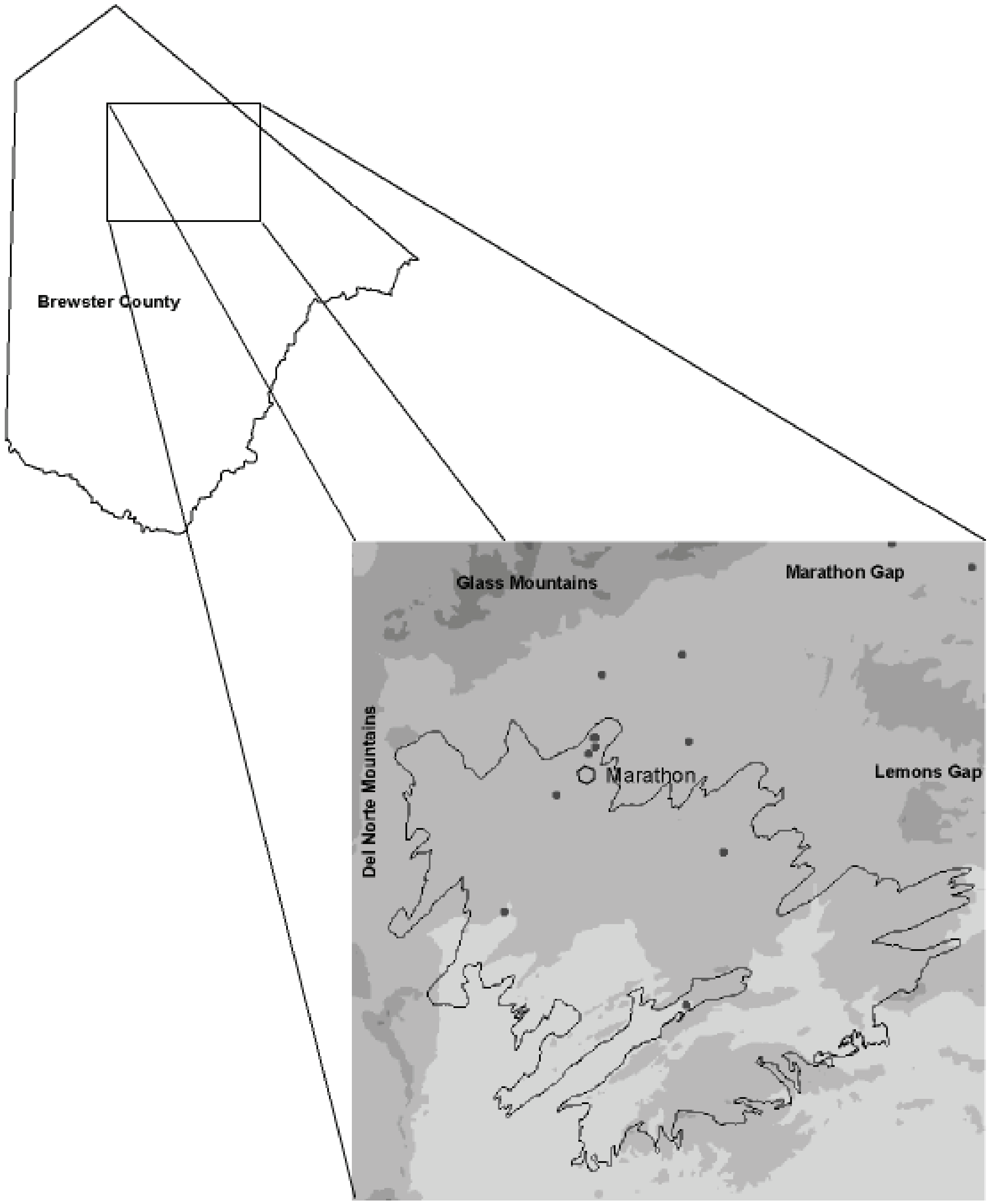


Figure 14-1: Location map of the Marathon Basin showing the Marathon Limestone aquifer outline. Additional points indicate wells within the basin that produce from Quaternary alluvium and terrace gravel.

Figure 14-2: The entrance to the Marathon Basin on Hwy 90 at Lemons Gap. This photo, taken looking to the west, shows a few of the low hills that are surficial remnants of Paleozoic structures once buried beneath Cretaceous sediments.

cactus are the dominant plants of this zone. At intermediate elevations vast grasslands occur in mountain basins, with white oak, juniper, and piñon woodlands dominating the slopes. Douglas fir, aspen, Arizona cypress, maple, Arizona pine, oaks, and madrone are found at the higher elevations. The fauna in Brewster County includes the pronghorn antelope, mule deer, white-tail deer, bobcat, mountain lion, desert bighorn sheep, black bear, coyote, raccoon, badger, prairie dog, pack rat, kangaroo rat, skunk, ringtail cat, porcupine, jackrabbit, cottontail, golden eagle, roadrunner, quail, dove, rock wren, white-winged dove, mourning dove, Canyon Wren, painted bunting, zone-tailed black hawk, and Colima warbler. Mineral resources include mercury, silver, lead, fluorspar, nonceramic clay, and lignite coal. Of these, the most important to the historical development of Brewster County was mercury. For most of the first half of the twentieth century the Terlingua Mining District in southern Brewster County was among the nation's leading producers.

The first European to set foot in what is now Brewster County may have been Álvar Núñez Cabeza de Vaca in 1535 (Casey, 1972). The presence in August 1583 of Antonio de Espejo's expedition, which probably passed the future site of Alpine via the Marathon Basin en route to La Junta de los Ríos, is more certain (Casey, 1972). Juan Domínguez de Mendoza is thought to have camped at Kokernot Spring, just northeast of Alpine, in 1684. But there was no extensive European presence in the Big Bend until the middle of the eighteenth century, when the Spanish began to explore the area in an effort to combat Indian raids into Mexico from the north (Gomez, 1990). In 1747, Governor Pedro de Rábago y Terán of Coahuila led an expedition into the Chisos Mountains, and, in 1772, Lt. Col. Hugo Oconór led an expedition to locate sites for forts along the Rio Grande. Oconór placed Capt. Francisco Martínez in command of the presidio at San Vicente, on the Comanche Trail on the Mexican side of the river. This was the first permanent European presence in the region (Gomez, 1990).

For much of the nineteenth century, the presence of Comanche raiding parties on their way to and from Mexico, combined with the forbidding local topography, discouraged European exploration of the Big Bend. The first Mexican and American explorers of the area, who arrived after the Mexican War, found harsh country indeed. In the summer of 1859 a camel expedition under 2d Lt. Edward L. Hartz set out from Fort Davis to explore the Comanche Trail and recommend a possible site for a fort on the Mexican border to protect against Indian raids. Hartz went south through the Marathon Basin to Persimmon Gap and down Tornillo Creek to the Rio Grande. A year later, a second camel expedition under 2d Lt. William Echols also explored along the Rio Grande, with the same goal as the Hartz expedition's. Before a fort could be built, however, the outbreak of the Civil War put an end to the plans.

After the war, three interrelated factors led to white settlement of what later became Brewster County: the presence of the United States Army, the development of the cattle industry, and the arrival of the railroad, all of which happened more or less simultaneously. Taking advantage of the Civil War, Indian cattle-rustling raids via the Comanche Trail rose sharply during the early 1860's and greatly reduced the number of cattle in northern Mexico. The high prices consequently paid by Mexican ranchers for imported cattle convinced Central Texas cattlemen to chance the long drive across the Big Bend country.

The revival of trade between Texas and Mexico along what has been called the Chihuahua Trail brought freighters and other transients to the future Brewster County. Kokernot Spring, where Mendoza had camped 2 centuries earlier, became a principal stopping place on the trail, renamed Burgess Waterhole after pioneering freighter John D. Burgess, whose wagon train was attacked by Indians there. In response to such threats, officials at Fort Davis established Camp Peña Colorada a few miles south of the future site of Marathon in 1879 (fig. 14-3).

Burgess and other freighters such as August Santleben helped spread the word about the open rangeland available in the Big Bend, and in the 1870's many ranchers from other parts of the state made plans to come west and investigate the area. Among them was Beverly Greenwood, from the Del Rio area, who came in 1878 and spent several months


Figure 14-3: Peña Colorada Springs, an oasis in the desert, was the site of Fort Peña Colorada during the mid- to late nineteenth century. Indians used the springs for thousands of years prior to European exploration. The springs flow from gravel deposits, and the water comes to the surface where Peña Colorada Creek crosses the very hard Caballos Novaculite. Flow has been measured at 150 to 450 gpm (DeCook, 1961).

exploring northern Brewster County. Mayer and Solomon Halff were San Antonio merchants who leased to the government the land on which Camp Peña Colorada was located. Later they became the first men to ship cattle into what is now Brewster County, along with John Beckwith, who in 1879 drove a herd of cattle to the vicinity of Peña Colorado Springs. These men contracted to supply meat to Camp Peña Colorada.

The burgeoning cattle industry got a major boost in 1882 when the Galveston, Harrisburg, and San Antonio Railway was built through the area. The gradual influx of cattlemen suddenly became a veritable flood, as a number of surveyors who had come with the railroad and the Texas Rangers who had been assigned to protect them, elected to stay. Among them were such men as Alfred S. Gage, James B. Gillett, and Joseph D. Jackson, who soon became the leading citizens of Brewster County. Initially, at least, ranchers generally settled in the northern part of what is now Brewster County for ease of shipping their cattle via the railroad. However, the Gage Ranch and the G4 Ranch, which started in the mid- to early 1880's, were the first major cattle operations in what is now southern Brewster County. Gage soon moved north to be nearer the railhead. Several towns sprang up along the rails, the most significant of which were Alpine, then called Murphyville, and Marathon.

These two towns quickly became shipping points and important supply centers for the booming cattle industry. Five years after the coming of the railroad, in 1887, Brewster County was marked off from Presidio County, as were Jeff Davis, Buchel, and Foley Counties. Brewster County was named for Henry P. Brewster, Secretary of War under David G. Burnet. Buchel and Foley Counties were not organized and were attached to Brewster County for judicial purposes. The first Brewster County elections were held on February 4, 1887, when Murphyville was selected as county seat; on March 14 of that year a contract was let for the construction of the Brewster County courthouse and jail. In 1890 Brewster County had just 710 residents, while Buchel and Foley Counties had only 298 and 25 residents, respectively. By 1897 Buchel and Foley Counties had still not been organized, and in that year their territory was officially added to that of Brewster County, making the latter the largest county in Texas.

Cattle ranching and mining have never regained the prominence in Brewster County that they had in the late nineteenth and early twentieth centuries. The county population rose from 6,478 in 1940 to 7,309 in 1950; dropped to 6,434 in 1960; and climbed again, to an all-time high of 7,780, in 1970, before declining slightly to 7,573 in 1980. The number of people employed in agriculture, however, steadily declined, from 712 in 1930 to 507 in 1950 and only 202 in 1970. Similarly, the number of people employed in mining dropped from 206 in 1930 to 147 in 1940 and 11 in 1950. In subsequent years, when the mercury mines enjoyed a brief renaissance, that figure rose again, to 32 in 1970 and 80 in 1980.

In the early 1980's Brewster County was 53rd among United States counties in land area and one of the most sparsely populated in Texas. The largest ancestry groups were Hispanic and English, both at 43 percent. In 1990 the population was 8,681. The largest town, Alpine, had 5,637 residents. By 1999, the population of Brewster County had increased to 8,793, with most of that growth in Alpine. Given the dry climate, coupled with the magnificent scenery, the Brewster County economy has become increasingly dependent on tourism.

# Physiography of the Basin

The Marathon Basin is in the Mexican Highlands physiographic province. The land surface consists of high plateaus, rugged peaks and sierras, and broad, shallow intermontane valleys. The Marathon area is situated on a structural uplift of the Ouachita fold belt. The crest has been eroded to a lower level than the flanks so that the central part is an irregular, circular basin surrounded by steep escarpments. The north and west basin margins formed by the Glass and Del Norte Mountains, respectively, consist chiefly of Permian and Cretaceous rocks that dip gently northward and westward. The relief on the

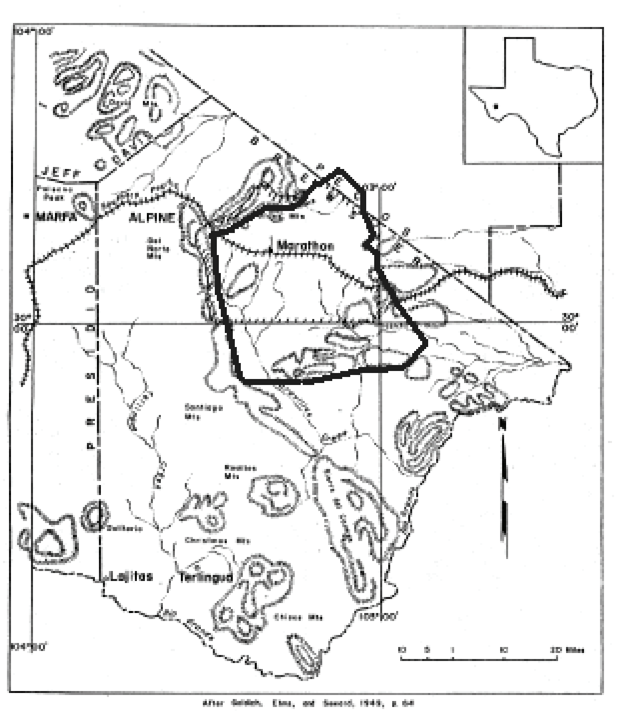


Figure 14-4: Location map of the Marathon Basin showing physiographic features and basin boundary.

rather steep inward-facing escarpments is about 1,000 to 1,500 ft. The highest peak in these ranges has an altitude of slightly more than 6,000 ft MSL (fig. 14-4).

The basin is composed of a series of shallow valleys and comparatively flat erosional surfaces separated by northeastward-trending, low, abrupt ridges. The valley floor in the basin varies in altitude from about 3,500 ft in the south to a high of 4,500 ft in the north along the base of the Glass Mountains. Relief within the basin is not great, the summits of the ridges being generally only about 300 to 700 ft higher than the adjacent valleys. The highest peak within the basin is Horse Mountain (southeast section) at 5,010 ft MSL.

# Climate

The mean annual temperature at Marathon is about 62° Fahrenheit (°F). The observed extremes vary from a high of about 110°F in the summer months to a low of below 0 in the winter. The mean monthly temperature varies from less than 46°F in December to more than 75°F in July and August.

According to records of the U.S. Weather Service, the long-term annual precipitation at Marathon is 13.59 inches. Most precipitation occurs during the summer months (fig. 14-5), largely in torrential rainstorms of irregular areal distribution.

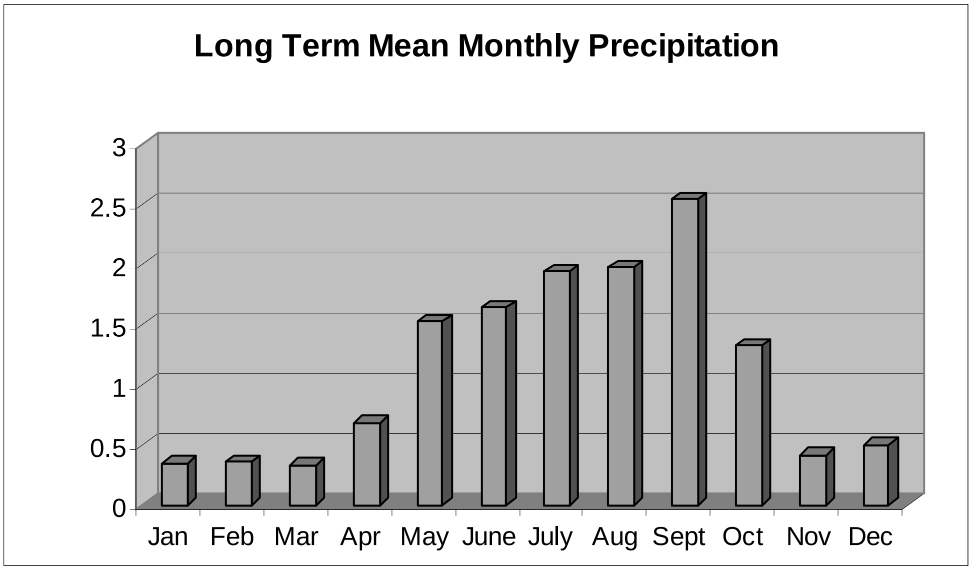


Figure 14-5: Annual precipitation distribution at Marathon, Brewster County, Texas 1946–1997. The value of the y-axis is inches.

The hot summers and low humidity contribute to high evaporation rates in the basin. The nearest long-term evaporation station is located at Balmorhea, 60 mi northwest of Marathon. Pan evaporation has been measured at about 70 inches per year, or over four times the mean annual precipitation at Marathon.

# Marathon Basin Drainages

The largest part of the Marathon area drains to the south owing to a southward slope. Maravillas, San Francisco, and some smaller creeks flow into the Rio Grande. The tributaries of Maravillas Creek include Dugout, Pena Colorada, Monument, Wood Hollow, and Hackberry Creeks. They drain the central and western sides of the basin. The eastern side of the basin is drained by San Francisco Creek, with its main tributary being Pena Blanca Creek. The northern extent of the basin is drained by Big Canyon, which flows generally eastward to the Pecos River.

These drainages and the unnamed tributaries to them are the principal recharge sources for the water-producing formations within the basin and the spring flow that issues from various points within the basin.

# General Geology of the Marathon Basin

The consolidated rocks exposed in the basin range in age from Paleozoic Cambrian to Cenozoic Tertiary. Intrusive igneous rocks occur in scattered areas, but extrusive rocks crop out only along the rim of the Del Norte Mountains to the west (King, 1937).

The Paleozoic rocks in the basin have a total thickness of about 21,000 ft. The majority of this thickness is composed of Pennsylvanian and Permian strata. Prior to the deposition of the Permian, the underlying rocks in the Marathon area were strongly folded and faulted, producing a series of northeastward-trending anticlinoria and synclinoria (King, 1937). Permian rocks, which are restricted to the south flank of the Glass Mountains and the northeast end of the Del Norte Mountains, lie unconformably on the folded and faulted surface of older rocks (Tauvers, 1988). The area was subjected to erosion prior to the deposition of the Cretaceous rocks, which resulted in the truncation of the folded and faulted rocks of the basement. Cretaceous sediments, measuring about 1,200 ft, were deposited on this erosional surface. The resulting Cretaceous surface was uplifted to form the Marathon Dome, which in turn was eroded to form a topographic basin (King, 1980). Finally, during the early part of the Tertiary period, igneous rocks intruded the Cretaceous strata (King, 1980).

Quaternary sediments mantle the present-day stream valleys and form alluvial-fan deposits at the base of the escarpments up to 125 ft thick. These units are generally very thin and serve as a catchment area for rainfall that then enhances recharge to the underlying Paleozoic rocks.

# Principal Water-Bearing Units of the Marathon Basin

## Marathon Limestone

The Marathon Limestone is the most productive aquifer in the Marathon area. DeCook (1961) noted that 92 wells, most in the town of Marathon, had yields from this formation that varied from a few gallons per minute (gpm) to over 300 gpm. The aquifer can best be characterized as a highly fractured limestone aquifer. Groundwater in the Marathon Limestone generally occurs under water-table conditions, but is locally (i.e., the City of Marathon) under artesian pressure and may rise a few feet above the point where it is first encountered.

The Marathon Limestone, which is Ordovician, crops out in the Marathon anticlinorium and at the northeast end of the Dagger Flat anticlinorium. The City of Marathon is situated at the north end of the Marathon anticlinorium, and wells in the city are completed in the Marathon Limestone. The thickness of the Marathon Limestone decreases from north to south, ranging from 800 to 900 ft at Marathon to about 350 ft at Dagger Flat. The Marathon Limestone is a dark-gray, flaggy limestone with gray to greenish clayey shale streaks. Sandstone and conglomerate are interbedded with the limestone and shale. The Marathon Limestone is underlain by the Cambrian Dagger Flat Sandstone, which does not produce water.

Approximately 100 domestic and industrial wells existed in the City of Marathon in 1956. A number of these wells were completed in shallow alluvial deposits, but the majority were drilled into the Marathon Limestone. Groundwater levels measured for the Marathon Limestone at the time indicated a high of 70 ft below the surface in the southwest part of Marathon compared with a low of 150 ft below the surface on the northeast side of town. There is not much basinwide information on the water levels in the Marathon Limestone. The plethora of wells in the city of Marathon has been replaced for the most part by two municipal wells operated by the Marathon Water Supply and Sewage Corporation.

In 1956, water levels on the north side of Marathon were 150 to 154 ft below surface level. In 1969, the City of Marathon well 52-55-104 was drilled to a depth of 468 ft and the water level was found to be 152 ft below the surface. An offset well drilled in 1974 (well 52-55-105) had a water level of 125 ft below the surface. These wells have measured yields of 85 and 55 gpm, respectively. According to a limited survey of the City of Marathon during July of 2001, it appeared that most houses and businesses were connected to the municipal supply.

## Alsate Shale

The Alsate Shale overlies the Marathon Limestone and crops out in the Marathon and Dagger Flat anticlinoria. The Alsate Shale, which consists of thin-bedded limestone; indurated, greenish shale; lenses of black chert; conglomerate; siltstone; and quartzose sandstone, is not known to yield water in the Marathon area.

## Fort Pena Formation

The Fort Pena Formation, which unconformably overlies the Alsate Shale in the Marathon area, forms low hogbacks roughly parallel to more prominent ridges formed by much younger novaculite. This formation is chiefly an alternating sequence of limestones and shales that yield water to some wells in the area. However, judging by the lithology, only small yields should be expected from the Fort Pena Formation.

## Woods Hollow Shale

The Woods Hollow Shale is poorly exposed in the Marathon area. It consists principally of light-gray-green to tan, slightly calcareous shale interbedded with laminated sandy limestone and fine-grained sandstone. It is assumed that wells are not completed in this formation.

## Maravillas Chert

The Maravillas Chert crops out primarily on the inner steep slopes of hogbacks formed by the overlying Caballos Novaculite. The Maravillas Chert is not known to contribute water to any wells in the area. However, it is a highly fractured formation and does convey water that emerges as springs along contacts with the underlying Woods Hollow Shale and the overlying Caballos Novaculite.

## Caballos Novaculite

The Marathon Limestone is basal Ordovician and the Maravillas Chert is considered the top of the Ordovician; whereas, the Caballos Novaculite is considered basal Devonian. The Caballos Novaculite is the principal ridge-forming formation in the Marathon area, making up the ridges or hogbacks that enclose the Marathon and Dagger Flat anticlinoria (fig. 14-4). The novaculite is not known to yield water to wells in the Marathon area, but springs issue from joints and fissures in the weathered parts of the formation.

## Pennsylvanian Formations

From oldest to youngest, the Tesnus, Dimple Limestone, Haymond, and Gaptank Formations are representative of the Pennsylvanian System in the Marathon Basin. All of these formations, with the exception of the Haymond, yield small quantities of water to stock wells in the area.

## Permian Formations

The Wolfcamp, Leonard, Word, and Capitan Formations are all Permian formations with no water production except for two wells in the Wolfcamp Formation and five wells in the Word Formation. These are all stock wells with very low yields.

The Cretaceous rocks that border the basin and the Tertiary rocks that are intrusives in the basin are not important sources of groundwater in the Marathon area.

# Groundwater Occurrence and Movement

The geologic structure in the Marathon Basin controls the occurrence, availability, and movement of all groundwater. The Marathon aquifer is composed of the Gaptank, Dimple, Tesnus, Caballos, Maravillas, Fort Pena, and Marathon Limestone Formations (Ashworth and Hopkins, 1995). The Marathon Limestone, which is the principal and most productive aquifer in the area, is found primarily in the Marathon and Dagger Flat anticlinoria, where upfolding has brought the formation to relatively shallow depths. In these areas, groundwater occurs under water-table conditions. In contrast, in the synclinorial belts where the Marathon Limestone is downfolded, younger rocks are generally tapped for groundwater. In those areas where the Marathon aquifer is overlain by relatively impermeable strata, groundwater is confined and is under artesian pressure.

Figure 14-1 shows the extent of the Marathon aquifer as delineated by Ashworth and Hopkins (1995). This delineation coincides with the surface outcrop of Ordovician rocks such as the Marathon Limestone, which is the primary groundwater-producing formation in the basin. However, numerous wells, primarily for stock, have been completed in the Quaternary alluvial and terrace sands, gravels, and silts that occur throughout the basin. The occurrence of these wells coincides with the entire basin configuration as seen in figure 14-4.

In general, groundwater in the Marathon Basin moves southward and southeastward toward the Rio Grande. This movement reflects the surface topography and the general drainage pattern of the area. Although data are not available to map the water table accurately, information gleaned from wells in the City of Marathon, in addition to spring information at Pena Colorada and other springs and spring-fed creeks, leads to the conclusion that subsurface flow is toward the south and southeast. This fact is probably not true in the immediate vicinity of the City of Marathon, where city wells have most likely drawn down water levels.

Groundwater is at relatively shallow depths in most of the Marathon Basin. Most wells are less than 250 ft deep. According to DeCook (1961), the depth to water for 205 wells in the area was less than 150 ft below the land surface, and in 72 wells it was less than 50 ft.

# Groundwater Recharge

Groundwater reservoirs in the Marathon area are recharged principally by infiltration of rainfall and stream runoff. Some underflow from outside the basin can be expected. The amount of recharge from precipitation is determined by the duration, intensity, and type of precipitation, the thickness of the vegetative cover, the porosity and permeability of the soil and underlying rocks, and the areal extent of the precipitation event.

A large part of the precipitation takes place in the summer months, when the evaporation rate is highest (fig. 14-5). Thus, only a small part of the precipitation escapes evaporation and becomes recharge. Runoff from the summer torrential downpours emerges from the steep slopes of the basin boundary areas, spreads out over the alluvial fans, and percolates into the coarse material forming the fans. If precipitation at Marathon is considered to be representative of the entire 760-$mi^2$ basin, the annual precipitation of nearly 13.6 inches is equal to about 550,000 acre-ft of water per year. Estimations of less than 5-percent recharge in the neighboring Alpine area (Littleton and Audsley, 1957) give rise to about 25,000 acre-ft of recharge in the Marathon area.

Recharge from underflow is only likely from the east, and any water entering the basin from this direction would most likely move southwestward, along San Francisco Creek.

# Groundwater Discharge

Groundwater is discharged from the basin via spring flow, evapotranspiration, underflow southward to the Rio Grande and pumpage from wells. Good estimates for spring flow and pumpage exist, but the other parameters can only be grossly estimated.

The amount of spring discharge was approximately 880 acre-ft in 1957 and 902 acre-ft in 1976. An undetermined, but probably large, part of the groundwater moves out of the Marathon Basin as underflow through the alluvium and permeable Paleozoic rocks. These preferential pathways include the stream valleys of Maravillas, Woods Hollow, Hackberry, and San Francisco Creeks, in addition to other minor drainages.

An additional quantity of water is removed via evapotranspiration. This varies with the season, with the summer months being the greatest with the highest temperatures. Discharge by direct evaporation occurs at several places along Pena Colorada Creek, Maravillas Creek, and other streams where the water table is at or near the land surface.

Groundwater is also discharged through pumpage of wells in the basin. In 1957, about 280 wells were known in the area. The total withdrawal was probably less than 400,00 gallons per day or 450 acre-ft per year. This amount is less than 2 percent of estimated natural recharge. Although the number of wells has been reduced owing to the Marathon Water Supply Corporation, the amount of pumpage has probably increased in the basin as a result of increased tourism and expanding development.

# Water Quality

The water quality of the groundwater in the Marathon Basin is generally good. Sampling of water from well 52-55-104 started in 1972, and samples have been taken and analyzed through 1998. Total dissolved solids (TDS) have averaged between 525 and 550 mg/L, chloride—115 mg/L, nitrate—11.5 to 12 mg/L, sulfate—84 mg/L, and sodium 72—mg/L. The TDS is a little high but still within acceptable limits (Ashworth and Nordstrom, 1989).

Some reports of contamination by oil, gas, and saline water have been noted, but it appears that these zones can be sealed off and uncontaminated water production achieved.

# Conclusions

The Marathon Basin contains two principal aquifers, the Marathon Limestone and the Quaternary alluvium near the streams and escarpments. The Marathon Limestone yields large quantities of good-quality water in the Marathon and Dagger Flat anticlinoria. In the synclinorial areas, production is generally from shallower producers.

Yields in the basin are low because most wells are for stock purposes. Larger yields could be achieved in numerous places through proper completion techniques.

Precipitation contributes about 25,000 acre-ft of recharge per year. Discharge is about 900 acre-ft/yr from springs, 450 acre-ft/yr from pumpage, and an unknown amount from underflow and evapotranspiration.

Water quality is generally good, although the water is hard. Some oil pollution exists and should be isolated.

# Recommendations

The Marathon Basin is a very undeveloped area of the State of Texas. Studies need to be conducted to expand databases throughout the basin. Well records, drillers' logs, geophysical logs, and field examinations of existing wells need to be gathered and analyzed. Water-level measurements need to be made throughout the basin. Pumping tests should be conducted to establish aquifer parameters.

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# Chapter 15

# Hydrogeology of the Rustler Aquifer, Trans-Pecos Texas

Radu Boghici[1] and Norman G. Van Broekhoven[2]

## Introduction

The Rustler aquifer is one of the less-studied aquifers in Texas, and this paper is an attempt to review and summarize all available hydrogeologic information on this aquifer. The Rustler Formation consists of up to 500 ft of carbonate and evaporite strata of Permian age deposited in the Delaware Basin of West Texas. The formation yields moderate to large quantities of fresh to brackish groundwater, primarily from solution openings in its upper section. Recharge takes place by cross-formational flow from deeper aquifers and percolation of surface water through the formation outcrop. Discharge is predominantly to pumping wells and by flow into overlying aquifers. Geochemical data indicate the main processes impacting the groundwater chemical composition are the dissolution of calcite, dolomite, gypsum, and halite and cation exchange.

## Regional Geologic Setting

The Rustler Formation underlies the Delaware Basin in West Texas and Southeastern New Mexico and is the youngest unit of the Late Permian Ochoan Series. The formation outcrops in a north-to-south-trending belt in the Rustler Hills of Culberson and Reeves Counties and the contiguous plains to the east, where it unconformably overlies the Salado Formation. The 250- to 670-ft-thick Rustler strata extend downdip toward the center of the Delaware Basin. In outcrop they consist of dolomite, dolomitic limestone, limestone breccia, gypsum, and mudstone, with minor siltstone and sandstone near the base (Hentz and others, 1989). Six subsurface formation members have been identified in the Rustler Hills area (table 15-1).

The Rustler Formation becomes thinner (40–200 ft) and conformably overlies the Salado Formation toward the eastern margin of the Delaware Basin and across the Central Basin Platform and Val Verde Basin (fig. 15-1). Hentz and others (1989) recognized three distinct subsurface members in Pecos County (table 15-2).

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Table 15-1: Subsurface stratigraphy of the Rustler Formation in Culberson and western Reeves Counties (from Hentz and others, 1989).

| Division | Thickness Range (ft) | Lithology |
|---|---|---|
| Forty-niner Member (Vine, 1963) | 70–90 | Two beds of white to gray massive and nodular anhydrite and gypsum separated by a thin gray to red gypsiferous mudstone or siltstone bed |
| Magenta Member (Adams, 1944) | 20–28 | Interbedded gray dolomite and gray gypsiferous dolomite |
| Tamarisk Member (Vine, 1963) | 105–125 | Two beds of white to gray massive and nodular anhydrite and gypsum separated by a gray gypsiferous mudstone bed |
| Culebra Member (Adams, 1944) | 50–55 | Grey laminated dolomite, locally brecciated |
| Lower gypsum and mudstone member | 35–50 | Gray and tan mudstone and gypsum interspersed with thin gypsiferous dolomite beds |
| Siltstone member | 95–145 | Gray and locally red dolomitic siltstone and mudstone in the lower part; gray dolomite at top |

Table 15-2: Subsurface stratigraphy of the Rustler Formation in Pecos County (descriptions from Hentz and others, 1989).

| Division | Thickness Range (ft) | Lithology |
|---|---|---|
| Upper Member | 10–55 | Gray dolomite, locally calcareous and oolitic. Vuggy porosity. |
| Middle Member | 40–65 | Calcareous siltstone, sandstone, red and gray shale, with interspersed anhydrite, gypsum, and shale, locally massive anhydrite and gypsum, and sandy dolomite. |
| Lower Member | 10–50 | Brownish-gray dolomite, locally calcareous, argillaceous, oolitic, or sandy. Vuggy porosity common. |

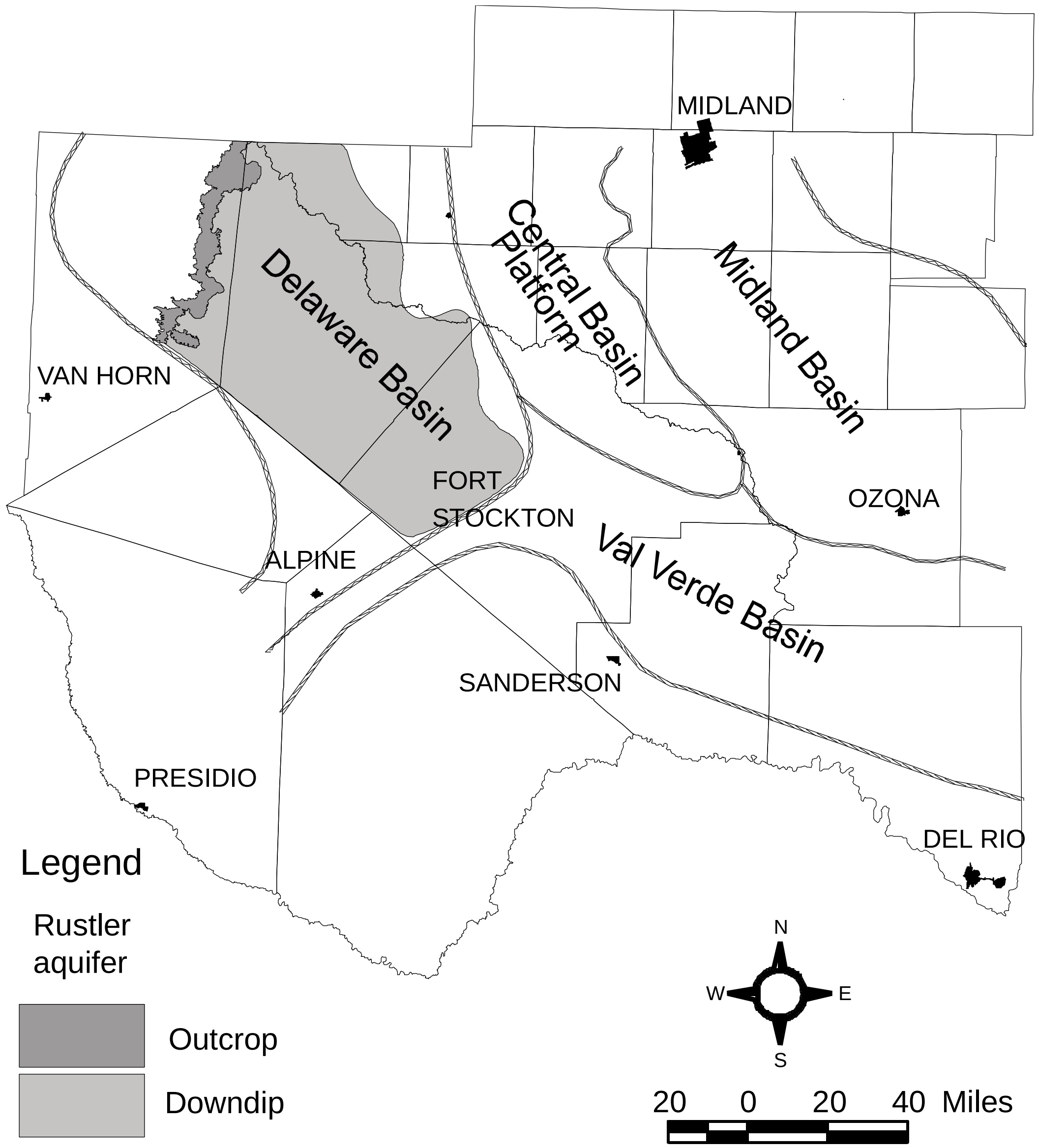


Figure 15-1: Regional Permian structure (modified from Small and Ozuna, 1993).

# Structure

The most prominent structural features in the Rustler region are the Delaware Basin, the Central Basin Platform, and the Val Verde Basin (fig. 15-1). The structure of the Rustler rocks in the study area closely reflects the structure of the older Permian strata. Generally Rustler beds dip away eastward along a wide, irregular monocline. Late Tertiary tectonic activity in the Basin and Range area left northeast-trending fault patterns that are visible today throughout the Trans-Pecos gypsum plain. An important effect of Tertiary faulting and eastward tilting of the Delaware Basin was the commencement of dissolution of Ochoan evaporites. This process caused the Permian beds to collapse and form a network of deep troughs and depressions throughout West Texas. Salt solution troughs have been mapped directly above the Capitan Reef on the eastern edge of the Delaware Basin, as well as under the west-central Delaware Basin (Hiss, 1976). Triassic, Cretaceous, Tertiary, and Quaternary rock and sediments gradually filled the troughs and underwent subsidence, faulting, and folding. Today they form prolific aquifers.

Structure for the top of the Rustler Formation (fig. 15-2) reveals several solution troughs resulting from the union of many lens-shaped subsurface depressions. The Balmorhea-Pecos-Loving Trough (Hiss, 1976) originates near Balmorhea and extends northward to Pecos and on into Eddy County, New Mexico. The Belding-San Simon Trough (Hiss, 1976) follows the Capitan Reef trend from Belding in Pecos County, Texas, to the San Simon swale in Lea County, New Mexico.

# Aquifer Delineation

In Texas, the Rustler aquifer underlies an area of approximately 480 $mi^2$ encompassing most of Reeves County and parts of Culberson, Pecos, Loving, and Ward Counties (fig. 15-3). The southwestern Rustler aquifer boundary was arbitrarily traced along the Jeff Davis-Reeves-Pecos County line because of lack of well coverage in that area. Although the Rustler Formation is present in Brewster and Jeff Davis Counties, no Rustler water-well data are currently available for these counties. The 5,000-mg/L total dissolved solids (TDS) isoline was designated as the downdip limit of the aquifer to the northeast and southeast (fig. 15-3).

# Aquifer Properties

Pump-test data for the Rustler aquifer were not available at the time this paper was written. Aquifer permeability is thought to be low except where porosity has been enhanced by carbonate and evaporite dissolution (Muller and Price, 1979). Reported well yields vary from 7 gallons per minute (gpm) to 4,400 gpm. Prior to 1955, when well acidizing became common in the area, few wells could produce from the Rustler in Reeves County (Ogilbee and others, 1962). The acidizing practice “almost eliminated dry holes” (Ogilbee and others, 1962) and resulted in temporarily increased yields of up to 1,000 gpm.

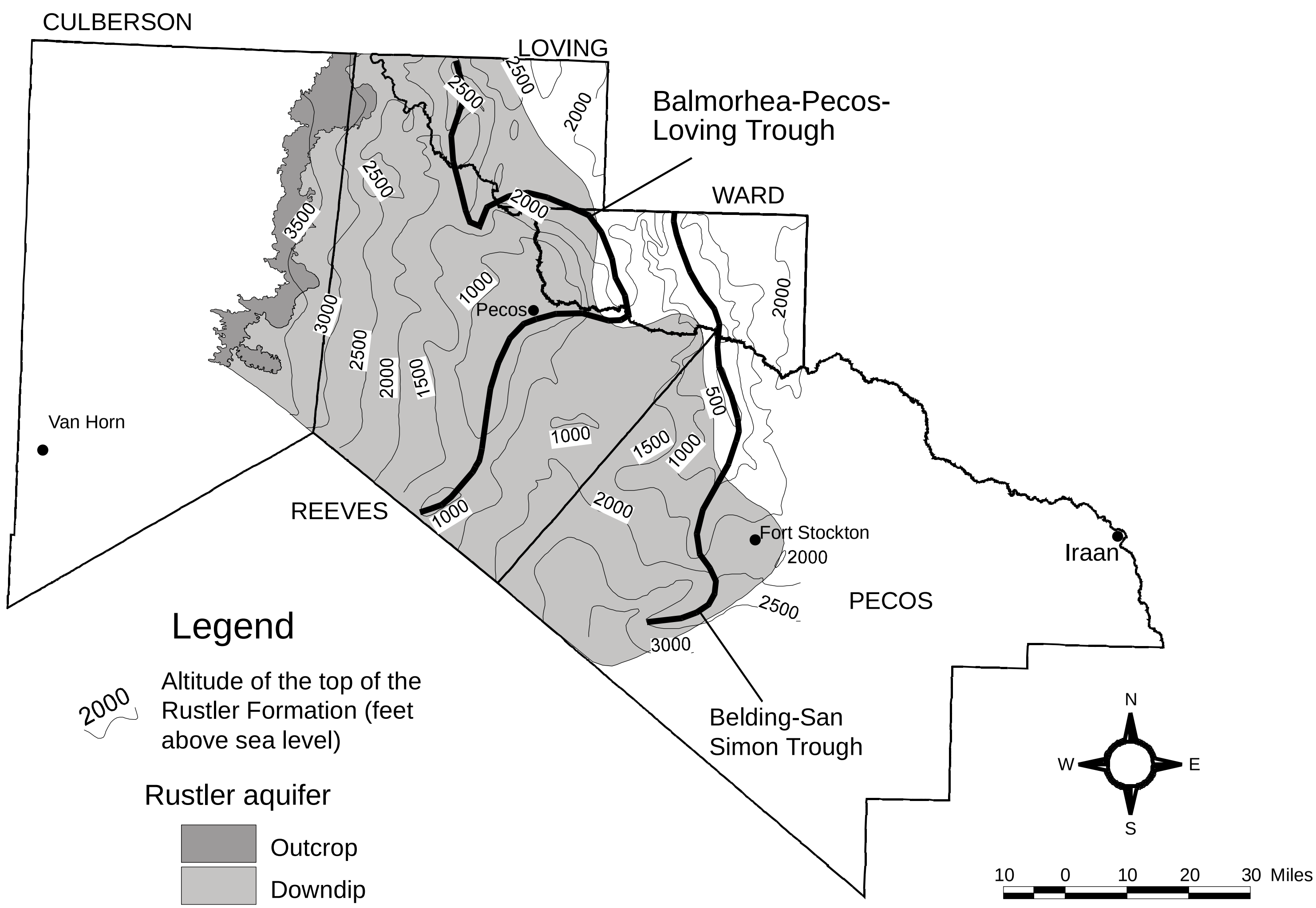


Figure 15-2. Structure map at the top of the Rustler Formation. Contours and modified trough locations from Hiss (1976).

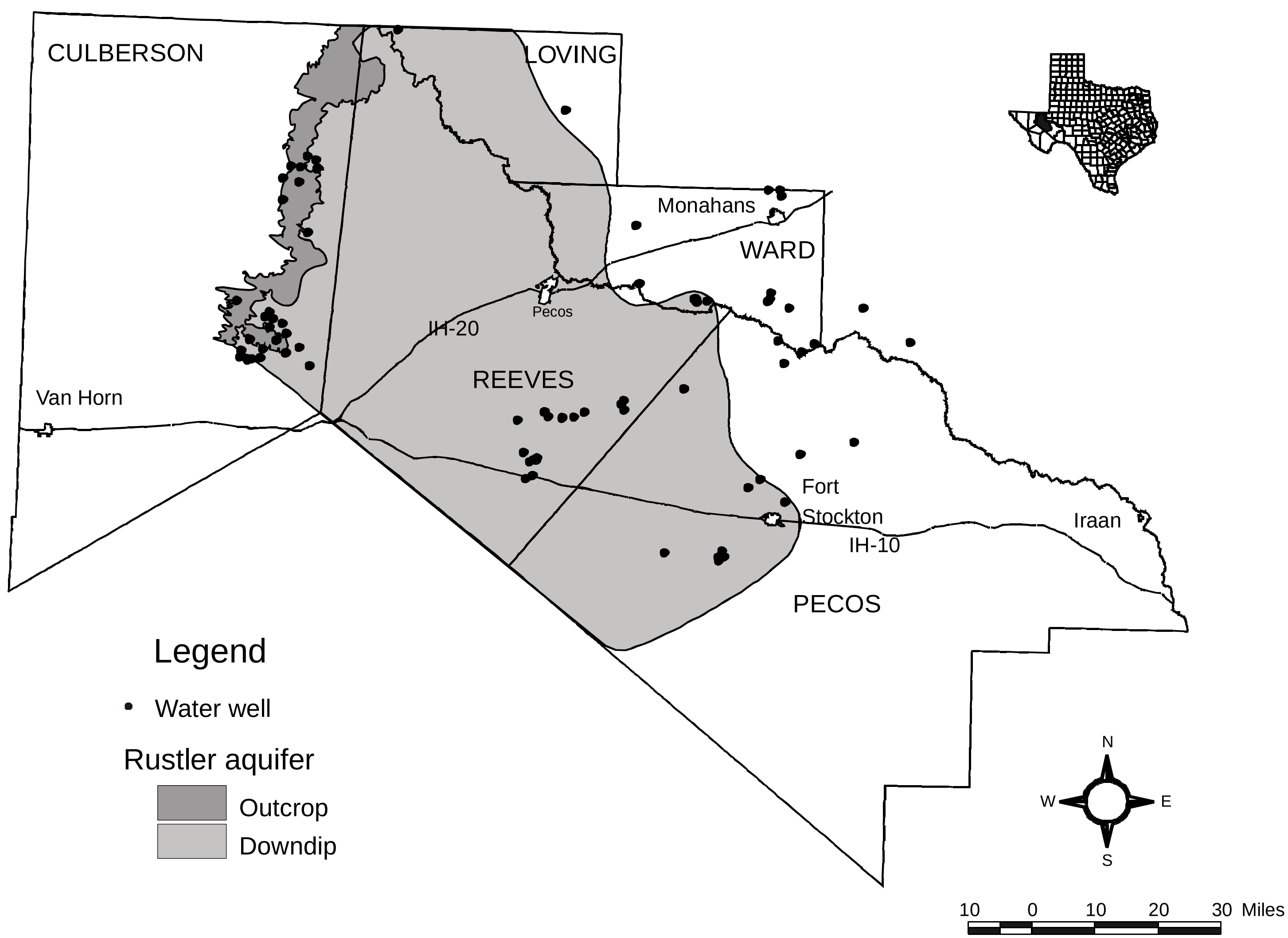


Figure 15-3. Location map and available Rustler wells. Map includes wells outside the aquifer limits. Data from TWDB.

The most productive interval of the Rustler Formation is the Upper Member (table 15-2), which, during the 1950’s, was supplying 500 to 1,000 gpm to 30 irrigation wells in Reeves County (Ogilbee and others, 1962). A well drawing 4,400 gpm was drilled in 1964 near Belding in Pecos County. Many of the wells in the Rustler flowed when drilled, and some are still flowing today, albeit at greatly reduced rates. For example, well 46-40-801 was flowing 0.25 gpm in 1995, down from 900 gpm when drilled in 1932. This flow diminution is probably due to a lowering of the water levels in parts of the aquifer. Well-construction problems, such as ruptured casings and caving of the formation below the casing, could also explain the reduction in flow (Armstrong and McMillion, 1961).

# Potentiometry

Water-level data from for the Rustler aquifer are very sparse. Historically no more than 13 wells have been measured throughout the aquifer in any given year. The areal distribution of the wells with water-level measurements made it impossible to contour an aquiferwide potentiometric surface map. From the outcrop of the Rustler aquifer in southeastern Culberson County, groundwater moved generally to the east-northeast toward the Reeves County line in 1988 (fig. 15-4). The average hydraulic gradient in the area was 0.015. Hydraulic heads in the outcrop ranged from 3,255 to 3,368 ft. A large data gap spans the area between the Culberson-Reeves County line and the Pecos meridian. In western Pecos County and southeastern Reeves County groundwater moved toward the north-northwest, with an average hydraulic gradient of 0.004. Hydraulic heads in this area ranged from 3,058 ft in the Belding area southwest of Fort Stockton to 2,612 ft near the Pecos-Reeves County line.

A water-level measurement from a Crane County well was used to constrain the downdip water-level configuration. Boghici (1997) delineated flow directions on the basis of earlier hydraulic heads outside the formal limits of the aquifer. The flowlines (fig. 15-5) show a centripetal pattern that converges under an area north of Fort Stockton. Trends in the potentiometric surface suggest the presence of a high-permeability area funneling groundwater flow in eastern Reeves County. In 1988, depths to groundwater in the Rustler outcrop were between 50 and 150 ft. The water levels were deeper in southeastern Reeves County, where they ranged from 130 to over 250 ft. Groundwater was 134 and 139 ft deep in two wells owned by the Belding Farms in Pecos County.

Time-series hydrographs of selected wells in the study area (fig. 15-6) illustrate long-term temporal fluctuations in aquifer storage. Depletion of storage due to pumping has occurred in Pecos County wells 51-16-608 and 51-16-609 from the mid-1960’s through late 1970’s. Declines in water levels of up to 100 ft have been recorded during this period.

Beginning in the 1980’s, reductions in groundwater withdrawals resulted in water-level recovery to levels above those encountered when these wells were drilled. The cessation of irrigation pumping in Reeves County well 46-60-902 has resulted in water levels rising 170 ft from 1960 to 1968.

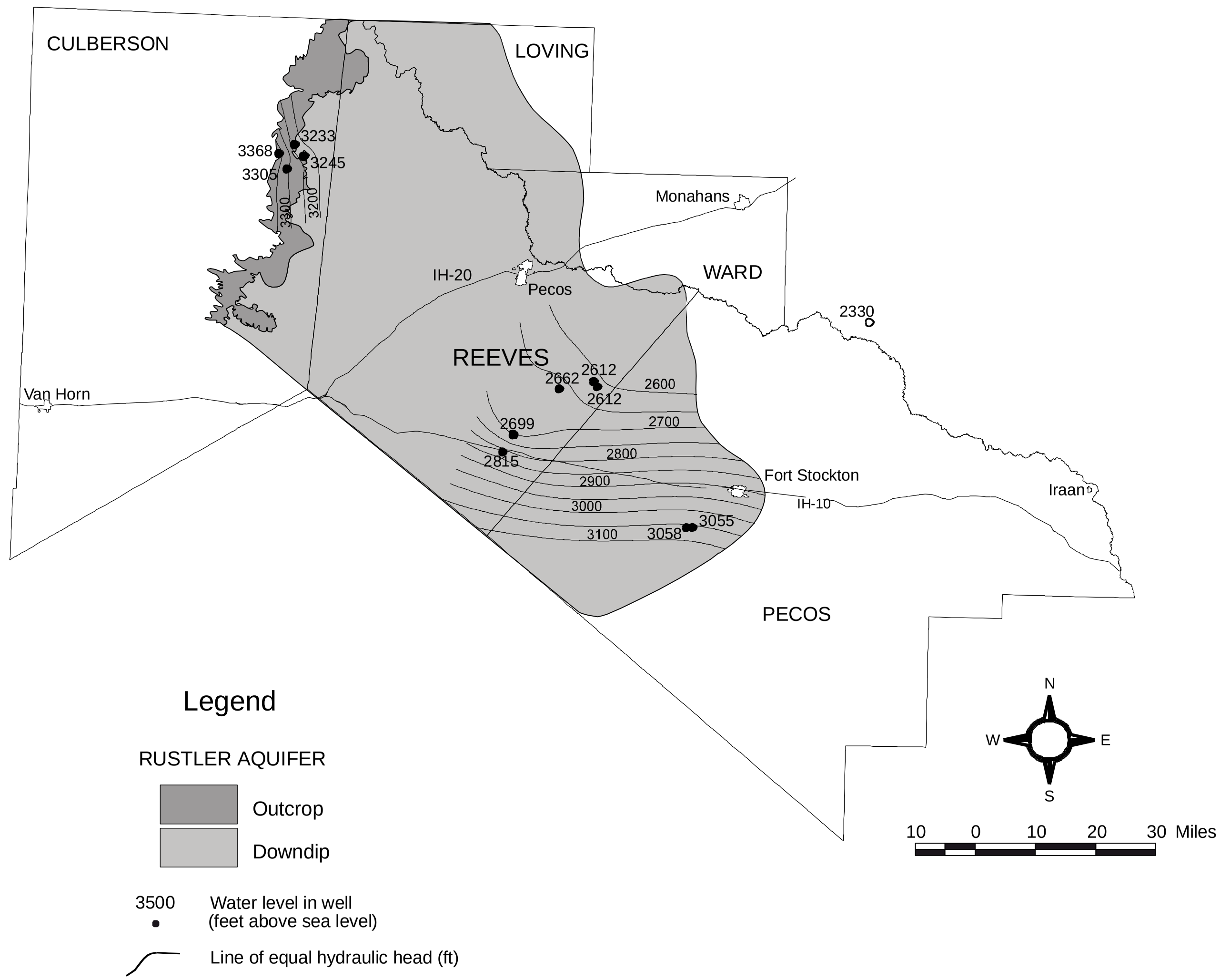


Figure 15-4. Potentiometric surface map for Rustler aquifer, 1988. Data from TWDB.

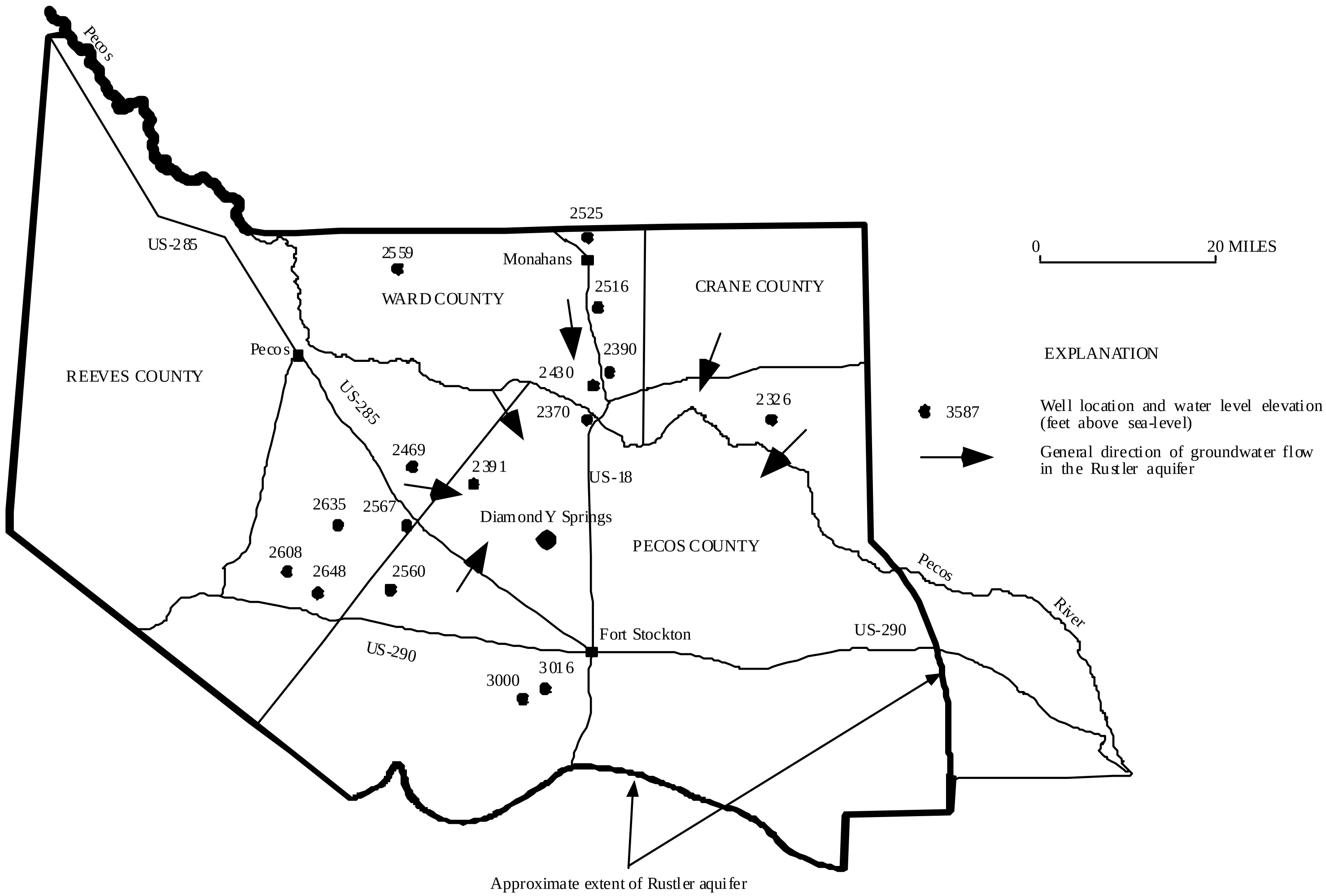


Figure 15-5. Selected wells, water levels and inferred flow directions in the Rustler aquifer. Data from Richey (1987).

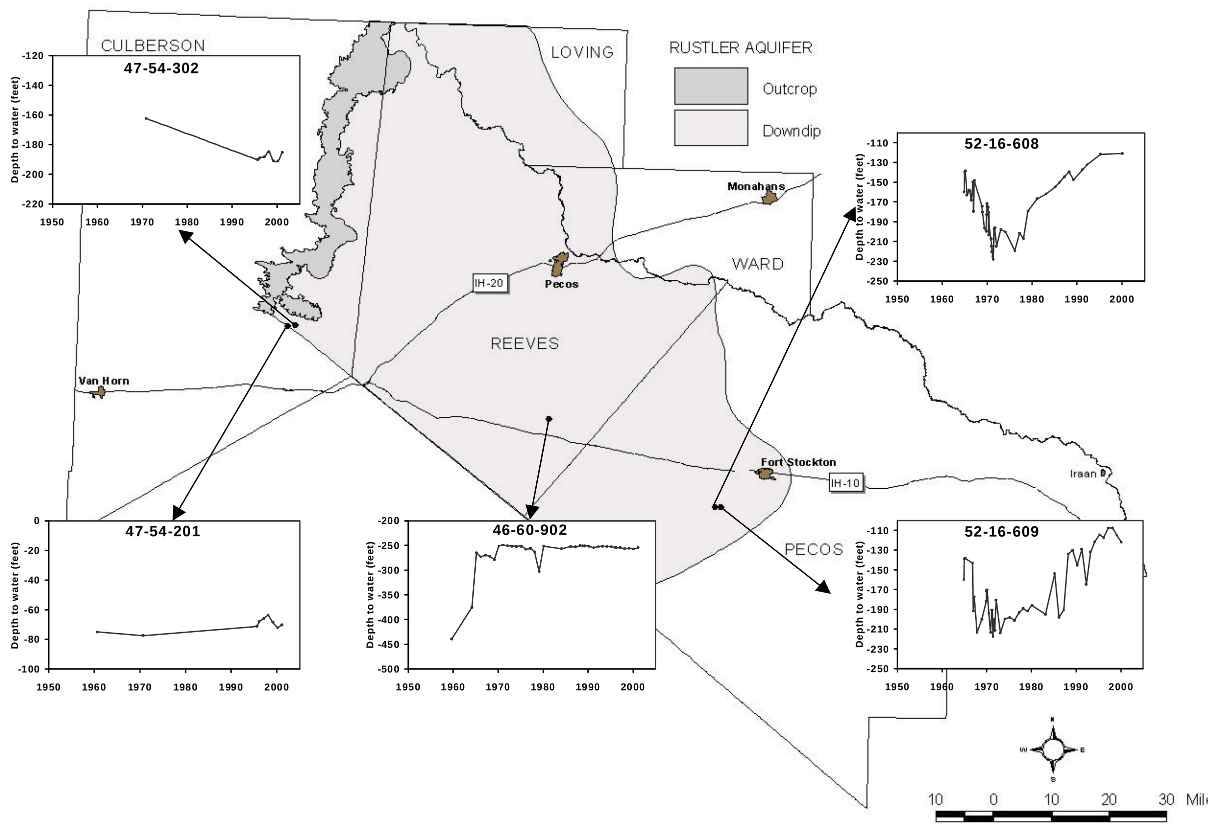


Figure 15-6. Hydrographs for selected wells in the Rustler aquifer. Data from TWDB.

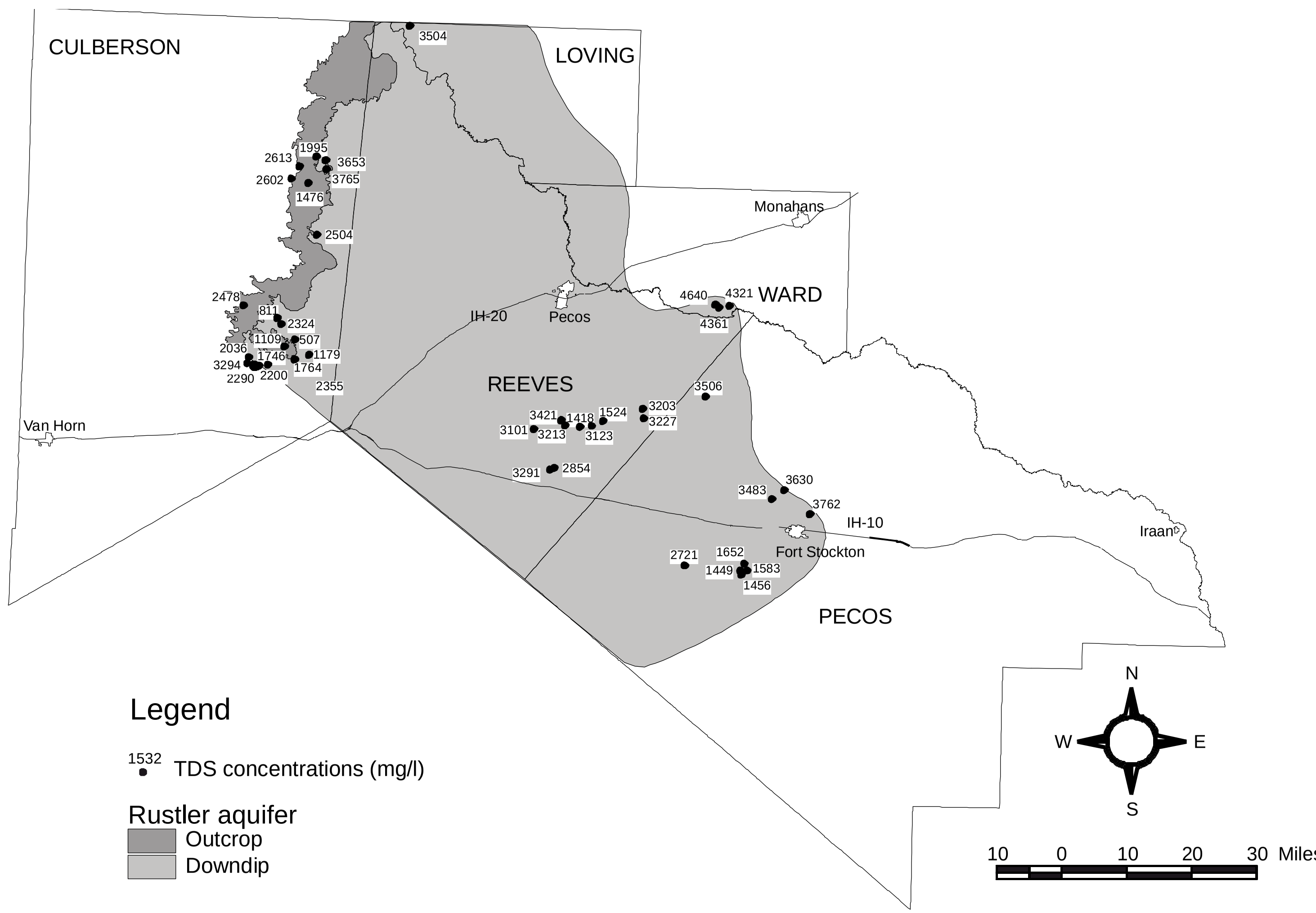


Figure 15-7. Salinity distribution in Rustler aquifer. Data from TWDB.

Table 15-3. Isotope composition in Rustler groundwater samples, Pecos County

| State Well Number | Date Sampled | Tritium (TU) | pmC[1] |
|---|---|---|---|
| 53-01-203 | Aug. 1996 | 0.00 | N/A |
| 53-01-203 | Aug. 2000 | 0.00 | 5.25 |
| 52-16-613 | Aug. 2000 | 0.06 | 13.91 |

[1] Percent Modern Carbon

The two stock wells in southeastern Culberson County shown on figure 15-6 are completed in different intervals of the Rustler aquifer. Pre-1995 data are sporadic at best but seem to show a trend of storage depletion for well 47-54-302 and relatively steady-state conditions for well 47-54-201. The post-1995 portions of the hydrographs show similar trends in both wells and indicate possible hydraulic communication between them.

The fluctuations shown by the hydrographs most likely reflect long-term variations in water-use patterns. Armstrong and McMillion (1961) estimated that some 7,500 acre-ft of Rustler water was pumped in 1958 from Pecos County, mainly for irrigation. From 1980 to 1993, the average groundwater pumpage stood at 354 acre-ft/yr (TWDB Water Use Survey). The aquiferwide water use increased to over 1,550 acre-ft/yr from 1994 through 1997 (TWDB Water Use Survey). The resumption of irrigation pumpage of the Rustler in Pecos County accounts for this escalation. Despite the rise in withdrawal rates, water levels in wells 52-16-608 and 52-16-609 continued to recover from 1994 through 2001 (fig. 15-6).

# Recharge, Discharge, and Water Availability

Ogilbee and others (1962) stated that recharge to the Rustler aquifer occurs by precipitation and infiltration of streams in its Rustler Hills outcrop, as well as by cross-formational flow. The Tessey Formation, an equivalent to the Rustler Formation that crops out in the Glass Mountains of Pecos and Brewster Counties, is also thought to contribute recharge to the Rustler aquifer.

Boghici (1997) looked at the tritium and $^{14}C$ composition of groundwater from two Rustler wells in Pecos County. Tritium and $^{14}C$ are radioisotopes used to determine the age of water (table 15-3). The samples are virtually devoid of tritium and exhibit low radiocarbon activities, which is typical for older waters in slow-moving flow systems and not for permeable aquifers with short groundwater residence times, as the Rustler aquifer is purported to be. Water temperature at the time of sampling ranged from 28º to 31º Celsius, 3º to 7º warmer than the rest of the wells in the Rustler aquifer. These findings imply that, at least in Pecos County, very little recharge is by percolation of recent rainfall and stream seepage, but most of it may be contributed by cross-formational flow of old water from deeper formations. Recharge by rainfall could also be impeded by the high potential evapotranspiration in the area, which is about nine times higher than the precipitation rate (Armstrong and McMillion, 1961). Veni (1991) suggested that shallow groundwater north of Fort Stockton may be in part derived from the underlying Capitan Limestone of the Permian Delaware Mountain Group. Low tritium activities in the Capitan groundwater (Boghici, 1997) support Veni's assertion and may point to it as potential source of the flow in the Rustler.

Groundwater discharges from the Rustler aquifer primarily through well withdrawals as springs and seeps (for example, at Diamond Y and Rustler Hills) and as cross-formational flow into the overlying Edwards-Trinity aquifer. Results of numerical groundwater flow modeling in Pecos County by Boghici (1997) indicate that 260 acre-ft/yr of water from the Rustler aquifer may be discharged through the Diamond Y fault system, and some 3,800 acre-ft of water per year may be upwelling into the overlying Cretaceous strata in the Belding area. Locally, where the water table is shallow, some discharge may take place by evapotranspiration. The Texas Water Plan (1997) estimates that approximately 4,000 acre-ft of Rustler water should be available for use every year without using water from storage.

# Groundwater Geochemistry

The salinity distribution for waters of the Rustler aquifer is shown in figure 15-7, which was built using data from 40 samples collected by TWDB between 1956 and 2000. All but two groundwater samples are brackish, with total dissolved solids (TDS) concentrations ranging between 507 and 4,640 mg/L. There is no clear pattern in salinity variations along the presumed direction of groundwater flow. Fresher waters have been found in some downdip wells, more so than in the outcrop.

Groundwater in the study area is predominantly of the Ca-Mg-$SO_4$ facies (fig. 15-8), reflecting the prevailing dolomitic-gypsiferous nature of the Rustler Formation. Several downdip samples (shown as triangles in fig. 15-8) are more dilute (1,500–1,700 mg/L TDS) and show distinctly different compositions of a Na-Cl- $SO_4$ type. These samples are from areas underlain by the Belding-San Simon Trough (Hiss, 1976). The dissolution-induced thinning of the Rustler and extensive deep faulting in this locale could provide an opportunity for Na-Cl water from the underlying Ochoan section to up well and mix with the sulfate-rich Rustler aquifer. Mineral saturation indices computed by the geochemical modeling program PHREEQC (Parkhurst, 1995) show that groundwater from the Rustler aquifer is undersaturated with respect to calcite and dolomite and at equilibrium with gypsum.

The aquifer mineralogy, mineral equilibria, and chemical composition suggest that carbonate and gypsum dissolution may be the main processes affecting the groundwater chemistry of the Rustler aquifer.

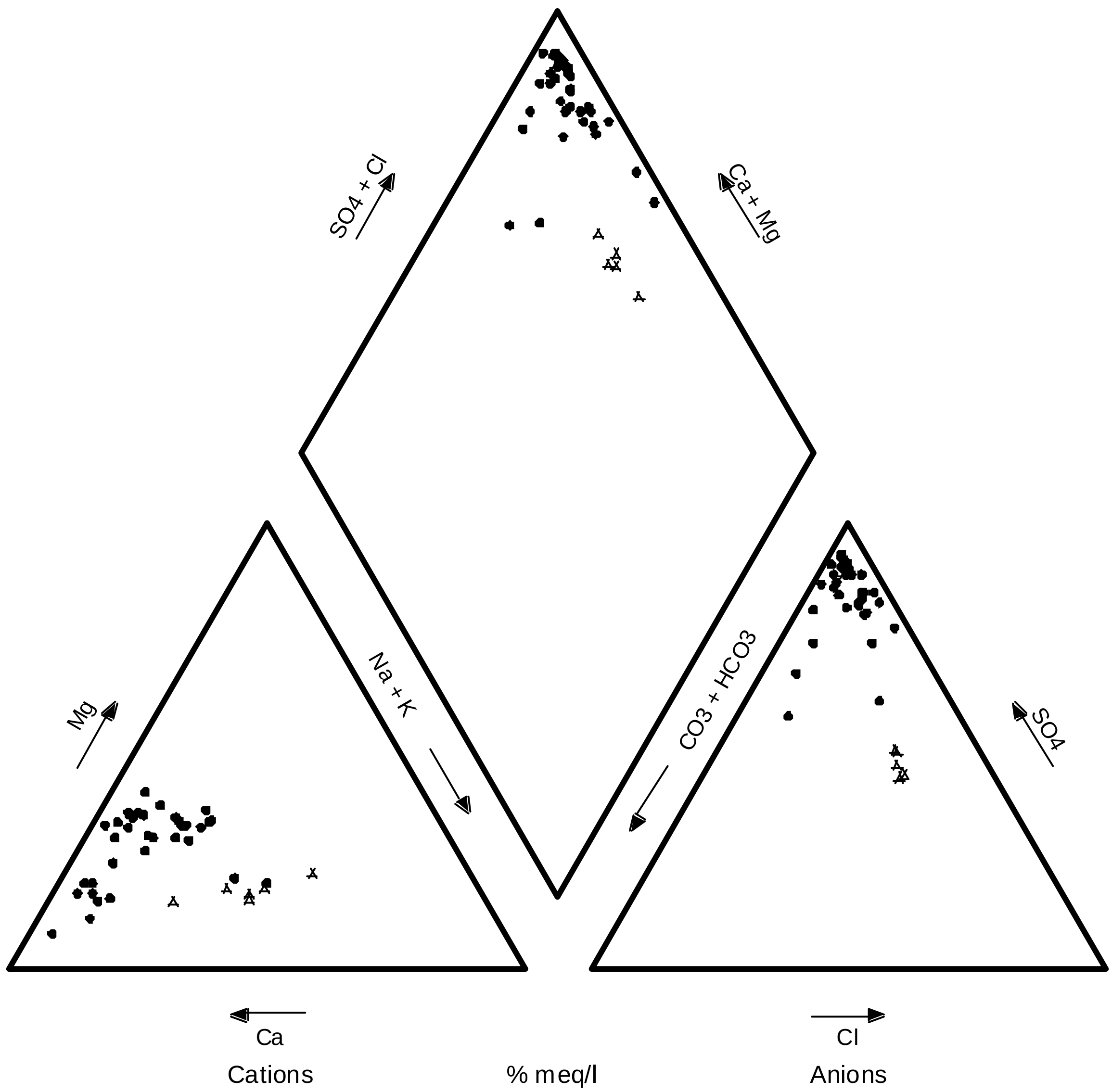


Figure 15-8. Piper plot showing Rustler groundwater composition. Data from TWDB.

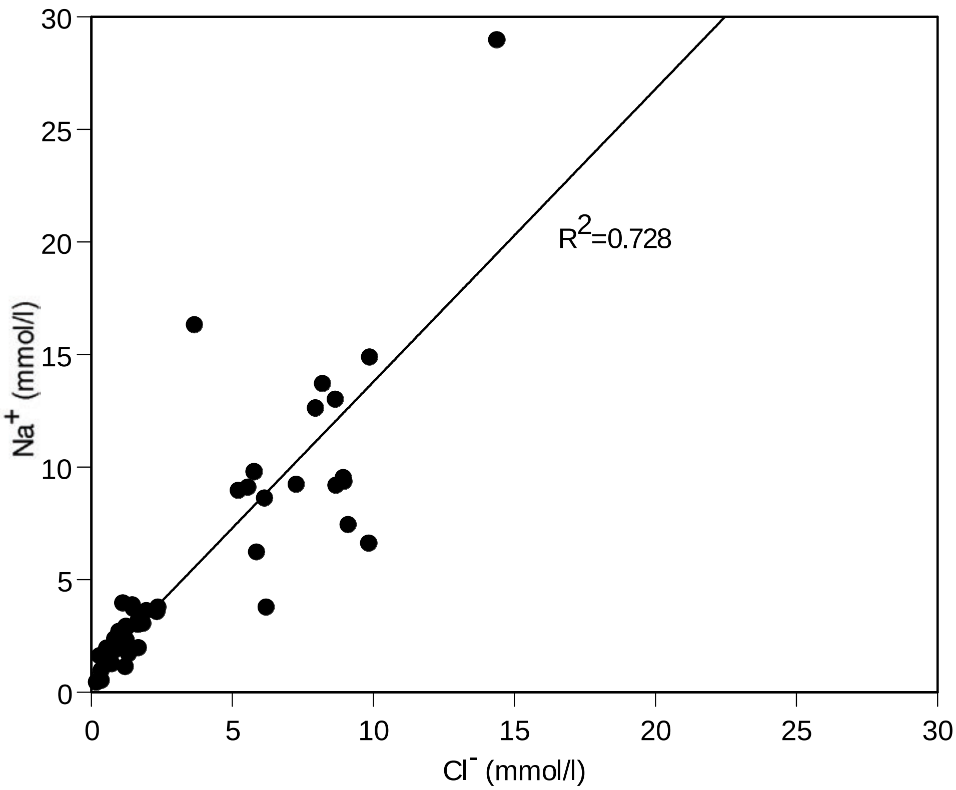


Figure 15-9. Plot of $Na^+$ versus $Cl^-$.

Following are the governing equations for prominent mineral dissolution and precipitation reactions occurring in aqueous systems:

Calcite dissolution and precipitation:

$$CaCO_3 + CO_2 + H_2O \Leftrightarrow Ca^{2+} + 2HCO_3^- \quad (1)$$

Dolomite dissolution:

$$CaMg(CO_3)_2 + 2\,CO_2 + 2H_2O \Leftrightarrow Ca^{2+} + Mg^{2+} + 4HCO_3^- \quad (2)$$

Gypsum dissolution:

$$CaSO_4 \,.\, 2H_2O \Leftrightarrow Ca^{2+} + S\,O_4^{2-} + 2H_2O \quad (3)$$

Halite dissolution:

$$NaCl + H_2O \Leftrightarrow Na^+ + Cl^- + H_2O \quad (4)$$

Ion exchange:

$$2Na(clay) + Ca^{2+} \Leftrightarrow Ca(clay) + 2Na^+ \quad (5)$$

A plot of sodium against chloride (fig. 15-9) is roughly linear, with a slope of 1.3 and an intercept near origin, indicating that some sodium and chloride may come from halite. The predominance of sodium over chloride indicates a source of sodium beyond halite dissolution.

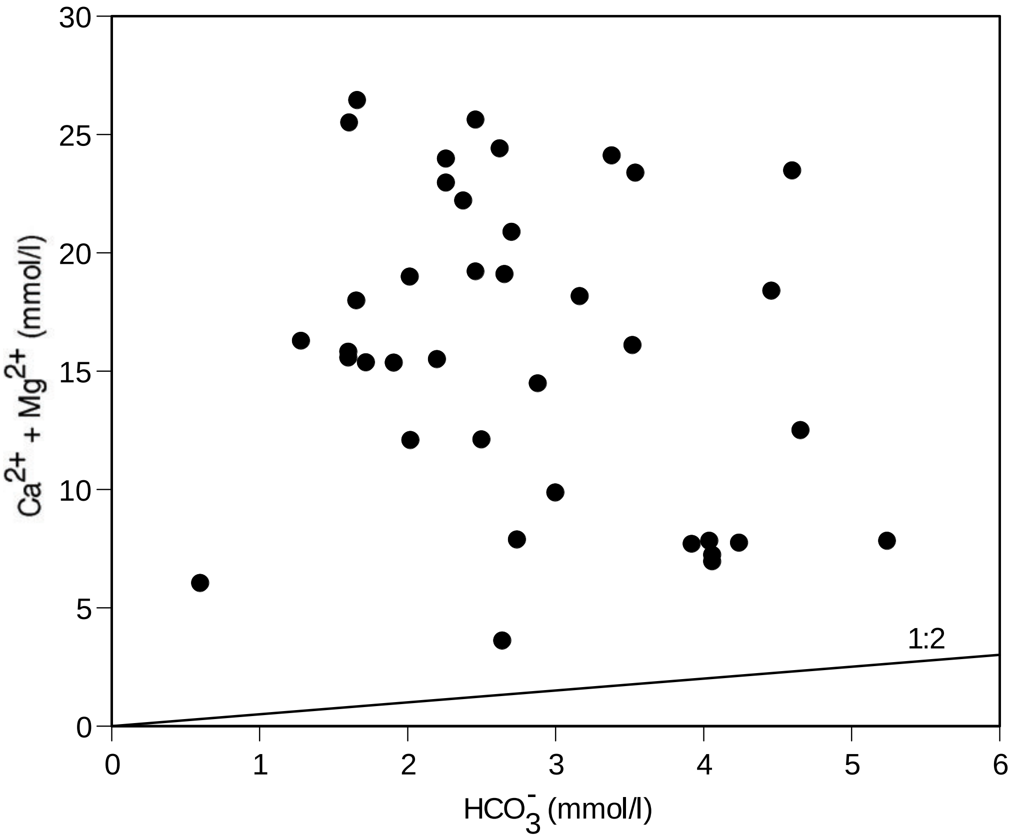


Figure 15-10. Plot of $Ca^{2+}$+$Mg^{2+}$ versus $HCO_3^-$.

Figure 15-10 shows the relationship between the concentration of calcium and magnesium versus bicarbonate. If all calcium and magnesium were derived from calcite and dolomite dissolution, then data would plot along a line with a slope of 1:2, as stated by equation 1. All points in this figure are above the 1:2 line, indicating an additional source of calcium and magnesium.

Additional calcium can be found in abundance in the gypsum-bearing Rustler Formation. To account for the calcium derived from gypsum dissolution, calcium and magnesium molar concentrations are summed up and plotted against the sum of sulfate and half of bicarbonate concentration (fig. 15-11). The major-ion water chemistry suggests that the calcium, magnesium, sulfate, and bicarbonate present in the water are the result of a simple dissolution of the available dolomite or magnesium-calcite, along with gypsum or anhydrite. In an ideal case, such dissolution reactions would result in these samples plotting on a straight line through the origin with a slope of one. The Rustler waters plot along an obvious line with a slope of 0.875 (fig. 15-11), a good coefficient of correlation ($R^2$=0.974), and an intercept near zero. The slope of the trend line suggests that there is a partial loss of calcium plus magnesium relative to the amount of bicarbonate and sulfate present. This is consistent with a partial cation exchange where some of the calcium plus magnesium is lost from the water and sodium is gained. This interpretation explains why most of the water samples have a higher ionic concentration of sodium than chloride (fig. 15-9), which indicates that there is a source of sodium beyond halite dissolution.

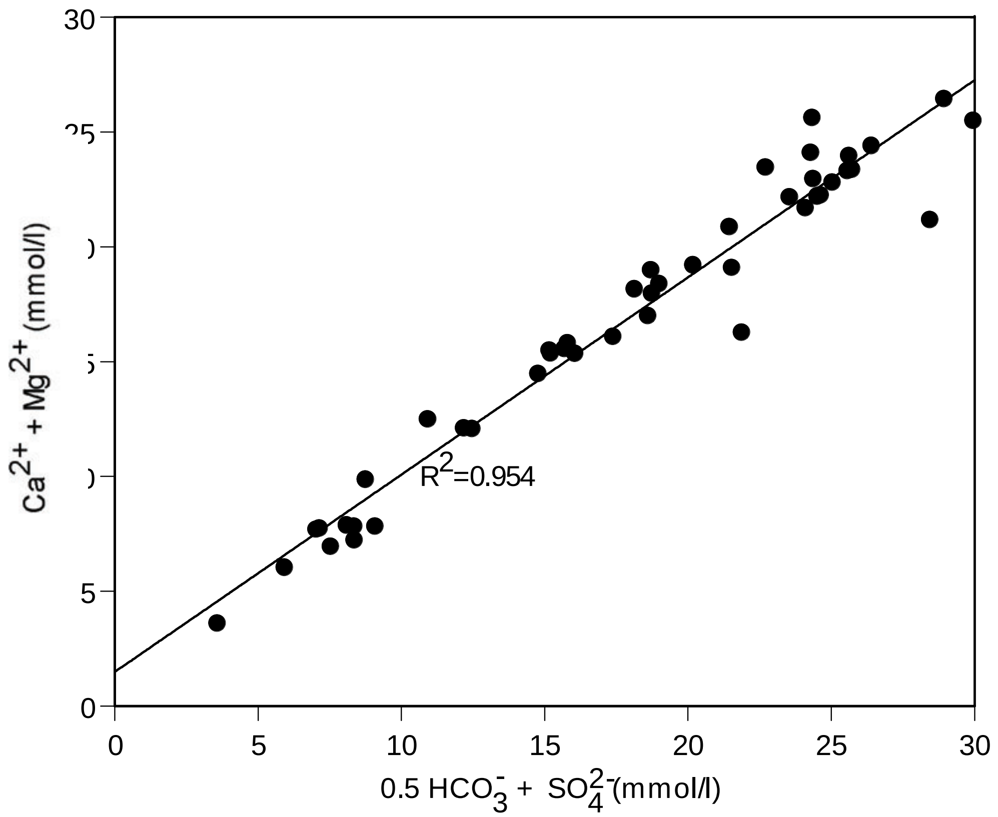


Figure 15-11. Plot of $Ca^{2+}$+$Mg^{2+}$ versus $SO_4^{2-}$+1/2 $HCO_3^-$.

To test the ion exchange hypothesis, the concentration of ($Na^+$–$Cl^-$) is plotted against ($Ca^{2+}$+$Mg^{2+}$-$SO_4^{2-}$–1/2 $HCO_3^-$). The quantity ($Na^+$–$Cl^-$) represents excess sodium, that is, sodium coming from sources other than halite dissolution, assuming that all chloride is derived from halite. The quantity ($Ca^{2+}$+$Mg^{2+}$-$SO_4^{2-}$–1/2 $HCO_3^-$) represents the calcium and/or magnesium coming from sources other than gypsum and carbonate dissolution. These two quantities represent the maximum amount of sodium and calcium plus magnesium available for ion-exchange processes.

The samples plot near a line with a slope of 2:1 (fig. 15-12), suggesting that some cation-exchange reactions are taking place where the aquifer media permit it. Waters undergoing exchange of calcium and magnesium for bound sodium on clays will gradually become of sodium-sulfate type. The fact that calcium is still the dominant cation in most of these samples indicates that exchange reactions have not yet occurred extensively. However, a close examination of these data shows that the cation exchange is somewhat more involved than this. There is more magnesium in the water than can be accounted for by

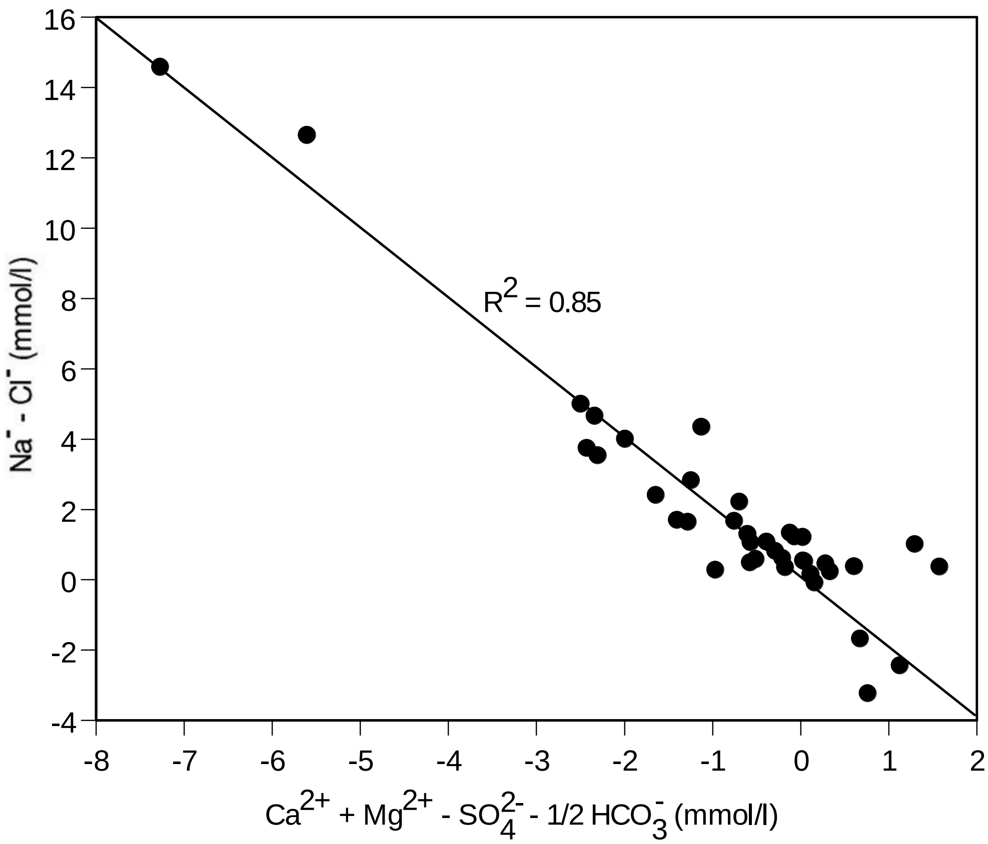


Figure 15-12. Plot of $Na^+$-$Cl^-$ versus $Ca^{2+}$+$Mg^{2+}$-$SO_4^{2-}$ -1/2 $HCO_3^-$.

the dissolution of dolomite. It is thought that these data also represent a significant amount of ionic exchange where calcium is lost and magnesium is gained.

More than half of the data points in figure 15-12 represent wells located in the Rustler outcrop in Culberson County. For ion exchange to take place it is necessary that, in addition to having a suitable clay medium, enough time be allowed for the reaction to proceed. This observation, coupled with the high salinities observed in outcrop samples, seems to indicate slow groundwater recharge rates and longer groundwater residence times in this recharge area.

# Summary

The Rustler aquifer of Trans-Pecos Texas is in the carbonates and evaporites of the Rustler Formation, of Late Permian age. The aquifer yields brackish water to irrigation and stock wells. Most of the water production comes from fractures and solution openings in the formation Upper Member. Recharge to the Rustler aquifer is by precipitation on its outcrop in Culberson County and by cross-formational flow from deeper aquifers. Geochemical and isotopic data indicate that parts of the aquifer contain old water, and modern recharge events may be less common than previously thought. Discharge from the Rustler takes place mainly through wells, by seeps and springs, and by leakage into the overlying strata. Rustler water quality is variable and ranges from fresh to brackish. The geochemical data for this aquifer fit a carbonate, gypsum, and halite dissolution and base-exchange model fairly well.

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# Chapter 16

# The Aquifers of Red Light Draw, Green River Valley, and Eagle Flat

Bruce K. Darling[1] and Barry J. Hibbs[2]

## Introduction

The water-planning studies conducted as part of Senate Bill 1 for the Far West Texas Regional Water Planning Group have generated much interest in the groundwater resources of the westernmost counties of Texas. This paper is intended to provide a basic description of Red Light Draw, Green River Valley, and Eagle Flat—all of which are part of the complex of West Texas bolsons. As such, the paper is based principally on (1) a study of the fresh and slightly saline groundwaters of westernmost Texas (Gates and others, 1980), (2) an evaluation of the suitability of the Eagle Flat Basin to be the location for a repository for low-level radioactive waste (Darling and others, 1994), and (3) a Ph.D. dissertation on the hydrogeology of the basins (Darling, 1997).

## Location and Physiographic Setting

Eagle Flat, Red Light Draw, and Green River Valley (fig. 16-1) are located approximately 100 mi east of the City of El Paso. The only settlements in the area are the unincorporated villages of Sierra Blanca and Allamoore.

The Diablo Plateau is a low-relief upland that slopes toward the north from an escarpment that forms the southern boundary of the extensive tableland and the northernmost extent of the region described in this paper. Three topographic basins lie south of the plateau.

Eagle Flat covers an area of approximately 560 mi$^2$. The northwestern area of Eagle Flat is a closed topographic depression (the Blanca Draw Watershed), which drains through Blanca Draw, an ephemeral stream, into a playa known as Grayton Lake. Water accumulates in Grayton Lake only after periods of heavy rainfall. Eagle Flat Draw drains the southeastern area of the larger Eagle Flat Basin. This drainage area is referred to later as the Southeast Eagle Flat Watershed. Toward the north-northeast, the watershed is bounded by the Carrizo Mountains. The Blanca Draw Watershed is bordered along the southwest by the rugged sandstone and limestone spines of Devil Ridge and by the Eagle

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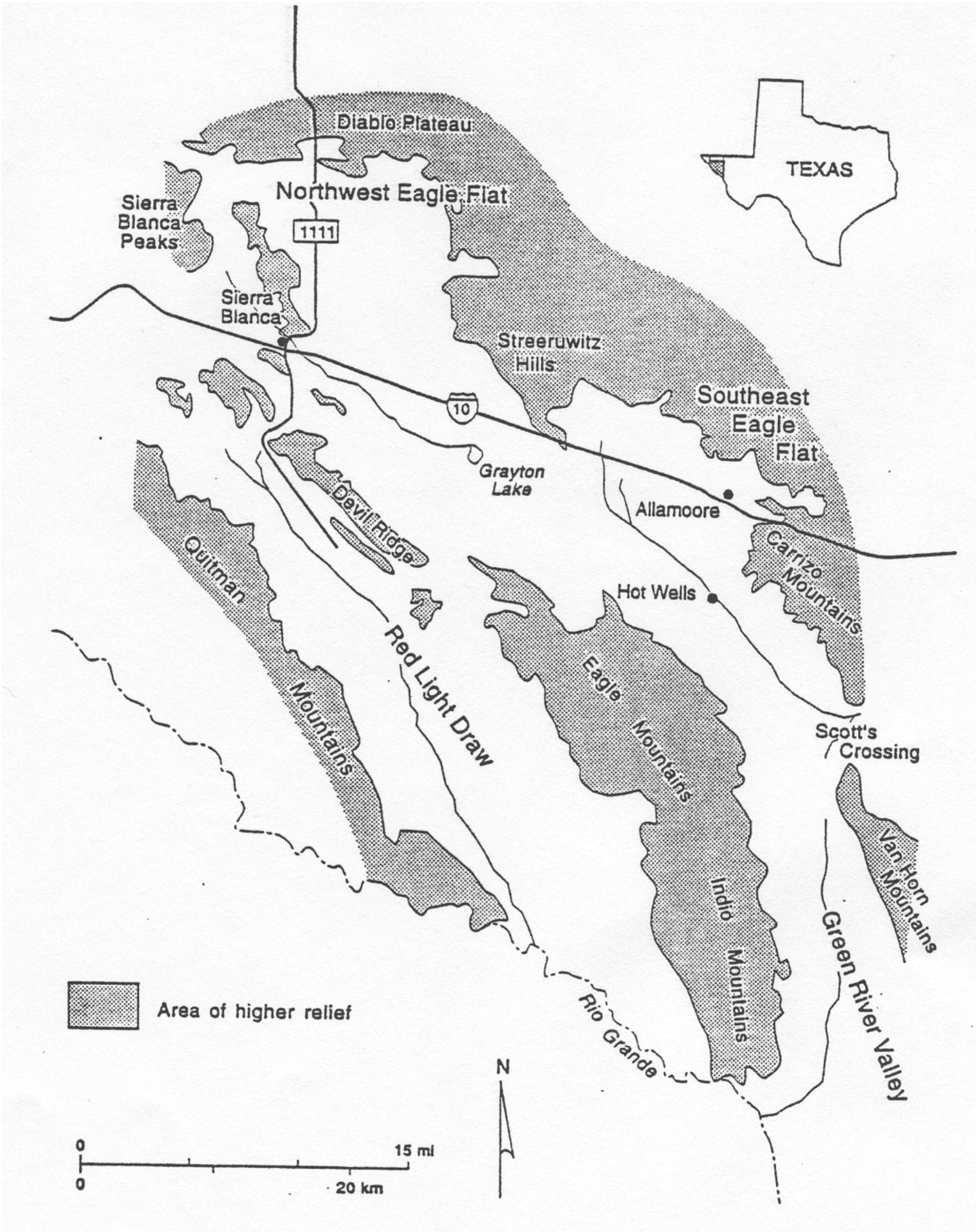


Figure 16-1: Location and major physiographic features.

Mountains. The Southeast Eagle Flat Watershed is bordered along the southwest by the Eagle Mountains.

Red Light Draw is flanked along the northeast by the Indio Mountains, the Eagle Mountains, Love Hogback, and Devil Ridge. The Rio Grande forms the southern boundary of the basin. Red Light Draw encompasses an area of approximately 370 $mi^2$. This watershed is drained by a Red Light Arroyo, an ephemeral tributary of the Rio Grande. The basin is bounded along the southwest by the Quitman Mountains. The Rio Grande is the only perennial stream in the area.

Green River Valley lies in parts of Hudspeth, Culberson, Jeff Davis, and Presidio Counties. This basin, which lies between the Indio Mountains on the west and the Van Horn Mountains on the east, is drained by the Green River, an ephemeral tributary of the Rio Grande. The surface area of the watershed is approximately 160 $mi^2$.

## Elevations

The Eagle Mountains form the highest topographic point in the area, reaching an elevation of 7,484 ft above mean sea level (msl). Sierra Blanca Peak rises to an altitude of 6,800 ft above msl. The Carrizo and Van Horn Mountains are more than 5,200 ft above msl, and the highest elevation of the Quitman Mountains is approximately 6,200 ft above msl. The villages of Sierra Blanca and Allamoore are at 4,500 ft above msl, and the center of the Grayton Lake playa of Northwest Eagle Flat lies at an elevation of 4,270 ft above msl, the lowest point in the Blanca Draw Watershed. Over a distance of more than 25 mi, the floor of Red Light Draw decreases from nearly 4,500 ft above msl in the northernmost reaches of the watershed to between 3,200 and 3,100 ft above msl along the Rio Grande. The floor of Green River Valley decreases from approximately 4,250 ft above msl in the northernmost area of the basin to around 3,100 ft above msl along the Rio Grande.

## Climate

The mean annual temperature is 65°F. The average annual low temperature is 48°F, and the average high is 81°F. Sierra Blanca receives approximately 10 to 12 inches of precipitation each year. Mean annual evaporation is 84 inches. Precipitation decreases to between 7 and 8 inches at El Paso, where evaporation is approximately the same as at Sierra Blanca. Hudspeth County and neighboring counties usually record the lowest annual precipitation of reporting stations in Texas. Low rainfall and high evaporation combine to create drought conditions during all or part of most years (Larkin and Bomar, 1983).

Most precipitation occurs during the months of July through October as widely scattered thunderstorms (Larkin and Bomar, 1983; Nativ and Riggio, 1989, 1990). Most winter rainfall is associated with widespread frontal systems that originate over the Pacific Ocean (Elliot, 1949; Nativ and Riggio, 1989, 1990). Winter storms account for less than one-third of the region’s total precipitation.

# Aquifers

Red Light Draw, Green River Valley, and Eagle Flat are part of the complex of West Texas bolsons (Ashworth and Hopkins, 1995). It is generally understood by hydrogeologists that each bolson encompasses a separate aquifer; however, the degree to which adjacent basins are hydrogeologically integrated is not well understood because there has been little research that would allow such inferences to be made with certainty.

## Red Light Draw Aquifer

Wells in Red Light Draw produce groundwater from Cretaceous rocks, Cenozoic basin fill, Tertiary igneous rocks, and Quaternary river alluvium. Cretaceous limestones and sandstones predominate in the northernmost areas of the draw, giving way to basin fill in the central and southern areas of the basin. Wells within the Rio Grande floodplain produce groundwater from coarse- to fine-grained sand and silt of the Rio Grande alluvium. Although the Texas Water Development Board (TWDB) considers the Rio Grande alluvium of the Hueco Bolson to be separate from the Hueco Bolson aquifer (Hibbs and others, 1997), we regard the alluvium and the basin fill of Red Light Draw to be part of the larger Red Light Draw aquifer. The thickness of the basin fill ranges from less than 400 ft in the northernmost reaches of Red Light Draw to between 1,000 and 2,000 ft in the central area of the basin, and to more than 3,000 ft in the vicinity of the Rio Grande (Gates and others, 1980). The base of the saturated section of the formations that underlie the basin fill is unknown. The depth to the potentiometric surface is 200 ft or less in the mountainous areas that surround the basin (fig. 16-2) and as much as 400 ft beneath the northern and central areas of Red Light Draw. (The potentiometric surface is the level to which groundwater rises in a well.) The depth to the potentiometric surface decreases to less than 25 ft within the Rio Grande floodplain.

The Red Light Draw aquifer is not a source of water for municipal supply. All current production is either for domestic use or for the watering of livestock and wild game. Large-capacity wells completed in the Rio Grande alluvium supplied water to irrigate cotton fields during the 1950’s and 1960’s. The farms in this area of Red Light Draw were abandoned in the 1970’s, and the irrigation wells are no longer in use.

## Green River Valley Aquifer

The Green River Valley aquifer consists of limestone, sandstone, conglomerate and siltstone of Cretaceous age, and Tertiary volcanics. The maximum thickness of the basin-fill deposits ranges from 1,700 to 2,000 ft. The basin fill includes thick sequences of coarse-grained volcanic material eroded from the surrounding mountains. Near the Rio Grande, the thickness of the basin fill is more than 2,000 ft. In this area, the basin fill consists predominantly of clay, silt, and possibly altered tuff (Gates and others, 1980). The depth to the potentiometric surface ranges from less than 200 ft in the mountains that bound the basin to as much as 400 ft within the central area of the basin (fig. 16-2). The depth decreases to less than 25 ft within 1 mi of the Rio Grande. A few windmills and wells equipped with submersible pumps supply water to the ranches in Green River

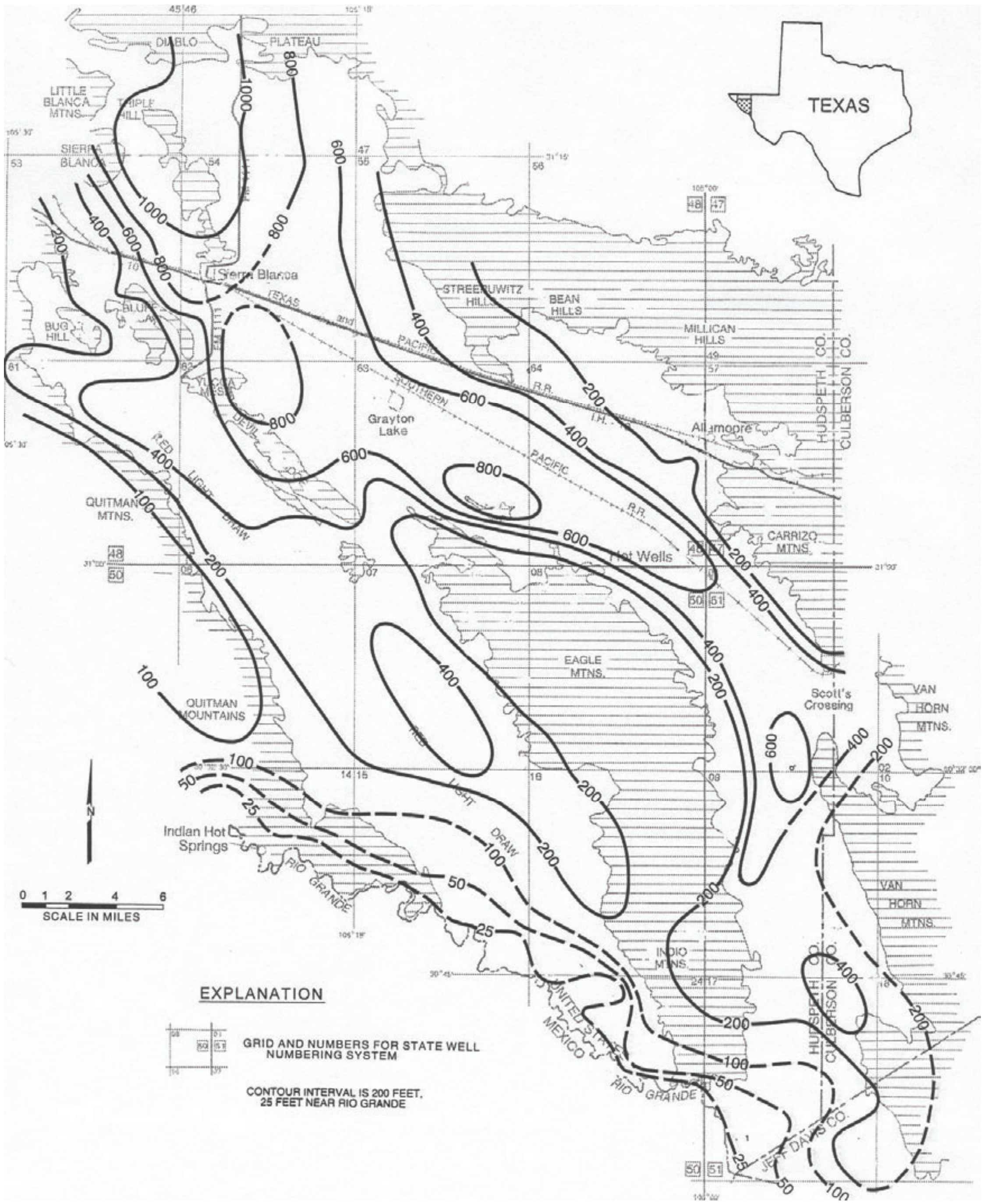


Figure 16-2: Depth to the potentiometric surface (adapted from Darling, 1997).

Valley, one of the most remote and rugged of the bolsons of West Texas. The depth of most water wells is less than 200 ft, and well yields vary widely, from less than 25 gpm in shallow wells along the margins of the basin to more than 100 gpm in irrigation wells (now abandoned) along the Rio Grande. All groundwater production is for the watering of livestock or wild game.

## Eagle Flat Aquifer

Metamorphosed rocks of Precambrian age make up the Eagle Flat aquifer between the Streeruwitz Hills and the Carrizo Mountains of Southeast Eagle Flat. Wells in this area of the basin are either windmills or small-diameter boreholes equipped with submersible pumps. These wells, most of which are 100 to 200 ft deep, are sufficiently productive to provide water for domestic use and for watering of livestock. The depth to the potentiometric surface is generally 200 ft or less in the area between the Streeruwitz Hills and the Carrizo Mountains (fig. 16-2).

Farther to the south and southeast, the Eagle Flat aquifer consists of interbedded sequences of sand, silt, and clay. In the 1970’s, the U.S. Geological Survey (USGS) drilled four test wells to assess availability of the fresh –to –slightly saline groundwater resources of westernmost Texas (Gates and White, 1976). (A concise summary of the drilling program in the Red Light Draw/Eagle Flat area is Gates and Smith, 1975). The USGS drilled one of the four wells midway between Scott’s Crossing and Hot Wells (locations marked on figs. 16-1 and 16-2). This well penetrated 2,100 ft of sand, silt, and clay without encountering bedrock. At Hot Wells, two wells drilled in the early 1900’s to supply water for steam locomotives produced groundwater from coarse-grained alluvial-fan material shed from the Eagle Mountains. Each well was 1,000 ft deep, and each was fitted with 10-inch (internal diameter), slotted steel casing (Gates and others, 1980). The wells are reported to have been capable of producing several hundreds of gallons of water per minute. The wells, which lie within the right-of-way of the Union Pacific Railroad, are no longer in service. The depth to the potentiometric surface of the Allamoore System ranges from 400 to 600 ft (fig. 16-2). In the middle to upper sections of the alluvial fan that borders the northwest face of the Eagle Mountains, the depth to the potentiometric surface is 200 ft or less. Scattered windmills and small-diameter wells equipped with submersible pumps produce water for domestic use and for watering of livestock and wild game in this area.

Limestone and sandstone formations of Cretaceous age make up the aquifer beneath Northwest Eagle Flat. The Cenozoic basin fill, which is as much as 500 to 700 ft thick in the central area of Northwest Eagle Flat, is not known to be a source of groundwater in this area of the basin (Gates and others, 1980). The saturated thickness of the Cretaceous formations that lie beneath the basin fill is unknown. Wells drilled to evaluate the suitability of Northwest Eagle Flat to be the location of a repository for low-level radioactive waste did not fully penetrate the Cretaceous bedrock (Darling and others, 1994). Pumping tests conducted in conjunction with the investigation indicate that the transmissivity of the Cretaceous bedrock formations is highly variable. In many cases, drawdowns of 100 ft or more were recorded at pumping rates that ranged from 10 to 15 gpm (Darling and others, 1994). A smaller number of wells appeared to be capable of substantially larger yields (Gates and others, 1980; Darling and others, 1994; Darling, 1997). The wide variability of well yields is probably related to fault-induced fracturing of the bedrock. The wells with higher yields may be located within zones of denser fracturing. Wells with lower yields may be completed in blocks that have not been highly fractured. The depth to the potentiometric surface ranges from 600 ft along the margins of the basin to between 800 and 1,000 ft beneath the floor of the basin (fig. 16-2).

The Cretaceous formations used to be the only source of drinking water for the unincorporated village of Sierra Blanca (population ~700). In the early 1970's, a newly developed municipal well field 5 mi to the west-southwest of Sierra Blanca was abandoned when water levels fell precipitously after 1 yr of operation (Gates and others, 1980). The rapidly falling water levels were probably related to stresses caused by pumpage from rocks of relatively low permeability and low groundwater storage capacity. Sierra Blanca now gets its water from a well field at the Van Horn municipal airport in Culberson County, 35 mi to the east.

# Potentiometric Map

A map of the potentiometric surface (fig. 16-3) delineates four major groundwater divides in the area. A groundwater divide is a naturally occurring hydrologic boundary between adjacent basins or within a basin represented by a high in the potentiometric surface. A groundwater divide is a hydraulic barrier to the direct lateral flow of groundwater. The first divide, which is roughly parallel to the rim of the Diablo Plateau, separates the groundwaters of the Diablo Plateau aquifer to the north from those of the Eagle Flat aquifer to the south. This hydrologic barrier is the *Plateau groundwater divide*.

The second divide lies between the Carrizo Mountains and the Eagle Mountains, forming a narrow saddle beneath the valley between both mountain ranges. This is the *Eagle Flat groundwater divide*. Groundwaters flowing northward from the Eagle Mountains and southward from the Carrizo Mountains converge beneath the floor of Southeast Eagle Flat to form the Eagle Flat groundwater divide. This divide creates two separate flow systems within the Eagle Flat aquifer. The system east of the divide is referred to as the *Allamoore flow system*, and the component west of the divide is the *Sierra Blanca flow system* (Darling, 1997). The potentiometric surface of the Allamoore system is broad and flat, and the 3,800-ft contour is open toward the east in the vicinity of Scott's Crossing, indicating flow toward the east (that is, toward the Lobo Valley aquifer). The potentiometric surface of the Sierra Blanca system is broad and flat, and the 3,600-ft contours in the center of the system are closed. The closed contours representing the surface of the Sierra Blanca system indicate no direct pathway for flow out of the basin.

The third divide extends northwestward from the Eagle Mountains and extends beneath Love Hogback and Devil Ridge. This barrier, referred to as the *Devil Ridge groundwater divide*, lies between the Red Light Draw aquifer and the Sierra Blanca flow system of the Eagle Flat aquifer. The Red Light Draw aquifer originates in the mountains and uplands surrounding Red Light Draw. The potentiometric surface of the Red Light Draw aquifer

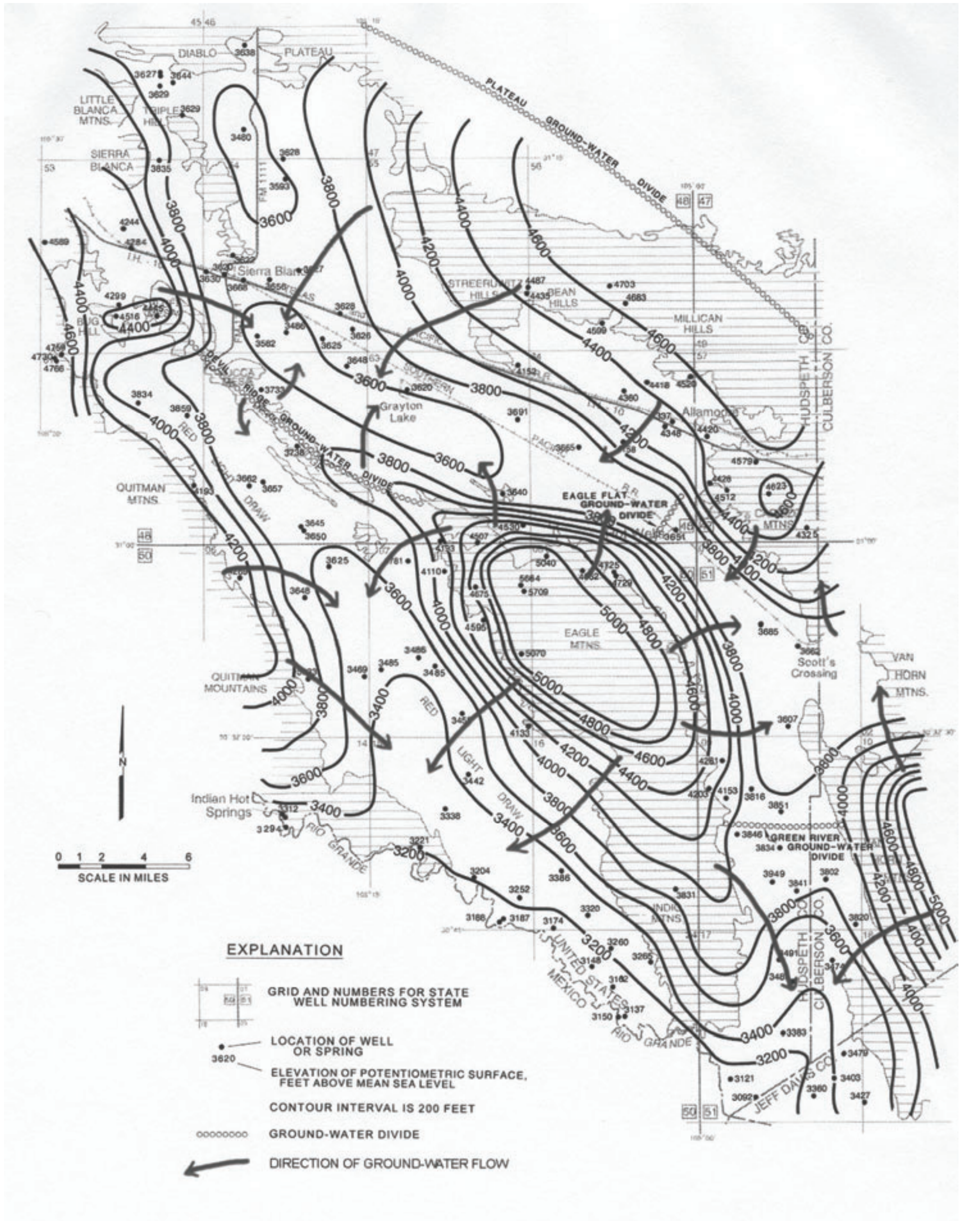


Figure 16-3: Map of the potentiometric surface (adapted from Darling, 1997).

slopes toward the Rio Grande, decreasing from approximately 3,600 ft above msl beneath the northernmost part of the draw to between 3,200 and 3,100 ft above msl in areas adjacent to the river.

The fourth divide extends from beneath the Eagle Mountains eastward to the Van Horn Mountains, forming a broad potentiometric high beneath the topographic high that establishes the boundary between Southeast Eagle Flat and Green River Valley. This is the *Green River groundwater divide*. This divide separates flow in the Green River Valley aquifer from the groundwaters of the Allamoore flow system. Over a distance of approximately 10 mi, the elevation of the potentiometric surface of the Green River Valley aquifer decreases from around 3,900 ft along the highest point of the groundwater divide to between 3,200 and 3,100 ft along the Rio Grande.

# Generalized Groundwater Flow Paths

The map of the potentiometric surface (fig. 16-3) provides a basis for delineating flow paths in each of these aquifers.

## Red Light Draw and Green River Valley Aquifers

The Red Light Draw and Green River Valley aquifers originate in the mountains that bound the basins. Groundwaters of these basins converge beneath the floors of their respective watersheds. The direction of flow is southward toward the Rio Grande, which lies within the discharge zone of each aquifer. This pattern of flow toward the Rio Grande is characteristic of all other West Texas bolsons that are dissected by the river (Hueco, Presidio, Redford Bolsons).

## Eagle Flat Aquifer

Lying to the north of both Red Light Draw and Green River Valley, Eagle Flat is not dissected by a major through-flowing stream such as the Rio Grande or by a smaller stream that drains to the Rio Grande. The floor of the basin lies at a higher elevation than that of the floors of Red Light Draw and Green River Valley, and, with the exception of minor springs in the middle to upper elevations of the Eagle Mountains, there is no area within Eagle Flat where groundwater discharges to the surface. The incision of bolsons by the Rio Grande 2 million yr ago initiated a set of conditions that have allowed for the flow of groundwater from these basins toward the river. The groundwater flow regimes of Eagle Flat and of other basins that have not been dissected by the Rio Grande (e.g., Ryan Flat, Lobo Valley, Wildhorse Flat, and Michigan Flat) are unlike those of Red Light Draw and Green River Valley (as well as the Hueco Bolson and the Presidio and Redford Bolsons). A few researchers have suggested that these aquifers might be linked to a deep regional flow system that transports groundwater toward the east (Nielson and Sharp, 1985; Sharp, 1989). Eagle Flat is the westernmost of the bolsons that are not bordered by the Rio Grande. The degree to which this basin is hydrogeologically integrated with others is not well understood. The following two sections of this paper present a set of hypotheses regarding possible directions of flow from this aquifer.

### *Allamoore Flow System*

Flow lines indicate that the direction of flow from the Allamoore system is toward the east (fig. 16-3). This interpretation is based partly on consideration of water-level measurements from Lobo Valley (Gates and others, 1980; Darling and others, 1994; LBG-Guyton Associates, 1999), where the potentiometric surface manifests a northward-directed gradient at elevations ranging from approximately 3,700 to 3,600 ft above msl immediately east of Scott's Crossing (Gates and others, 1980; LBG-Guyton Associates, 1999). If bedrock formations are sufficiently permeable to permit the flow of groundwater from the Allamoore system to the Lobo Valley aquifer, then the lower hydraulic potential east of the Scott's Crossing basin boundary should allow for the possibility of flow from the Allamoore system to the Lobo Valley aquifer. The lower hydraulic head of the Green River Valley aquifer (between 3,200 and 3,100 ft along the Rio Grande) cannot be ignored because the elevations along the river also suggest the potential for flow toward the south beneath the Green River groundwater divide from deeper sections of the Allamoore system. However, the few data in this area do not permit the partitioning of flow between eastward and southward components from the Allamoore system at this time. The best that can be argued is that the lower hydraulic head in basins to the east and to the south of Southeastern Eagle Flat underscore only the potential for flow in one or both directions.

### *Sierra Blanca Flow System*

Both the Eagle Flat and the Devil Ridge groundwater divides appear to limit direct lateral flow from the Sierra Blanca system to the Allamoore system and to Red Light Draw. The most likely avenue for the transfer of groundwater from the Sierra Blanca system is along vertical pathways to more porous and permeable rocks that underlie the Cretaceous bedrock. The depth to the static water level (fig. 16-2) and the relatively flat potentiometric surface (fig. 16-3) suggests the influence of a drain that permits flow downward to an intermediate or a regional flow system.

At least one well offers support for the occurrence of higher porosity and permeability in bedrock formations of the Eagle Flat Basin. In 1965, Texaco, Inc., drilled a 1,700-ft core test (Capitan Drilling Co., No. 1 Espy Ranch) approximately 5 mi to the west of the Eagle Flat groundwater divide. The surface elevation at the well was reported to be 4,368 ft above msl. The drilling record on file at the TWDB shows that the borehole penetrated 240 ft of basin fill before encountering limestone. The record also reports lost circulation in bedrock between depths of 1,590 ft and 1,700 ft. The drillers were unable to regain circulation, and the well was plugged and abandoned at a depth of 1,700 ft.

This core test was one of the deepest recorded penetrations of bedrock in Eagle Flat. The loss of circulation reported for this test occurred at an elevation of 2,778 ft above msl, or nearly 1,000 ft below the potentiometric surface in this part of the Sierra Blanca flow system. The loss of circulation in the well suggests that higher permeability pathways in bedrock formations might provide a deep bypass of local groundwater divides. Similar pathways for flow beyond basin boundaries in the Great Basin of southern Nevada have been described by the USGS (Winograd, 1962).

Two possible directions of flow from the Sierra Blanca system are postulated. The first is toward the east, beneath the Allamoore system. The second is toward the south, beneath the Red Light Draw aquifer. The lowest elevations of the potentiometric surface of the Sierra Blanca system are approximately 3,600 ft above msl; and within the Allamoore system, the lowest elevations are approximately 3,700 ft above msl in the vicinity of Scott's Crossing (fig. 16-3). Along the Rio Grande, however, the elevations are between

3,200 and 3,100 ft above msl. The river, therefore, establishes the zone of lowest hydraulic potential in the study area.

The higher hydraulic head of the Allamoore system may limit the potential for the eastward flow of groundwater from the Sierra Blanca system. However, the nearly 400-ft difference in head between the lowest areas of the Sierra Blanca system and the Red Light Draw aquifer indicates that groundwater of the Sierra Blanca system might flow southward, beneath the Devil Ridge groundwater divide—first along vertical flow paths and then southward in deeply buried Cretaceous formations and Tertiary igneous rocks. Gates and others (1980) also observed that “the available water-level data are not sufficient to trace the movement of ground water in northwestern Eagle Flat, but the data suggest that the water may discharge through the Cretaceous rocks in the subsurface, probably toward the Rio Grande to the south.”

# Recharge of the Aquifers

Researchers with the USGS have surmised that the aquifers of West Texas are recharged by infiltration along the foothills of mountains and plateaus and locally along the channels of ephemeral streams in the basins (Gates and others, 1980). Their recharge estimates, however, are based on the assumption that 1 percent of average annual precipitation of 12 inches distributed uniformly across the area is available to recharge the aquifers of Red Light Draw, Green River Valley, and Eagle Flat. Proceeding from this assumption, the USGS (Gates and others, 1980) estimated that average annual recharge attributable to precipitation might be as much as 2,000 acre-ft in Red Light Draw, 1,000 acre-ft in Green River Valley, and 3,000 acre-ft in Eagle Flat (an acre-ft is equal to 325,851 gallons.)

Darling (1997), on the basis of his research using analyses of the radioactive isotopes carbon-14 and tritium to delineate recharge areas of Red Light Draw and Eagle Flat and also on a cross-sectional numerical flow model through Red Light Draw (Hibbs and Darling, 1995), concluded that recharge in Red Light Draw occurs only along the higher elevations of the mountain fronts and not within the middle to lower elevations of alluvial fans or along the floors of the basins. He estimated that total recharge might be as little as 14 percent (or 280 acre-ft) of the average annual estimates by Gates and others (1980). If this is typical of other areas, then recharge in Eagle Flat and Green River Valley might be as low as 430 and 120 acre-ft, respectively.

More recently, LBG-Guyton Associates and others (2001) used a modification of the approach by Gates and others (1980) to derive estimates of recharge for each of the bolsons of West Texas. The aquifer outlines as shown on maps published by the TWDB (Ashworth and Hopkins, 1995) were regarded as “storage” areas, and only the highlands that form the boundaries of the aquifers were considered to be primary recharge areas. One percent of average annual rainfall over the highlands was assumed to be available as recharge. All precipitation and runoff over the storage zones were assumed to be removed by evaporation and transpiration. (In Red Light Draw and Green River Valley, runoff to the Rio Grande also accounts for a pathway to remove surface water from the basins.)

Using this approach, LBG-Guyton Associates and others (2001) estimated that recharge might be 700 acre-ft/yr in Red Light Draw and Green River Valley and approximately 1,000 acre-ft/yr in Southeastern Eagle Flat. The lower estimates of recharge derived by Darling (1997) and by LBG-Guyton Associates and others (2001) should be interpreted to indicate only that a reasonable basis exists for inferring that the aquifers of this area might receive a much smaller amount of recharge than estimates based on an assumed relationship between annual precipitation and the entire surface area of a basin.

# Water Quality

The quality of water varies widely, not only between but also within the basins (fig. 16-4). Chloride concentrations in the central area of the Sierra Blanca flow system range from 5 to 20 millimoles/liter (180 to 710 milligrams/liter [mg/L]). The highest concentrations occur within the central areas of the flow system. In the Allamoore system and the Green River Valley aquifer, chloride concentrations are typically less than 35 mg/L. In the Red Light Draw aquifer, the concentration of chloride ranges from 35 to 70 mg/L except in areas along the river where concentrations increase to between 700 and 5,000 mg/L. The higher concentrations of chloride associated with the Rio Grande alluvium are attributed principally to the flushing of salts that accumulated over long periods of time from the evaporation of water used to irrigate crops and to the infiltration of salty irrigation-return water from the Rio Grande. The elevated concentrations of chloride are typically accompanied by sulfate concentrations that range from 300 to as much as 2,000 mg/L. The secondary drinking-water standards promulgated by the U.S. Environmental Protection Agency specify 250 mg/L as the maximum concentration of chloride and of sulfate. Although most high-chloride/high-sulfate waters are not regarded as suitable for consumption by humans, they are typically acceptable for watering of livestock.

# Potential for Development

The development potential of the Red Light Draw, Green River Valley, and Eagle Flat aquifers has not been fully examined. Research programs conducted by the USGS (Gates and others, 1980) and The University of Texas at Austin, Bureau of Economic Geology (BEG) (Darling and others, 1994) have not explicitly addressed this important issue. The disappointing performance of the municipal well field developed for Sierra Blanca (Gates and others, 1980) and the results of aquifer tests conducted in Eagle Flat by the Bureau of Economic Geology (Darling and others, 1994) could cause skepticism regarding the large-scale development potential of Northwestern Eagle Flat. Without significantly more detailed hydrogeologic investigations in other areas, nothing can be stated with certainty regarding the development potential of Southeastern Eagle Flat, Red Light Draw, or Green River Valley. Further comment on this matter would be little more than speculation.

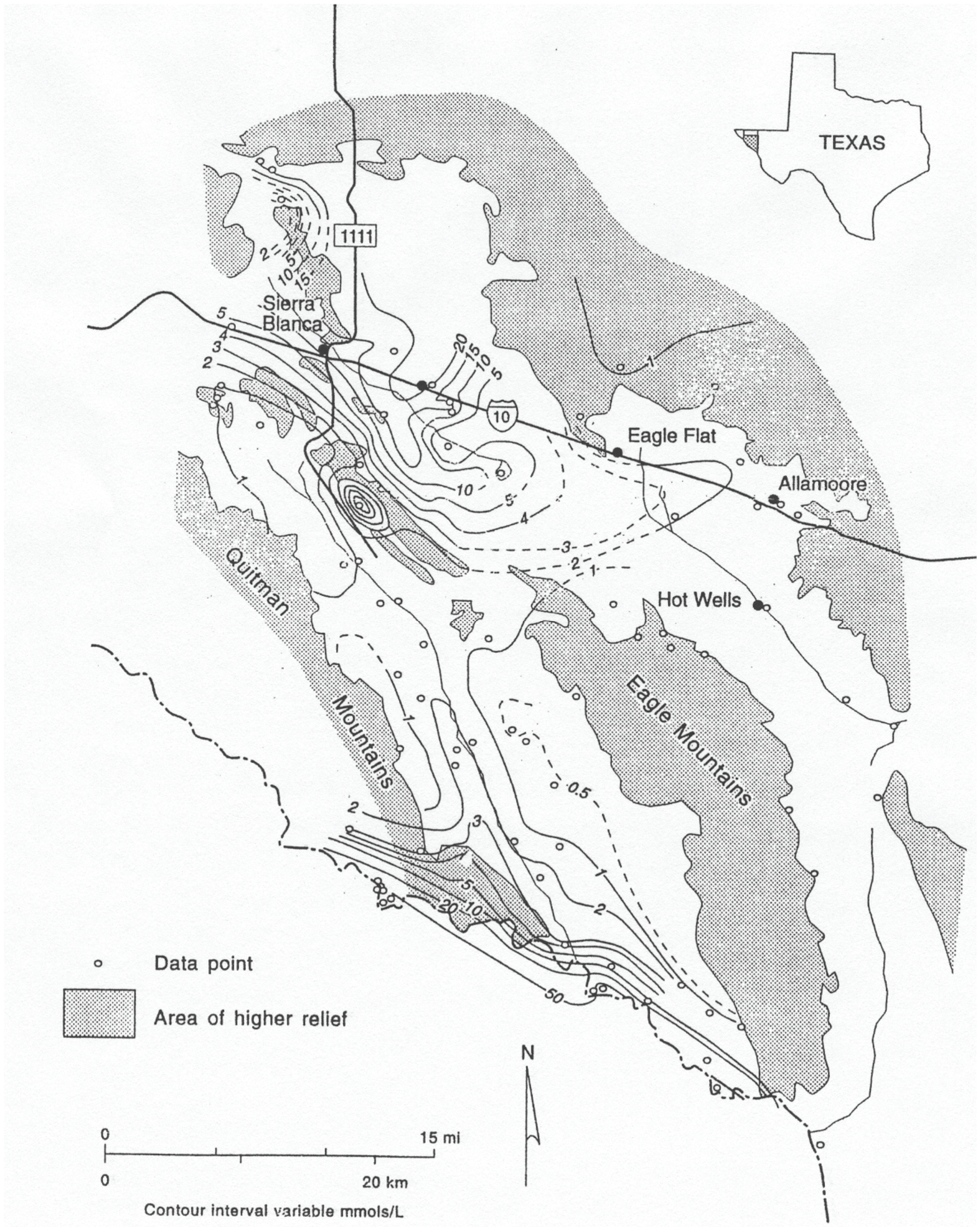


Figure 16-4: Chloride concentrations in groundwater (adapted from Darling, 1997).

# Conclusions

The aquifers of Red Light Draw, Green River Valley, and Eagle Flat present an interesting set of problems with regard to our understanding the hydrogeologic integration of the bolsons of West Texas. Groundwaters of Red Light Draw and Green River Valley, for example, flow toward the Rio Grande—in a manner consistent with all other bolsons down-cut by the river. The flow of groundwater from Eagle Flat, however, is not clearly apparent. Located north of Red Light Draw and Green River Valley, Eagle Flat appears to have more in common with other bolsons that are not bordered by the Rio Grande. There are insufficient data to allow flow paths from the Allamoore and Sierra Blanca systems to be delineated with certainty. At least two pathways from each system are possible, and a substantial amount of hydrogeologic research will be required to determine where the groundwaters of these systems flow.

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# Chapter 17

# Hydrogeology of the Salt Basin

Edward S. Angle[1]

## Introduction

The Salt Basin of West Texas has been a significant source of groundwater to local users in West Texas for most of the last century. In a region of normally low rainfall and high evaporation, groundwater is a vital resource to municipalities, industries, and landowners in the Salt Basin. Because El Paso is facing serious water shortages in the next 20 to 30 years, city and regional planners are looking, in part, to water resources in the Salt Basin. It is therefore important to understand how pumping and drought impact the aquifers of the Salt Basin to maintain its viability for West Texans in the future.

Although many people think of the Salt Basin as being the salt flats north of Van Horn, the Salt Basin as a physiographic feature extends much farther south and includes the Salt, Wild Horse, Michigan, Lobo, and Ryan Flats (fig. 17-1). The sediments beneath these flats hold water that form part of the West Texas Bolsons aquifer (Ashworth and Hopkins, 1995). The purpose of this paper is to present a brief overview of the hydrogeology of the area and to present results of water-level and water-quality information collected between 1992 and 1994 by the Texas Water Development Board.

### Physiography

The Salt Basin is located in the Trans-Pecos region of West Texas. It forms a valley that extends from just north of the New Mexico border in Hudspeth and Culberson Counties along a southeastern trend through western Jeff Davis County, where it ends in the northwest portion of Presidio County. The Salt Basin is approximately 140 miles long and 25 miles across at its widest point (fig. 17-1).

The Salt Basin is a classic expression of basin and range tectonism where a broad, flat valley trending roughly north and south is bounded on the east and west by uplifted, fault-block mountains (Underwood, 1980). The basin valley separates the Diablo Plateau and Sierra Diablo, Baylor, Beach, Carrizo, Van Horn, and Sierra Viejo Mountains in the west from the Guadalupe, Delaware, Apache and Davis Mountains to the east. The highest point in the basin is at Guadalupe Peak 8,751 ft above mean sea level (amsl), while the lowest point is 15 mi due west in the Salt Flats at 3,600 ft amsl. Along the basin

[1] Texas Water Development Board

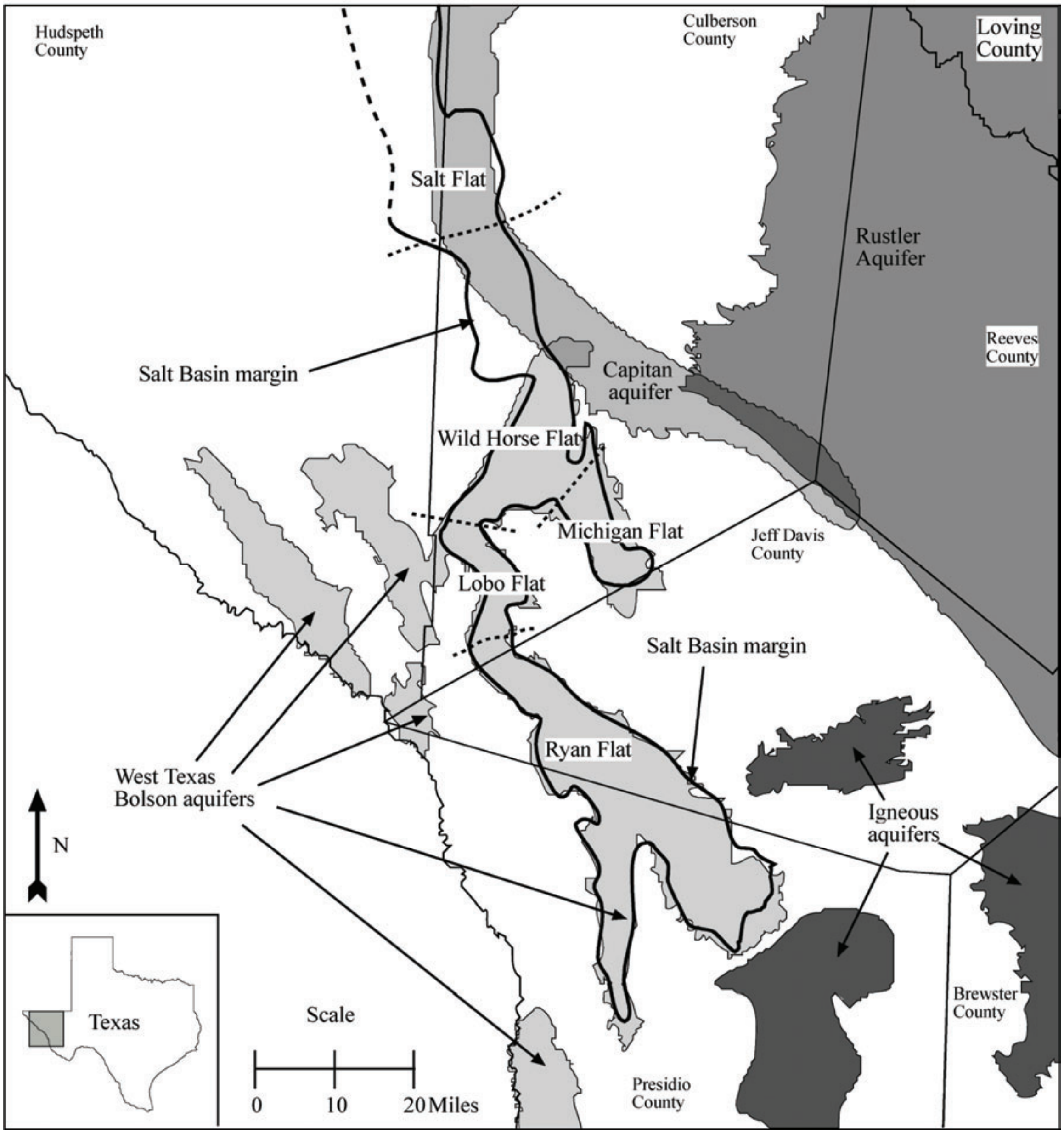


Figure 17-1: Location of the Salt Basin on base map showing the aquifer systems near the Salt Basin, the Salt Basin margins and the physiographic subdivisions of the Salt Basin.

axis from the north, the elevation of the valley floor begins at 3,600 ft amsl and remains fairly level until a midway point, where it climbs to 5,000 ft amsl in the very southern part of the basin. The basin terminates at the southern end in the Oak Hills.

Topographically the Salt Basin is a closed basin with surface drainage inward toward the center of the basin and northward to the Salt Flats (Underwood, 1980). There are no recorded perennial streams in the basin. Stream flow is often the result of rainfall activity in mountainous regions or locally intense rain showers that result in overland flow.

## Climate

The climate in the Trans-Pecos region typically varies with elevation (Underwood, 1980). Below 4,000 ft amsl, climate in this region is arid to subtropical. The higher elevations tend to be cool, temperate, and humid, with warm summers. The Salt Basin has a characteristic arid environment, with hot summers and mild winters. Average daily temperatures per year are fairly constant across the basin, ranging from a low of 60°F in Marfa (period of record: 1978–1997) to a high of 63°F in Van Horn (1942–1997) (HED, 1995).

Average annual precipitation varies across the Salt Basin, although the annual amount of rainfall is low across the basin. Average rainfall totals increase with elevation and toward the southern portion of the basin. In the Salt Flats area, data collected from 1959 through 1977 show an average of 9.1 inches/yr of rainfall. In comparison, the Guadalupe Mountains, 15 mi due east, record an average of 18 inches/yr during the period 1987–1997 (HED, 1995). Farther south in the Salt Basin, the cities of Van Horn (1942–1997), Valentine (1978–1997) and Marfa (1940–1997) show an average of 10.3, 13.4 and 15.5 inches/yr of rainfall, respectively (TWDB, 2001).

Rainfall events in the Salt Basin typically occur as brief, local, high-intensity thunderstorms that deliver .05 to .2 inches of water. Three-quarters of the total yearly rainfall for the basin falls between the months of May and October. This high-rainfall period also corresponds with the period of highest temperature and evaporation. The evaporation rate is 76.2 inches/yr (TWDBEDP, 2001), five to eight times greater than average annual rainfall.

## Geology

The stratigraphic record in the Salt Basin includes the Precambrian through the entire Phanerozoic, with only a few gaps between. Examples of volcanism, metamorphism, and sedimentary deposition can all be found in the Salt Basin. On the basis of geologic composition, the basin can be divided into a northern and southern portion. The basin flanks in the north consist mostly of Permian-age sedimentary formations, while the flanks in the south consist of Cretaceous and Tertiary volcanics. The basin fill in the north reflects the mostly carbonate source rock in the form of Tertiary and Quaternary alluvium, lacustrine sands, silts, muds, and evaporate deposits. The basin fill in the south reflects the volcanic sources on the flanks, as well as volcanic deposits, pyroclastic debris, lava flows, and tuffaceous deposits.

The structural history of the Salt Basin begins sometime during the Laramide orogeny, with a single-faulted monocline (Muehlberger and Dickerson, 1989). This initial fault was then later reactivated during Late Cenozoic time and subsequently evolved into the horst and graben valley we see now.

The north-trending ranges of the Guadalupe, Delaware, and Apache Mountains border the eastern side of the basin. The eastern margin of the Salt Flat is defined by the fault scarp that forms the western edge of the Guadalupe Mountains and exposes the massive, Permian-age Capitan Formation (Underwood, 1980). The Capitan Formation, composed of the Capitan Limestone and the Goat Seep Limestone, is a reef system deposited on the margins of the Delaware Basin (Bebout and Kerans, 1993). Farther south along this eastern flank, the Bell Canyon, Cherry Canyon, and Brushy Canyon Formations are exposed in the Delaware and Apache Mountains. It is in this northern area that the Salt Basin gets its name, from the numerous salt playas that have formed on the western side of the valley floor. West of the Delaware Mountains are the Sierra Diablo Mountains. These mountains are unique because an almost complete geologic record from the Precambrian through the Cretaceous can be seen here (Underwood, 1980). Only Cambrian, Triassic, and Jurassic ages are absent from this exposure. More toward the center of the basin, near the town of Van Horn, are the Beach and Baylor Mountains. These are small, up-thrown fault blocks composed of Precambrian, Ordovician, and Permian-age rocks. A third up-thrown block of Permian limestone forms the Wylie Mountains that bifurcate the valley into Lobo Flats to the west and Michigan Flats to the east.

The Tertiary volcanics of the Davis Mountains compose the eastern sides of the southern portion of the Salt Basin. The Precambrian strata of the Carrizo Mountains, the Cretaceous sandstones and limestones of the Van Horn Mountains, and the Tertiary volcanics of the Sierra Vieja Mountains form the western sides of the basin. The Van Horn Mountains act as a bridge from the Permian to the Tertiary, with exposures of the Cox Sandstone, the Bluff Mesa, Loma Plata and other Cretaceous Formations (Barnes, 1979). The southern portion of the Salt Basin is dominated by Tertiary volcanic formations. The Petan Basalt, Bracks Rhyolite, Capote Mountain Tuff, the Sheep Pasture Formation and many others make up the southern half of the Wylie Mountains, the Davis mountains in the east, and the Sierra Vieja mountains to the west (Barnes, 1979). The southern end of the basin terminates in the Tertiary conglomerates that form the Oak Hills.

# Hydrogeologic Setting

## Hydrostratigraphy

The Salt Basin consists mostly of late Tertiary- and Quaternary-age deposits that fill the basin. The basin fill reflects the local composition of the bordering mountains

Table 17-1. Characteristics of water-bearing units in the Salt Basin area (after Gates and others, 1980; Boyd and Kreitler, 1986).

| Erthem [what?] | System | Unit | Physical and Lithologic Characteristics | Water-Bearing Characteristics |
|---|---|---|---|---|
| Cenozoic | Quaternary and Tertiary | Bolson deposits | Unconsolidated clay, silt, sand, and gravel derived from weathering and erosion of local rock and deposited by ancestral drainages within the Salt Basin; commonly 1,000 to 2,000 ft thick. Interbedded carbonates, gypsum, and salines of the playas derived from evaporation of groundwater originating in Permian strata of the surrounding highlands. | Supplies moderate to large quantities of fresh to saline water to the northern parts of the Salt Basin, mostly in fine-grained lacustrine and alluvial deposits. |
| | Tertiary | Volcaniclastic and volcanic deposits | Reworked tuffs and alluvial deposits consisting almost exclusively of volcanic debris (volcanic clastics) interbedded with ash-flow tuffs and volcanic flows up to 6,000 ft thick in Ryan Flat. | Supplies small to large quantities of fresh water in Ryan and Lobo Flats; permeable zones probably most common in the uppermost 1,000 ft and may include well-reworked tuff, well-sorted volcaniclastics, weathered zones above and below volcanic flows, and possibly fractured volcanic-flow rocks. |
| Mesozoic | Cretaceous | Cox Sandstone | Mostly quartz sandstone with some pebble conglomerate and siltstone, shale, and limestone; very fine to medium grained; commonly less that 200 ft thick but can be as much as 700 ft thick. | Supplies small to moderate quantities of fresh to moderately saline water in eastern and southern Wild Horse Flat. |
| Paleozoic | Permian | The Capitan and Goat Seep Limestone, undifferentiated limestones and sandstones, including the Delaware Mountain Group | Capitan and Goat Seep limestones are massive, thick-bedded reef limestones and dolomite; Capitan is 1,000 –2,000 ft thick in the Guadalupe Mountains and Beacon Hill area and up to 900 ft thick in the Apache Mountains area; the Delaware Mountain Group is sandstone and limestone with some siltstone; aggregate thickness is on the order of 3,000 feet. Permian gypsum deposits of the Guadalupe and Delaware Mountains contribute significant amounts of sulfate to the groundwater system. | Capitan and Goat Seep Limestones supply moderate to large quantities of fresh to slightly saline water in the Beacon Hill area. The Capitan supplies moderate to large quantities of fresh to slightly saline water in the Apache Mountain area. The sandstones and limestones of the Delaware Mountain Group supply small quantities of slightly to moderately saline water along the eastern side of the northern Salt Basin and foothills of the Delaware Mountains. |

(table 17-1). In the Salt Flats, Wild Horse Flats, and Michigan Flats area, the fill is mostly lacustrine and fluviatile deposits of clay, silt, sand, gravel, caliche, and gypsum (Barnes, 1983). This fill reflects the surrounding highland Permian sedimentary deposits of similar composition. In the Ryan and Lobo Flats area the basin fill is similar to that of the northern areas, but the clay, silt, and sandstone are red in color and the conglomerates are composed of volcanic materials—pebbles and cobbles of vesicular, aphanitic, and porphyritc textures (Barnes, 1979).

## Recharge

The basin fill is recharged by infiltration of precipitation along basin boundaries and by groundwater flow between flats. Groundwater recharge from surface infiltration generally occurs along the valley margins and foothill regions, where surface sediments are courser grained and more permeable (Gates and others, 1980). Some recharge might occur in ephemeral streambeds, but the majority of this recharge is likely lost to evaporation or transpiration. Because recharge is difficult to quantify in West Texas, it is generally assumed to be about 1 percent of average annual precipitation (Gates and others, 1980). Assuming 11.5 inches of rain annually and a drainage area of 2,760 $mi^2$, the recharge to the Salt Basin from surface infiltration could be about 17, 000 acre-ft/yr (Gates and others, 1980). Basin fill beneath the flats is also recharged by groundwater flow within the Salt Basin. Groundwater flows from Ryan Flat into Lobo Flat, from Lobo Flat into Wild Horse Flat, and from Wild Horse Flat into Michigan Flat. Water levels in the southern portion of the Salt Flat indicate that groundwater flows from the Salt Flat area into the northern portion of Wild Horse Flat.

## Discharge

Discharge occurs by a variety means from the Salt Basin. Evaporation from the land surface occurs throughout the basin but is a primary means of discharge in the Salt Flat area. Cross-formational flow from the basin fill into adjacent formations is also important and seems most apparent in the Wild Horse Flat area. Groundwater pumping accounts for the majority of discharge from the entire Salt Basin.

### *Salt Flat*

Groundwater in the Salt Flat discharges by evaporation, leakage to other formations, and pumping. Boyd and Kreitler (1986) found that groundwater in the Salt Flat area discharges by evaporation where the land surface is bare and by transpiration from phreatophytes where the land is vegetated.

Because the hydraulic conductivity in the basin fill is one or two orders of magnitude less than the underlying Capitan Reef aquifer, very little water discharges from the Salt Flat to the Capitan (Hiss, 1980).

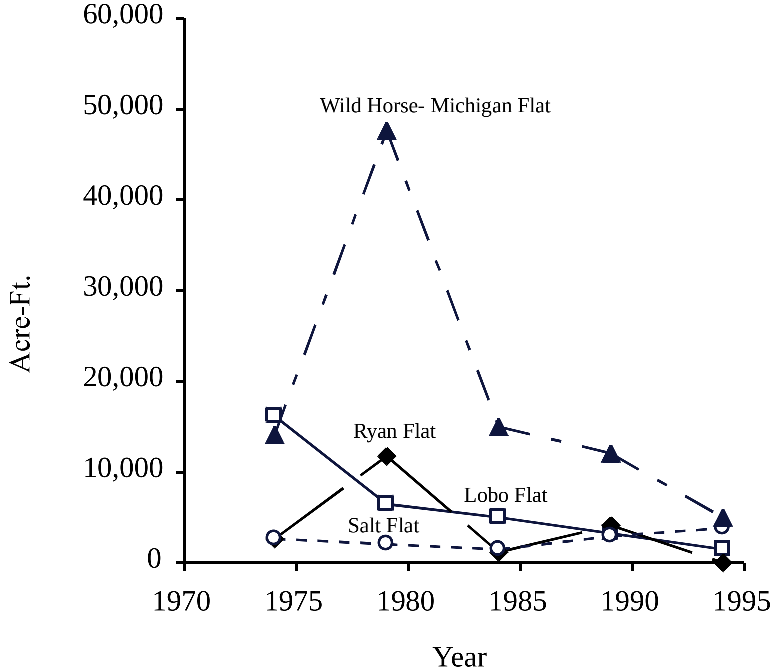


Figure 17-2. Volume of irrigation water used in Salt Basin.

The first irrigation wells were drilled in the Salt Basin area in the late 1940's. This development continued slowly until about 1973, when the last recorded irrigation well was drilled. The majority of groundwater pumped in the Salt Flat area comes from the Capitan and Goat Seep Limestones that lie under the basin fill. Pumping has been fairly steady at about 2,600 acre-ft/yr from 1974 through 1994 (fig. 17-2).

***Wild Horse Flat***

Groundwater in Wild Horse Flat discharges by leakage to other formations, pumping, and outflow from the bolson. Gates and others (1980) reported that water levels showed potential for discharge to the east. LaFave and Sharp (1987) suggested that a component of the spring flow in the Balmorhea area is a result of flow from the southern Salt Basin (the Michigan Flat area), through the Capitan Formation, into the Lower Cretaceous and then exiting at the springs. Although groundwater pumping for irrigation accounts for the greatest withdrawals in the Wild Horse and Michigan Flats area, it has been on the decline steadily since it peaked in 1984 (fig. 17-2). There was some pumping for industrial use at a talc-processing facility between 1972 and 1991. The yearly average of pumping at the talc plant was 5.5 acre-ft/yr. The City of Van Horn has six public supply wells that have averaged 654 acre-ft/yr of production between 1972 and 1994.

### *Lobo and Ryan Flat*

Most of the discharge from Lobo Flat is from pumping. Gates and others (1980) reported that between 1951 and 1972 about 320,000 acre-ft, or an average of about 15,000 acre-ft/yr, of water was pumped. However, pumping has slowly declined since the 1970's (fig. 17-2). Pumping in the Ryan Flat area accounts for the majority of discharge, and since the 1970's it has been in steady decline. By 1994, there was no pumping for irrigation in Ryan Flat according to the TWDB survey.

## Water Levels and Groundwater Flow

The TWDB has collected water-level information on specific wells in the Salt Basin area as far back as the 1950's. Water levels in the Salt Basin generally follow the topographic relief of the basin. The highest water levels are in the southern part of the basin, where the land surface is correspondingly high. Water levels in wells along the basin margins are generally higher that those toward the basin center. Water-level maps made from data collected by the TWDB between 1992 and1994 show that groundwater generally flows from the northern and southern ends of the basin into the central part of the basin (figs. 17-3, 17-4). Most wells in the Salt Flat area penetrate the basin fill and are completed in the underlying Capitan or Delaware Mountain Formations. Water levels in the Capitan Formation have decreased about 10 to 20 ft in the last 40 yr of measurement (fig. 17-5).

Water levels in Wild Horse Flat show that water is relatively shallow and groundwater flow is to the south (fig. 17-3). Water levels decline from 3,590 ft above mean sea level (amsl) to 3,550 ft amsl. At the midpoint of Wild Horse Flat, there appears to be a slight mounding of groundwater that forms a groundwater divide where the basin fill is bottlenecked between the Baylor Mountains and the Apache Mountains. Groundwater in the southern part of Wild Horse Flat forms a slight depression, were water levels only range about 15 ft, from 3,540 ft amsl to 3,525 ft amsl. Hydrographs from wells in Wild Horse Flat show an overall decline in water levels of about 30 ft since 1950. Groundwater also flows from Wild Horse Flat southeast into Michigan Flat. Water levels in Michigan Flat are also slightly depressed on the south and eastern sides of the basin, indicating either a shallow cone of depression or that groundwater is flowing in a easterly direction out of Michigan Flat and into the Cretaceous and Permian formations that form the basin walls.

Water levels in the southern portion of the Salt Basin indicate that groundwater flows from Lobo Flat at an elevation of 4,200 ft amsl north to Ryan Flat at an elevation of 3,600 ft amsl (fig. 17-4). From Ryan Flat, groundwater flows into Wild Horse Basin. Hydrographs in both Ryan Flat and Lobo Flat indicate that there has been a slight rise in water levels in the past 20 to 50 yr (fig. 17-5). One well shows an increase of about 20 ft in Ryan Flat since 1950, and a second well in Lobo flat shows about a 10-ft rise in water levels since 1978.

Water levels in the Salt Basin appear not to be declining as much during the last decade as they did during the 1950's through the 1980's. This is partly a result of decreased pumping due to the decline in irrigation since the 1980's.

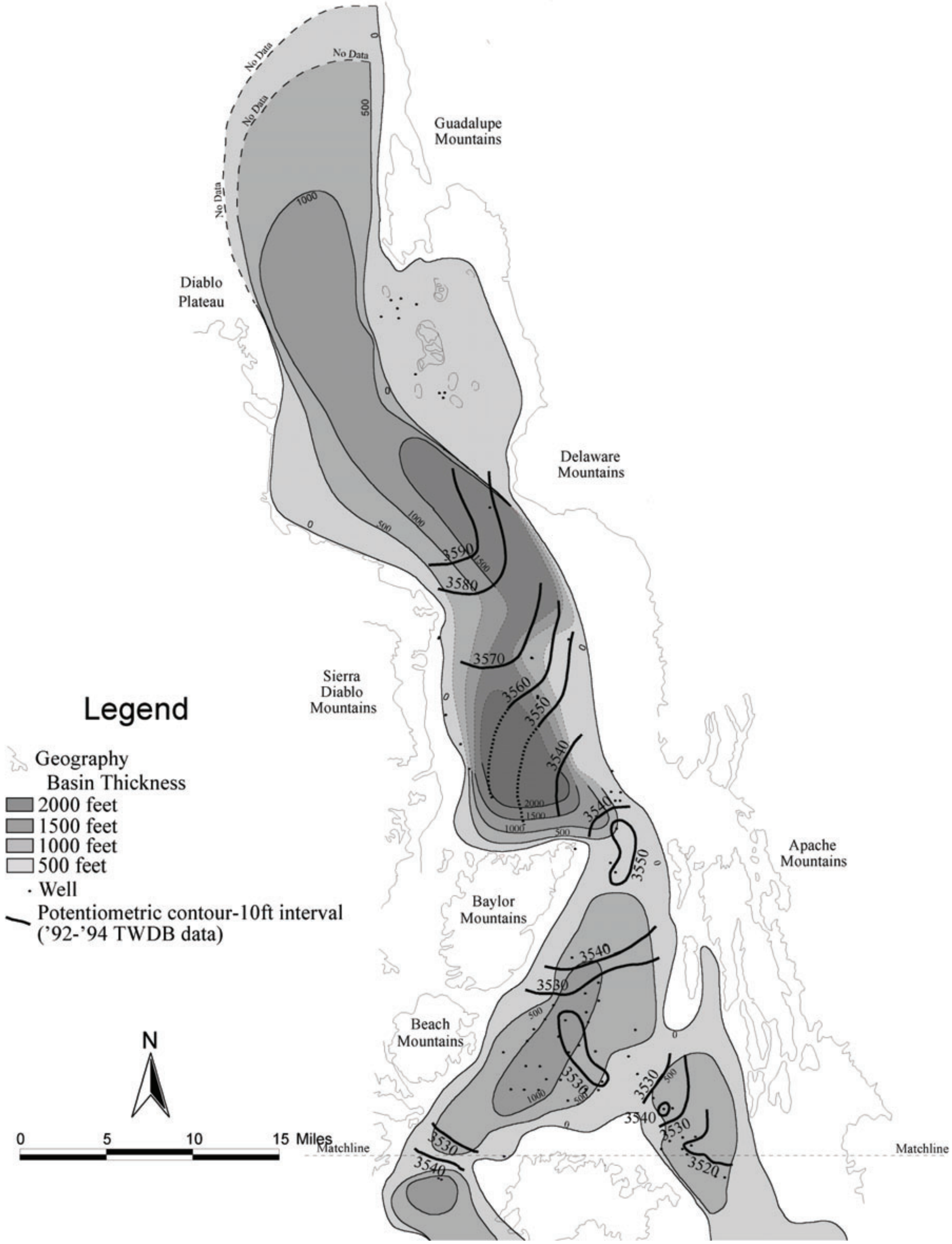


Figure 17-3: Thinkness of basin-fill and water-level contours in the northern Salt Basin (interpretation of basin fill modified from Gates and others, 1980).

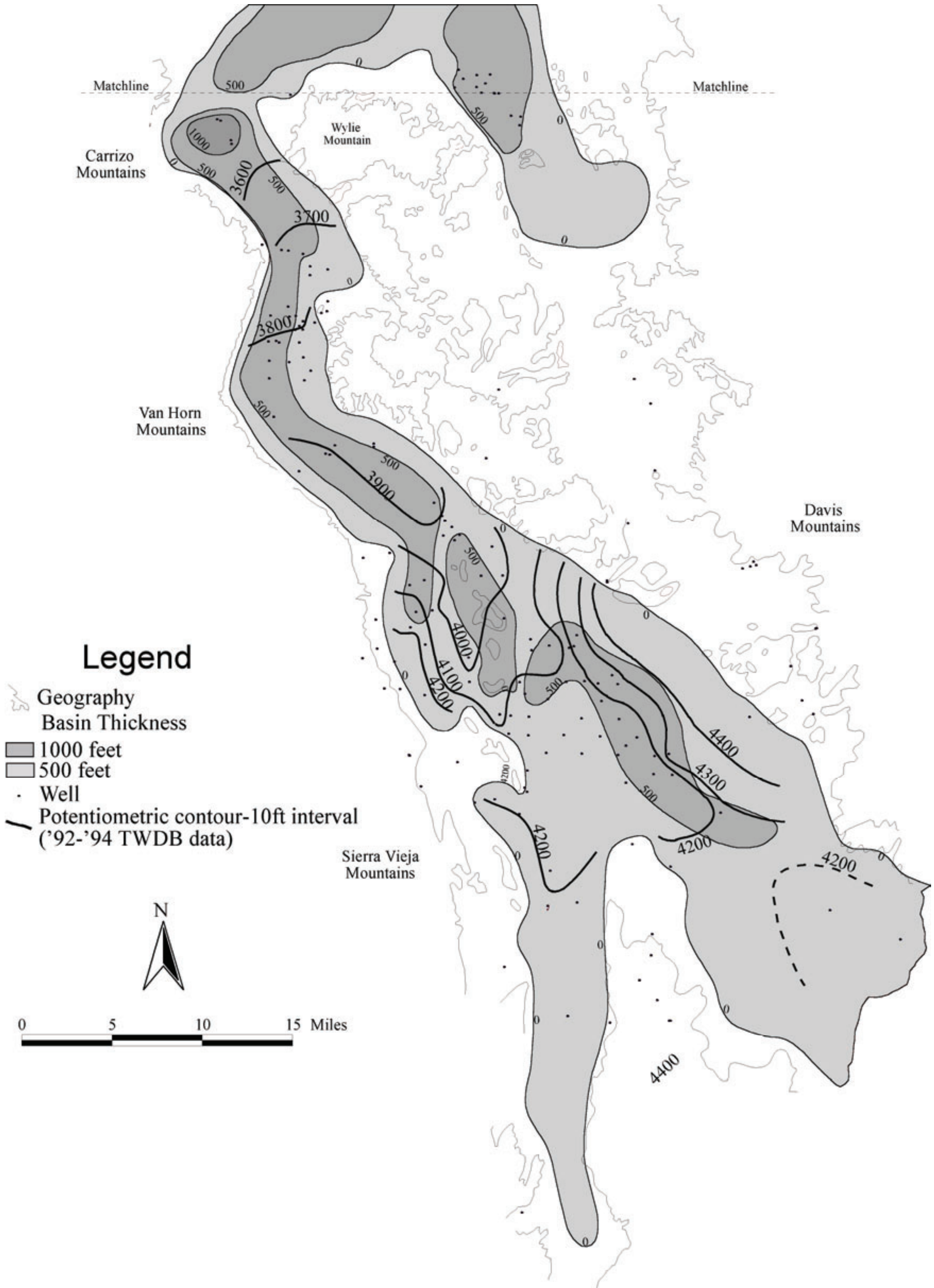


Figure 17-4: Depth of basin-fill and water-level contours in southern Salt Basin (interpretation of basin fill modified from Gates and others, 1980).

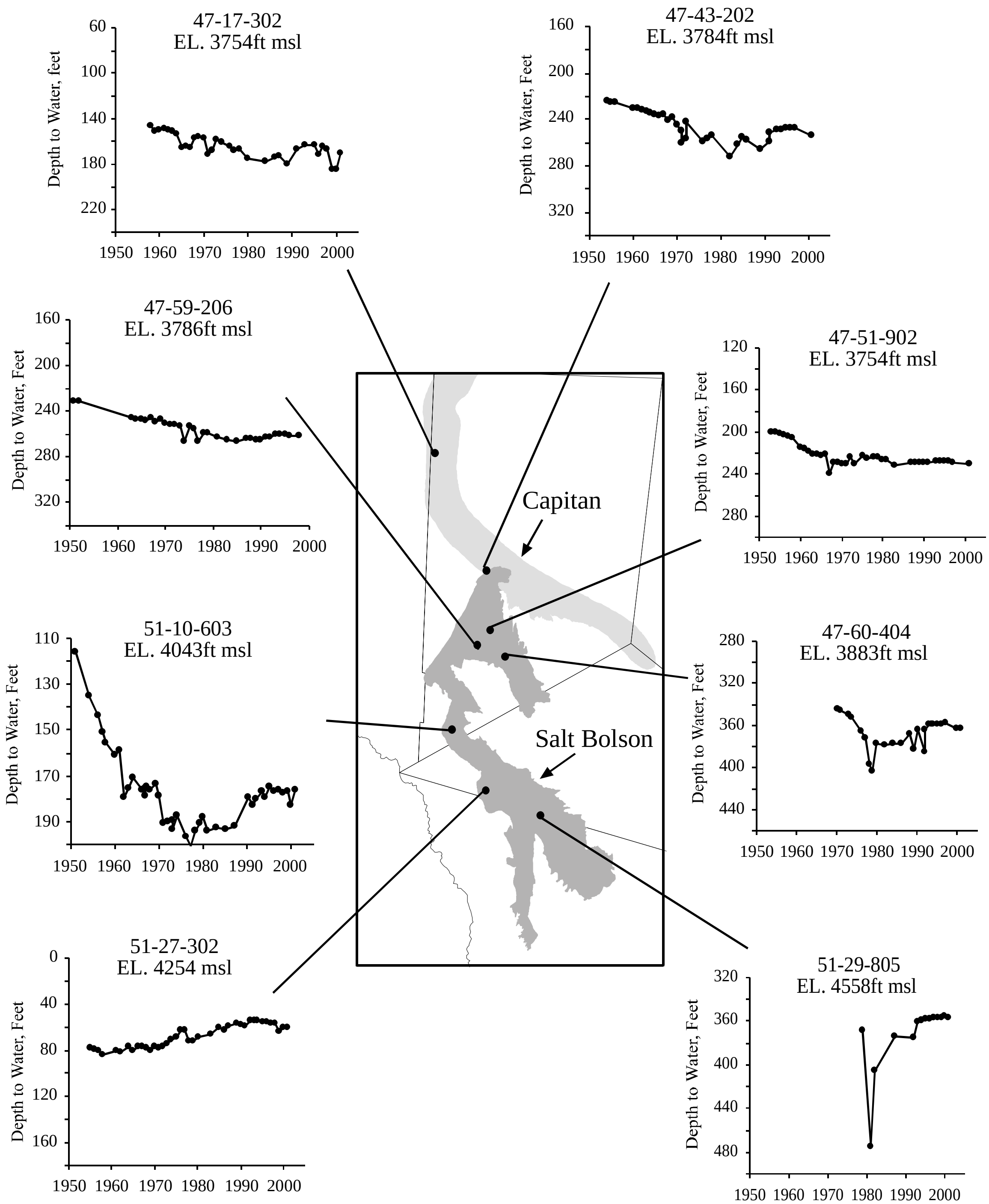


Figure 17-5: Hydrographs from various wells in the Salt Basin area. The Salt Bolson refers the bolson aquifers in the Salt Basin.

## Hydraulic Properties

Hydraulic properties of the aquifer in the Salt Basin vary considerably, depending on the local depositional character of the basin fill. Wells with high transmissivity values in the Salt Flat area are almost all completed partially or entirely in the underlying Capitan Formation. The highest transmissivity of any well in the Salt Basin occurs in well 47-09-207, which is completed in the Capitan Formation with a value of 80,000 $ft^2/d$. Wells in the Wild Horse Flat and Michigan Flat area tend to have higher transmissivity values when they are completed in the underlying Cretaceous formations than solely in the basin fill. In the southern portion of the Salt Basin, wells completed solely in the basin fill tend to have higher transmissivity values than wells that are also completed in the underlying volcanics or solely in volcanic formations (table 17-2).

## Water Quality

Water quality in the Salt Basin ranges from very saline in the northern part of the basin to fresh in the southern part. The basin fill in the Salt Flat area is thought to have little fresh or saline water, with the exception of the shallow groundwater associated with the salt playas (Gates and others, 1980). Total dissolved solids (TDS) of 3,000 to 6,000 mg/L are not uncommon in this area. With TDS values between 1,500 and 2,500 mg/L, well 47-34-603 (fig. 17-6) is representative of many wells in the Salt Flat area. Sodium, sulfate, and chloride levels are also elevated in these wells, rendering the water of limited use. Farther south in Wild Horse Flat, water quality improves slightly, where TDS is typically less than 1,000 mg/L. Wells 47-51-701, 47-59-101, and 47-60-707 are representative of average water-quality conditions for wells in Wild Horse Flat (fig. 17-6). Farther south in the Ryan Flat area, the aquifer has some of the freshest water in the Salt Basin area, with very low TDS sulfate, chloride and sodium values (fig. 17-6).

Overall, the wells in the Salt Basin show very little change in water quality over time. Only three of the six wells measured show any decrease in water quality over time, and these declines in water quality are very small (fig. 17-6). This lack of change in water quality indicates that there is very little impact from irrigation return-flow or migration of poorer quality water to the groundwater system in the Salt Basin at the present time. Although there have been several periods of intense pumping in the Salt Basin in the last 60 yr, the lack of change in water quality also indicates that up to now, there has been little impact from cross-formational flow from adjacent formations.

# Groundwater Availability

Well-completion strategies in the Salt Basin are governed by the need to access all available water when drilling. Therefore, many wells are completed in the basin fill and in basement formations where water is available. In the Salt Flat and northern Wild Horse Flat area, the groundwater in the basin fill is so poor that many wells are screened in both the underlying Capitan Formation and Delaware Mountain Group. In the southern portion of Wild Horse Flat and Michigan Flat, many wells are completed in the underlying

Table 17-2. Hydrologic data from the Salt Basin area.

| Location | State Well Number | Aquifer | Transmissivity $ft^2/d$ | Source |
|---|---|---|---|---|
| Salt Flat | 4709207 | Capitan Reef Complex and Associated Limestones | 80,000 | CSC |
| | 4717218 | Salt Bolson | 2,500 | CSC |
| | 4717317 | Capitan Reef Complex and Associated Limestones | 11,000 | AT* |
| | 4717321 | Salt Bolson and Capitan Reef Complex | 45,000 | CSC |
| | 4717903 | Capitan Limestone | 1,400 | AT* |
| | 4718402 | Delaware Mountain Formation or Group | 400 | AT* |
| | 4718404 | Salt Bolson and Delaware Mountain Group | 500 | CSC |
| | 4718707 | Salt Bolson and Delaware Mountain Group | 4,300 | CSC |
| Wild Horse Flat | 4734902 | Capitan Reef Complex and Associated Limestones | 10,000 | AT* |
| | 4743503 | Delaware Mountain Formation or Group | 1,100 | CSC |
| | 4751403 | Salt Bolson and Permian Rocks | 19,000 | AT* |
| | 4751807 | Salt Bolson | 4,100 | CSC |
| | 4752301 | Capitan Reef Complex and Associated Limestones | 500 | CSC |
| | 4752601 | Capitan Reef Complex and Associated Limestones | 11,000 | CSC |
| | 4752602 | Capitan Reef Complex and Associated Limestones | 2,000 | AT* |
| | 4758502 | Salt Bolson | 900 | CSC |
| | 4758505 | Salt Bolson | 1,600 | CSC |
| | 4758602 | Salt Bolson | 5,000 | AT* |
| | 4758602 | Salt Bolson | 6,300 | AT* |
| | 4759102 | Salt Bolson and Cretaceous Rocks | 6,000 | AT* |
| | 4759209 | Cretaceous System | 2,600 | AT* |
| Michigan Flat | 4759307 | Salt Bolson and Cretaceous Rocks | 1,900 | AT* |
| | 4759603 | Cretaceous System | 2,000 | CSC |
| | 4760404 | Salt Bolson | 1,000 | CSC |
| | 4760601 | Permian System | 30 | CSC |
| Lobo Flat | 5102906 | Salt Bolson | 8,600 | CSC |
| | 5102918 | Salt Bolson | 1,100 | CSC |
| | 5102923 | Salt Bolson | 4,900 | CSC |
| | 5102926 | Salt Bolson | 2,500 | AT* |
| | 5103702 | Salt Bolson | 6,400 | CSC |
| | 5103703 | Salt Bolson | 500 | CSC |
| | 5110306 | Salt Bolson | 1,500 | CSC |
| | 5110309 | Alluvium and Tertiary Volcanics | 5,800 | CSC |
| | 5110316 | Salt Bolson | 5,100 | CSC |
| | 5110317 | Alluvium and Tertiary Volcanics | 2,400 | CSC |
| | 5110321 | Salt Bolson | 6,900 | CSC |
| | 5110322 | Salt Bolson | 2,000 | CSC |
| | 5110328 | Salt Bolson | 4,800 | CSC |
| | 5110331 | Alluvium and Tertiary Volcanics | 4,200 | CSC |
| | 5110332 | Salt Bolson | 4,700 | CSC |
| | 5110603 | Salt Bolson | 3,000 | AT* |
| | 5110603 | Salt Bolson | 2,400 | CSC |
| | 5110624 | Salt Bolson | 380 | CSC |
| | 5111105 | Salt Bolson | 3,400 | CSC |
| | 5111106 | Salt Bolson | 1,600 | CSC |
| | 5114501 | Volcanics | 70 | CSC |
| | 5119104 | Salt Bolson | 3,000 | CSC |
| | 5119301 | Salt Bolson | 5,200 | CSC |
| | 5120403 | Salt Bolson | 800 | CSC |
| | 5120404 | Salt Bolson | 1,700 | CSC |
| Ryan Flat | 5127302 | Volcanics | 910 | CSC |
| | 5128303 | Salt Bolson | 1,900 | CSC |
| | 5128606 | Salt Bolson | 1,200 | CSC |
| | 5129104 | Salt Bolson | 30 | CSC |
| | 5129105 | Salt Bolson | 230 | CSC |
| | 5129403 | Salt Bolson | 2,000 | CSC |
| | 5128404 | Salt Bolson | 5,500 | CSC |
| | 5128406 | Salt Bolson | 3,000 | CSC |
| | 5128702 | Salt Bolson | 9,200 | CSC |
| | 5129704 | Salt Bolson | 1,900 | CSC |
| | 5129705 | Salt Bolson | 4,900 | CSC |
| | 5128701 | Salt Bolson | 10 | AT* |
| | 5136601 | Salt Bolson | 9,900 | AT* |

*AT - Aquifer Test *CSC - Calculated from Specific Capacity

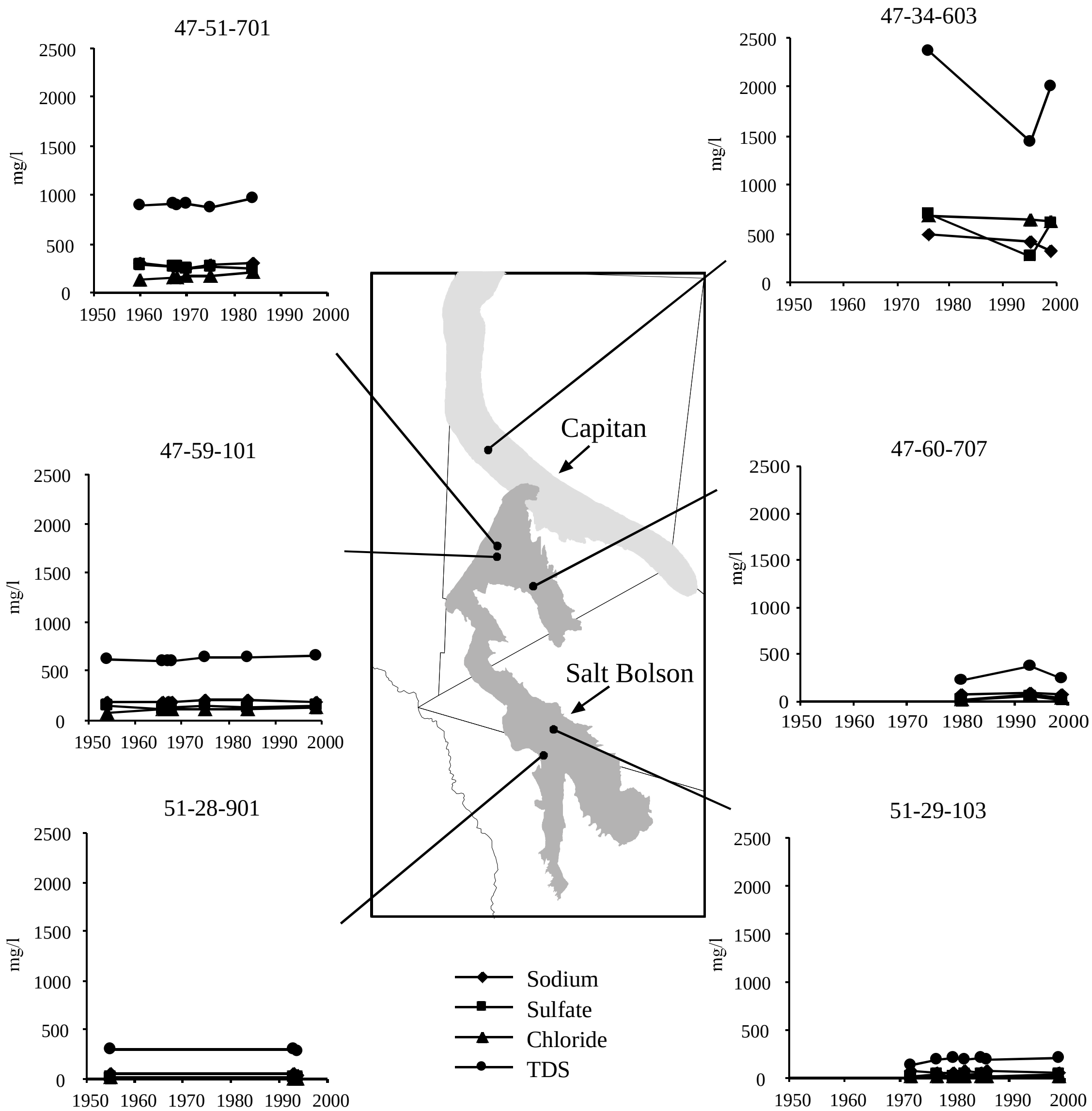


Figure 17-6: Water quality from the Salt Basin.

Cretaceous formations, as well as the basin fill. In the Lobo and Ryan Flat areas, some wells are completed in the basin fill and underlying Tertiary volcanic formations.

Gates and others (1980) investigated the availability of fresh and saline water in the West Texas Bolsons, including the Salt Basin. They concluded that the greatest amounts of fresh water were found in the Lobo and Ryan Flats area, with a combined volume of approximately 4.6 million acre-ft. However, they also noted that the development of groundwater from these basins would result in large water-level declines because of thelow recharge rate. Brown and Caldwell (2001) investigated the potential of developing 15,000 acre-ft/yr of groundwater from the West Texas Bolsons to pipe to El Paso. They noted that when the aquifer was pumped at 15, 000 acre-ft/yr between 1949 and the early 1980's, water levels declined up to 144 ft. Because pumping was greatly reduced in the 1980's, water levels have only recovered 30 percent (Brown and Caldwell, 2001). Brown and Caldwell (2001) concluded that water from Lobo and Ryan Flats may not be economically feasible owing to low recharge, declining water quality, the high cost of development, and limited availability of groundwater.

The Beldon Foundation is funding work on a model of the bolson aquifer in parts of Wild Horse, Michigan, and Lobo Flats (CCGCD, 2001). This model will be a useful tool to assess the possible impacts of increased pumping on water levels.

# Conclusions

Fresh to slightly saline water in the Salt Basin aquifer occurs in the basin fills beneath the Salt, Wild Horse, Michigan, Lobo, and Ryan Flats and is part of the West Texas Bolsons aquifer recognized by the TWDB. Basin fills in the Salt Basin consist of Tertiary and Quaternary alluvium, lacustrine sands, silts, muds, evaporate deposits, pyroclastic debris, lava flows, and tuffaceous deposits. Many of the wells in the Salt Basin also penetrate and produce water from underlying formations, including the Capitan Reef and Igneous aquifers. Because of the dry climate, geology, and topography, recharge is low and focused along basin boundaries. Most of the discharge from the aquifer is by pumping, discharge to the Salt Flats, and cross-formational flow. The aquifer generally has good well yields and good water quality in the southern part. Water levels have declined in response to pumping, although the rate of decline has slowed because of decreases in irrigation. There does not appear to be a substantial decline in water quality over time.

Studies suggest that there may be a considerable amount of fresh water in the bolson fills of Lobo and Ryan Flats. However, because of low recharge rates, water pumped from these aquifers will cause large water-level declines. Recent studies suggest that producing large amounts of water from these areas may not be economically feasible.

# Acknowledgments

The author would like to acknowledge and thank all the landowners, ranch foremen, well drillers, and others who allowed access to property, wells, and information. Their continued cooperation greatly contributes to the ability of researchers to develop a better understanding of groundwater in West Texas. TWDB employees Doug Coker, Frank Bilberry, John Derton, and Stephen Moore spent many days in the field during 1992 and 1993 collecting water-level and water-quality measurements in support of West Texas studies, and their efforts are greatly appreciated. The author also thanks John B. Ashworth and Darrel S. Peckham for their early work on this project and Robert E. Mace for his inciteful review of this paper.

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# Chapter 18

# The Diablo Plateau Aquifer

William F. Mullican III[1] and Robert E. Mace[1]

## Introduction

Although the Texas Water Development Board has delineated several aquifers in the West Texas area (e.g., Ashworth and Hopkins, 1995; TWDB, 1997; Mace, this volume), there still may be cause to consider adding at least one more aquifer to the mix: the Diablo Plateau aquifer. The Diablo Plateau aquifer coincides with the Diablo Plateau, a relatively flat area that lies between the Hueco Bolson to the west, the Salt Basin to the east, and several mountain ranges to the south, extending northward into New Mexico (fig. 18-1) (Muehlberger and Dickerson, 1989). The Diablo Plateau consists primarily of limestone: some of the same limestones that compose the prolific Bone Spring-Victorio Peak aquifer in the Dell City area. Studies in the late 1980's on siting a low-level radioactive waste disposal facility concluded that the hydrogeology of the Diablo Plateau precluded the area from being suitable for waste disposal because of the potential as a future water resource (Mullican and others, 1987; Kreitler and others, 1987, 1990). These studies found good-quality water, good well yields, and evidence of recent recharge over most of the aquifer.

The purpose of this paper is to summarize past work characterizing the hydrogeology of the Diablo Plateau and to suggest that the Diablo Plateau aquifer be further evaluated as a potentially significant water resource for this region of West Texas. Ultimately, this area may warrant future consideration and possible designation as a minor aquifer of the State.

## Climate

The Diablo Plateau area has a subtropical arid climate characterized by high mean temperatures with marked fluctuations over broad diurnal and annual ranges (minimum and maximum average annual temperatures are 45º and 81ºF, respectively) and low mean annual precipitation (10 inches/yr) with widely separated annual extremes (Kreitler and others, 1990). Precipitation occurs primarily during late summer and early autumn rainfall from thundershowers. Rainfall events are locally intense but short lived, and surface water is ephemeral because of consistently high evaporation rates. Mean annual lake-surface evaporation potential in the study area is approximately 83 inches (Larkin and Bomar, 1983). For 19 of the 31 yr from 1951 to 1981, Hudspeth, Culberson, El Paso,

[1] Texas Water Development Board

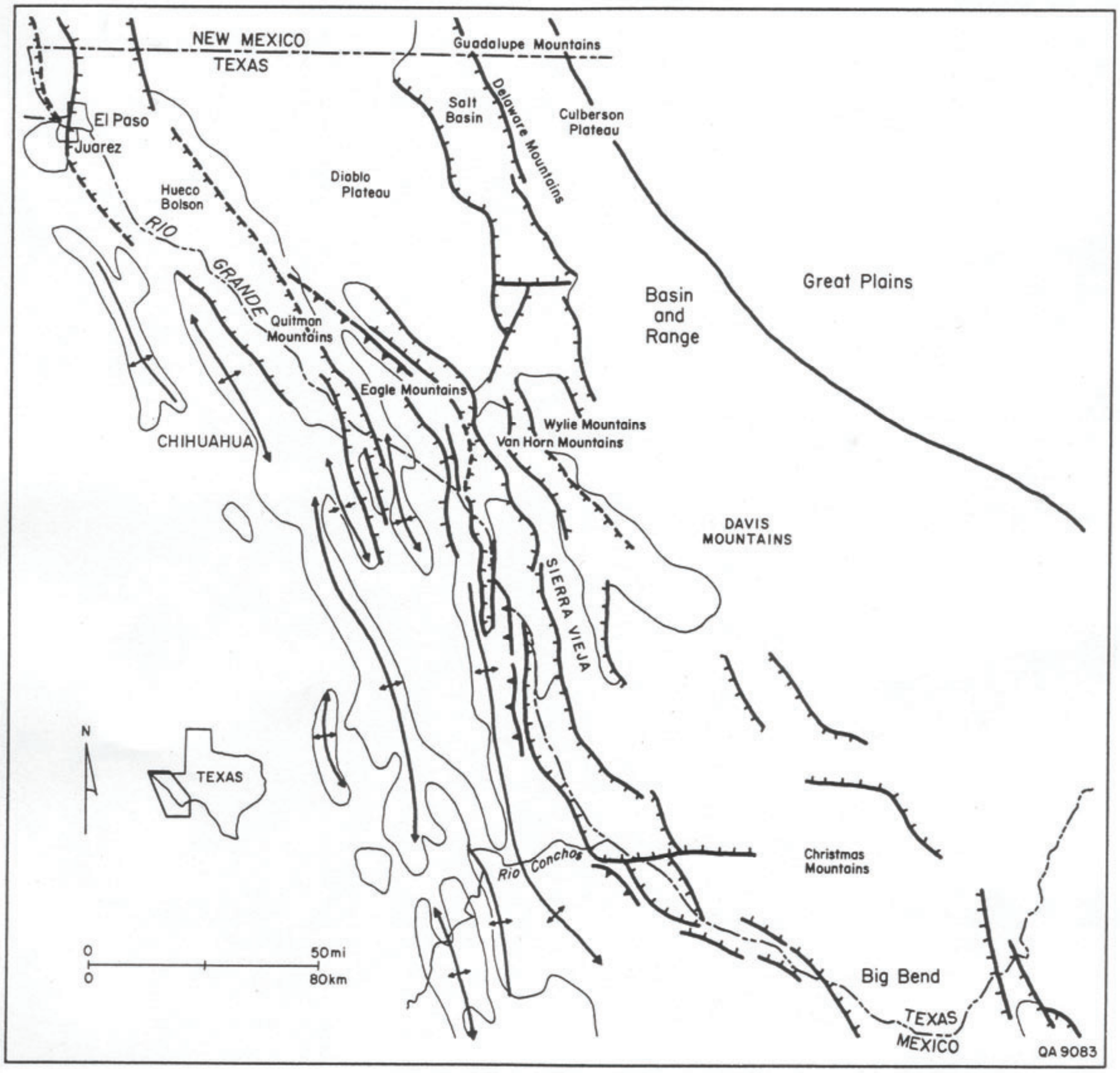


Figure 18-1: Tectonic and physiographic map showing the location of the Diablo Plateau (from Kreitler and others, 1990 [which was modified from Henry and Price, 1985]).

and adjacent counties recorded the lowest annual precipitation of any reporting stations in Texas (Bomar, 1995).

# Geologic Setting

The Diablo Plateau is in the southeastern part of the Basin and Range Province and is an uplifted, east-northeast-dipping homoclinal structure. The Diablo Plateau is bounded by major normal faults to the west at the Hueco Bolson and to the east at the Salt Basin and by several normal faults to the south near the Eagle Mountains (Barnes, 1983; Henry and

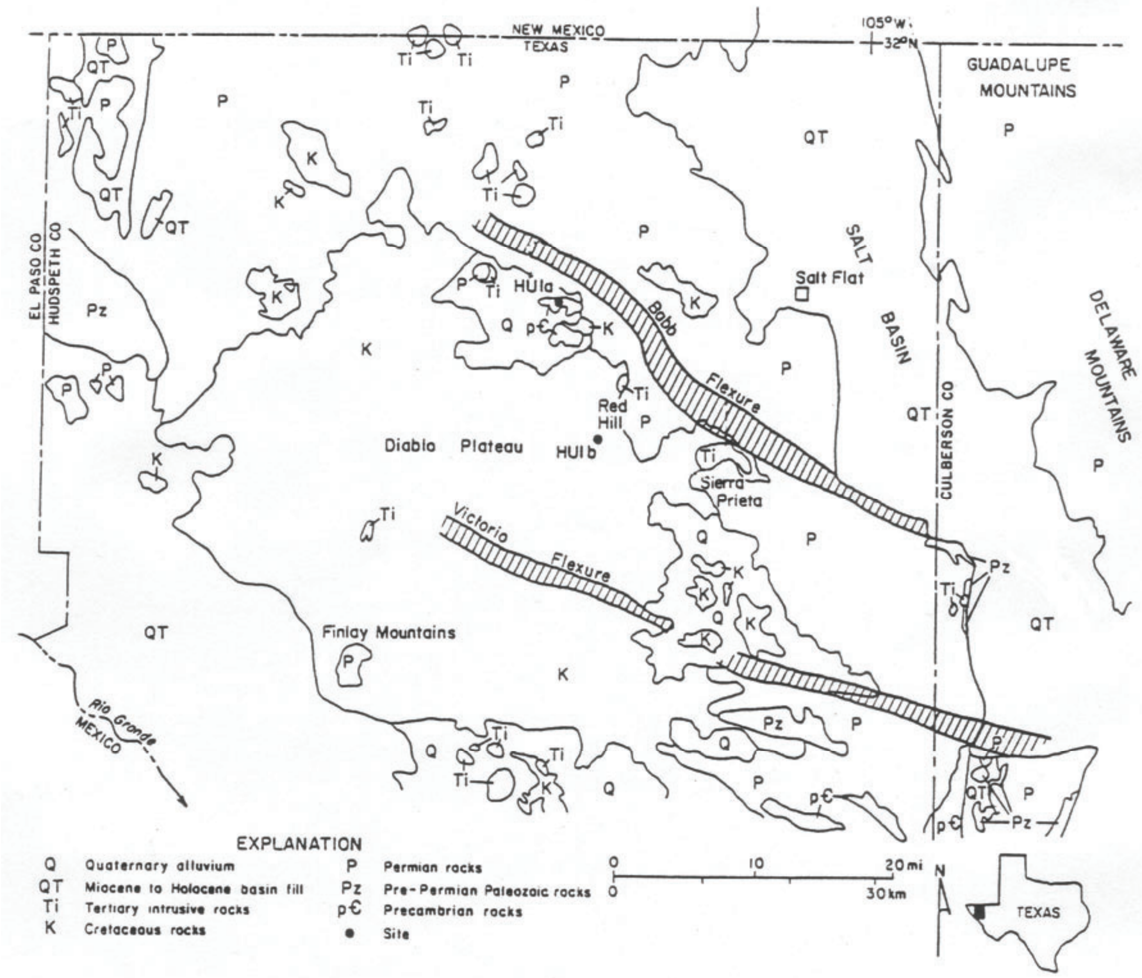


Figure 18-2: Geology of the Diablo Plateau area (modified from Kreitler and others, 1986 [with geology from Henry and Price, 1985]).

Price, 1985). The Diablo Plateau consists of Permian- and Cretaceous-aged limestones interbedded with sandstones and shales, with patches of Miocene to Holocene and Quaternary alluvium, occasional Tertiary intrusive rocks, and an area of Precambrian rhyolite and porphyry (fig. 18-2) (Henry and Price, 1985; Kreitler and others, 1986). Additional details on the geology in the area can be found in King (1965).

There are several structural features within the Diablo Plateau. The Babb flexure is a west-northwest-trending monocline about 1 to 2 mi wide with downward displacement of strata on the north side of the flexure (King, 1949; 1965) that may be traced about 40 mi northwestward from the Salt Basin (fig. 18-2). The flexure may be the Permian or post-Permian expression of a major pre-Permian strike-slip fault (Hodges, 1975). Farther to the south is the Victorio flexure (fig. 18-2). Fractures in the outcrop are associated with the flexures, Tertiary intrusions, and other faulting in the area.

# Hydrogeology

Water-level information suggests that there are two aquifers in the Diablo Plateau (Kreitler and others, 1986, 1990; see next section). These aquifers appear to correspond to the geology in the area: one aquifer is located in the Cretaceous rocks on the southwestern part of the plateau and another is located, at land surface, in the Permian rocks on the northern and northeastern part of the plateau. The aquifer in the Permian rocks underlies the aquifer in the Cretaceous rocks. However, the nature of the aquifer in the Permian rocks beneath the aquifer in the Cretaceous rocks is not known in great detail. Wells close to each other but drilled at different depths support two aquifers because of considerably different water levels (Kreitler and others, 1986). Both aquifers are primarily unconfined, although the aquifer in the Cretaceous rocks is locally perched confined to semiconfined (Kreitler and others, 1990). The aquifer in the Permian rocks is most likely confined beneath the Cretaceous rocks.

## Water Levels and Groundwater Flow

Water levels show that the Diablo Plateau aquifer is laterally connected to a number of aquifers in the area. The Cretaceous part of the aquifer is hydraulically connected to the Hueco Bolson aquifer (Mullican and Senger, 1990, 1992) to the west and to the Salt Basin and the Bone Spring–Victorio Peak aquifer in the Dell Valley area to the east (Peckham, 1963; Young, 1975; Kreitler and others, 1990; Mayer, 1995; Ashworth, this volume). The Bone Spring and Victorio Peak Formations or their equivalents are also part of the Diablo Plateau aquifer.

Depth to water in the aquifer can range from less than 5 ft  to more than 800 ft. The fresh-water part of the aquifer may be quite thick: the U.S. Soil Conservation Service drilled a borehole to 1,800 ft in the Dell City irrigation district on the northeastern side of the plateau and never crossed the base of the fresh/brackish water (Logan, personal communication, 1986).

Water levels show that there is a mound of water in the part of the Diablo Plateau aquifer south of Highway 62/180 corresponding to a local topographic high (fig. 18-3). Groundwater flows outward from this high to the southwest toward the Hueco Bolson, to the northeast toward the Salt Basin, and to the southeast toward the Finlay Mountains and northwest Eagle Flats (fig. 18-3). Limited information also suggests that a component of groundwater flows to the north (fig. 18-3). Groundwater flow north of Highway 62/180 generally flows eastward toward the Dell City area (Mayer, 1995).

Most of the water flows down the structural dip of the monocline toward the northeast, with only a minor amount of water flowing into the Hueco Bolson (Mullican and others, 1987; Kreitler and others, 1990). Hydraulic gradients are higher in the central Cretaceous part of the plateau and much lower along the Hueco Bolson and in the Permian part of the plateau.

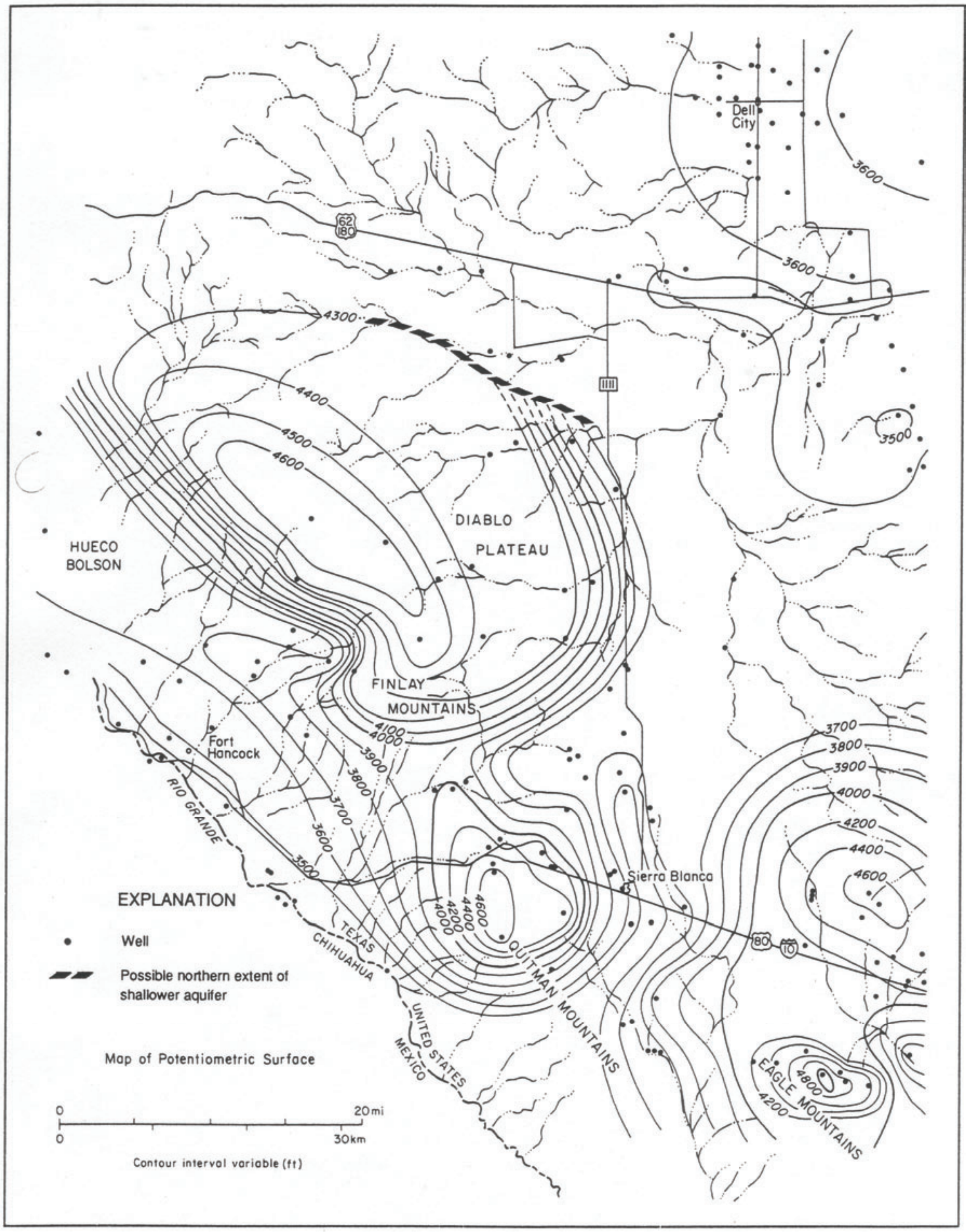


Figure 18-3: Potentiometric surface map of the Diablo Plateau area (from Kreitler and others, 1990).

Kreitler and others (1986, 1990) reported discontinuities in the potentiometric surface and suggested changes in hydraulic conductivity to partly explain the discontinuity (fig. 18-3). We think that the geology and topography can help explain the differences, acknowledging that the permeability of the Permian rocks is likely higher than the permeability of the Cretaceous rocks.

## Hydraulic Properties

The limestones of the Diablo Plateau may have the ability to transmit large amounts of water. Wells in the Bone Spring-Victorio Peak aquifer in the Dell City area have produced about 98,500 acre-ft/yr for 30 yr, with only about 33 ft of drawdown (Kreitler and others, 1990) from similar formations. The high production of the aquifer in this area is primarily due to fractures caused by faulting and subsequent dissolution of the host limestones. Well production is much greater in and near fracture zones than away from these zones. Individual wells located by lineament analysis can produce 2,000 to 3,000 gpm. Using aerial photography to locate areas of intense fractures, the U.S. Soil Conservation Service has successfully located 11 of 12 floodwater injection wells. Only 44 percent of the wells first drilled in the Dell City area were considered successful (Scalapino, 1950). In many cases, one well could produce greater than 2,000 gpm, while a well only 100 ft away would produce less than 100 gpm.

Specific capacity of the Bone Spring-Victorio Peak aquifer in the Dell City area ranges from 5 to 64 gpm/ft (Peckham, 1963). Using the Thomasson and others (1960, C = 1.2) approach to estimate transmissivity from specific capacity, these specific-capacity values correspond to transmissivities of 1,200 to 15,000 $ft^2$/d. Using the Bone Spring-Victorio Peak aquifer, similar transmissivities may be attainable in the Permian part of the Diablo Plateau aquifer. Mullican and others (1987) and Kreitler and others (1987, 1990) reported that in a majority of the pump tests conducted on wells completed in the Diablo Plateau aquifer, a majority were indicative of fracture flow. Several wells recently drilled and tested in the Diablo Plateau aquifer in northwestern Hudspeth County can produce 40 to 300 gpm for 48 h with no drawdown (LBG-Guyton Associates, 2001). Although the aquifer is not extensively used today, it has the potential to produce large volumes of fresh water.

## Recharge

Recharge occurs over the entire ~2, 900 $mi^2$ catchment area of the Diablo Plateau, as shown by the occurrence of tritium in nearly every well sampled on the plateau (fig. 18-4) (Mullican and others, 1987; Kreitler and others, 1990) (tritium is a relatively short lived radioisotope that suggests recent [<50 yr] recharge; see Scanlon and others, this volume, for a discussion on tritium as a tracer of recharge). This is in contrast to many of the bolson aquifers, where recharge is focused along mountain fronts (e.g., Darling, 1997, this volume; Scanlon and others, this volume). Most recharge probably occurs during flooding of the arroyos that traverse the Diablo Plateau. Chloride concentrations are significantly lower in soils in the arroyo than in soils between the arroyos, suggesting that the arroyos recharge at a much greater rate (Mullican and others, 1987; Kreitler and

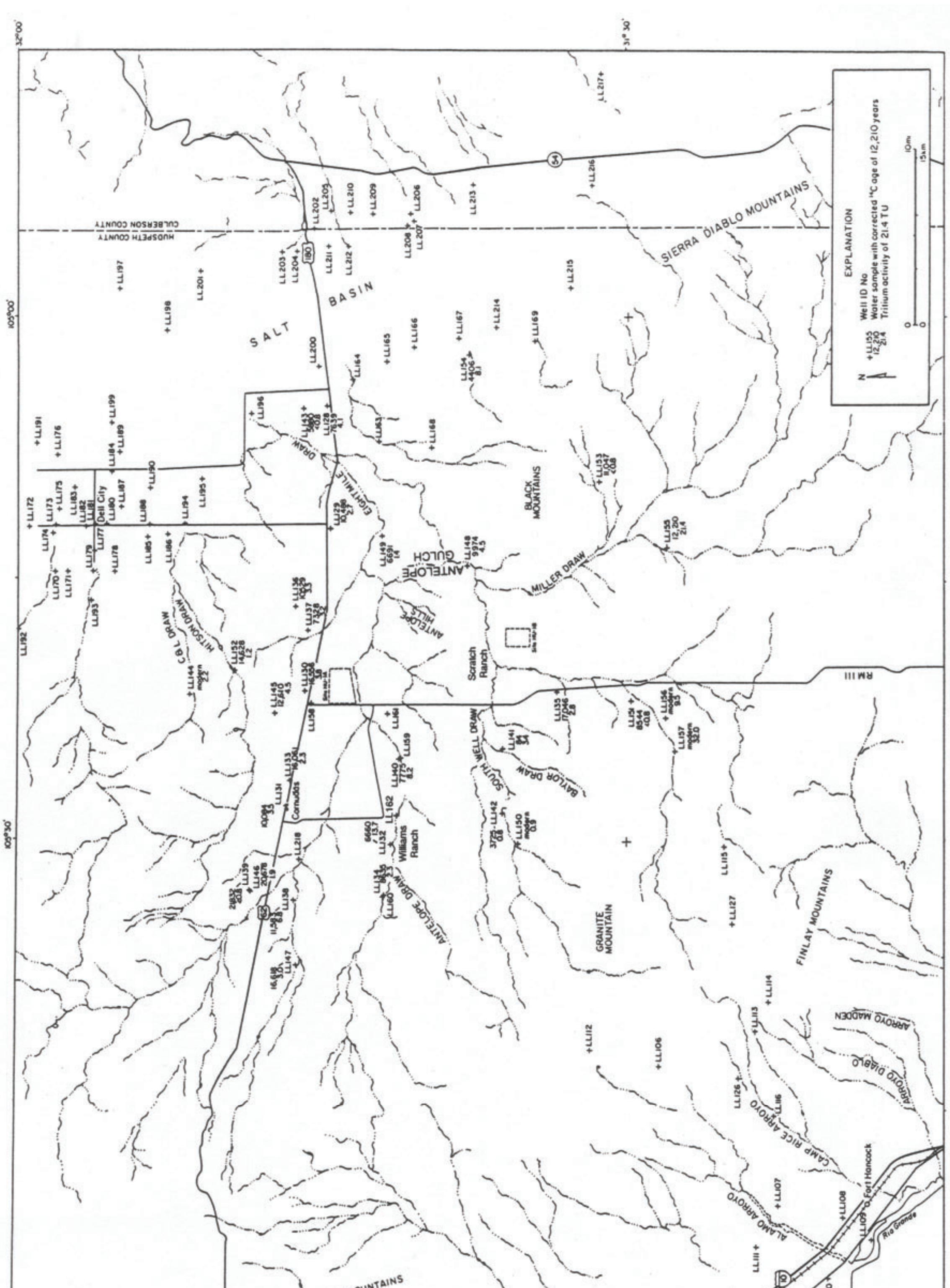


Figure 18-4: Areal distribution of tritium in water wells in the Diablo Plateau aquifer (from Kreitler and others, 1986).

others, 1987, 1990). Fractures, typically concentrated in arroyos, permit surface water to move rapidly through the thick unsaturated section. Peckham (1963) noted that the Bone Spring-Victorio Peak aquifer is partly fed by recharge in the Diablo Plateau to the west. To our knowledge, no one has estimated total recharge to the Diablo Plateau aquifer.

## Discharge

Based on the potentiometric surface map, it has been determined that most of the groundwater ultimately discharges naturally from the Diablo Plateau aquifer by evaporation and by interbasin flow. Groundwater flows from the Diablo Plateau aquifer into the Salt Basin. In the topographic low between the plateau and the Guadalupe and Delaware Mountains (the Salt Basin), the water table in the Salt Basin approaches the land surface (<3 ft depth to water), and large amounts of groundwater are evaporated. This evaporation precipitates gypsum, halite, and carbonates (Chapman, 1984; Boyd and Kreitler, 1986; Chapman and Kreitler, 1990). Gypsum may also be precipitating and reducing the permeability of sediments in the Salt Basin (Kreitler and others, 1990). The Diablo Plateau aquifer is thought to be the primary source of water to the Salt Basin. A minor portion of the groundwater in the Diablo Plateau flow to the south-southwest to ultimately discharge through cross-formational flow into the Hueco Bolson aquifer and ultimately the Rio Grande.

Groundwater may also discharge from the Diablo Plateau aquifer by interbasin flow beneath the gypsum flats of the Salt Basin to the south through Permian carbonates (Nielson and Sharp, 1985; Kreitler and others, 1990). This interbasin flow would eventually discharge to Balmorhea Springs or the Cenozoic Pecos Alluvium in Pecos County. Evidence of this is (Kreitler and others, 1990) (1) the absence of springs along the western edge of the Salt Basin, (2) the apparent restriction of flow in the Salt Basin due to limited thickness (3,280 ft, Veldhuis and Keller, 1980) and low permeability, and (3) the potential for a connection between the limestone of the Diablo Plateau and the limestones beneath the Salt Basin. Water levels in the Salt Basin suggest that water may flow to the south and to the east toward Balmorhea Springs. The $Ca\text{-}SO_4$ composition of the spring water supports the existence of such a large regional flow system.

Groundwater is also discharged from the aquifer by pumping. There is substantial pumping in the Dell City area, but much less in the rest of the Diablo Plateau.

## Water Quality

Groundwater from the Diablo Plateau aquifer ranges from a $Ca\text{-}HCO_3$ composition in the area of the groundwater divide between the Hueco Bolson and the Diablo Plateau to a $Ca\text{-}SO_4$ to a $Na\text{-}SO_4$ composition along the flow path to the Salt Basin. Water quality is generally fresh to brackish with total dissolved solids ranging from 715 to 3, 803 mg/L. Freshwater in the Diablo Plateau aquifer is generally restricted to Cretaceous rocks although freshwater is found in the Permian section in the more upgradient area of the aquifer. Water from the Cretaceous part of the aquifer has elevated $NO_3$, probably from animal waste or septic heads (Kreitler and others, 1986).

# Conclusions

The need for additional water resources in West Texas has been clearly established by many recent water-supply planning efforts. The rocks of the Diablo Plateau clearly warrant further evaluation as a potential water resource for West Texas. The Diablo Plateau aquifer has high well yields, good water quality, and is actively recharged. The aquifer consists primarily of Cretaceous and Permian limestones. Water levels indicate that the aquifer is laterally connected to neighboring aquifers, including the Hueco Bolson, Bone Spring-Victorio Peak, and Salt Basin aquifers. Water levels also indicate that there are two hydraulically distinct parts to the aquifer: one part in Cretaceous rocks and another in Permian rocks. Water quality is good, especially in the Cretaceous part of the aquifer, although nitrates might locally be a concern. Potential well yields in the aquifer are promising with wells in the Permian part of the aquifer producing as much as 300 gom without any measurable drawdown. Well yields are affected by faulting with much higher yields coming from wells that intersect fractures. The aquifer is widely and actively recharged over the entire Diablo Plateau. Most of the water that recharges the Diablo Plateau discharges to the Salt Basin with a lesser amount discharging to the Hueco Bolson. There is some evidence to suggest that water that originates on the Diablo Plateau discharges as far away as the springs in Balmorhea and into the Cenozoic Pecos Alluvium aquifer.

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# Chapter 19

# Aquifers of West Texas Field Trip Guide

Edward S. Angle

## Stop 1: Sul Ross University Center Parking Lot—Start and End of the Field Trip

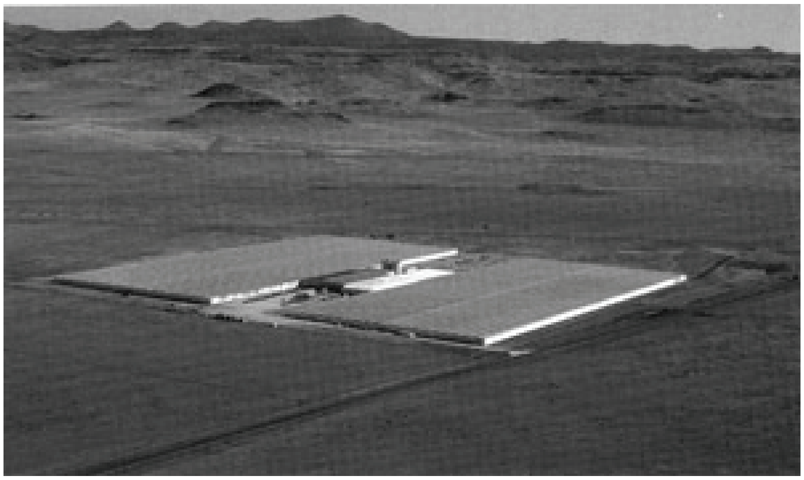
Sul Ross State University came into existence under authorization by act of the Thirty-Fifth Legislature in 1917 and was named for Lawrence Sullivan Ross, Texas governor from 1887 to 1891 (Tyler and others, 1996a). The university, originally established to train and certify teachers, currently offers a broad range of studies and promotes scientific research in biology, geology, and range animal science with emphasis on Chihuahuan Desert studies (Tyler and others, 1996a). For the 1999-2000 Fall and Spring semesters there were 4,157 enrollments and 115 faculty. The governing body is the Board of Regents of the Texas State University System. During fiscal year 1999, the university had $30.8 million in revenues and $29 million in expenditures (Sul Ross State University, 2000).

## Stop 2: Kokernot Springs–West Texas Springs: Igneous Aquifers

The Kokernot Springs, now dry, are located at the Kokernot Lodge inside the city limits of Alpine, Texas. The springs were originally known as Charo de Alsate, named after a powerful Apache war chief. Later the springs were called Burgess Springs or Burgess Water Hole after John Burgess, a cattle driver in the 1860's. The springs served as a water supply for countless generations of native peoples and later to many Spanish explorers, including de Vaca, de Espejo, and Mendoza.

In October of 1929, flow at Kokernot Springs was recorded at 222 gallons per minute (gpm) and later in 1947, at 396 gpm. As a result of well development in the Alpine area, the springs ceased to flow in 1950 (Brune, 1981).

## Stop 3: Village Farms–Commercial Application of Groundwater



This facility is one of several Village Farms greenhouses in Texas. This 41-acre greenhouse took 25 mo to develop and build, from ground breaking in March of 1996 to full-scale production in January of 1997 (Village Farms, 2001). The abundant wintertime light and cooler nighttime temperatures in the summer make this location ideal for greenhouse-tomato production.

There are nine wells associated with this Village Farms facility. The average well depth is 245 ft and the average well yield is about 200 gpm (Alan Standen, personal communication, 2001). Water usage was approximately 286 acre-ft/yr from three wells for 2000, according to Jeff Davis County UWCD (Janet Evans 2001, personal communication).

Some interesting facts about the Fort Davis Village Farms facility:

| | |
|---|---|
| ***Total Growing Area:*** | 41 acres (1,785,960 $ft^2$) |
| ***Packing and Support Facilities:*** | 76,230 $ft^2$ |
| ***Construction:*** | Aluminum, steel, and glass |
| ***BTU Capacity:*** | 96 million |
| ***Computer System:*** | Hoogendoorn Vitaco state-of-the-art system to control ventilation, shading, heating, fertigation, $CO_2$ levels, recirculation and pasteurization of nutrient feed |
| ***Number of Plants/Yield:*** | 416,000 plants, planted twice yearly, yield approximately 19 million pounds of tomatoes annually |
| ***Variety:*** | Beefsteak |

## Stop 4: Fort Davis National Park—West Texas Springs: Igneous Aquifers

The Fort Davis Spring is located on the southeast side of Fort Davis National Park. Nearby, crown-polished boulders indicate the use of the spring by early native peoples. The Spanish explorer Espejo stopped here in 1583 on his travels through West Texas (Brune, 1981). In the mid-1800's the area was known as Painted Comanche Camp because the Indians had painted pictures on many trees (Brune, 1981). From 1875 to 1883, the spring was used to supply drinking water to the fort. The men stationed at the fort suffered dysentery from unsanitary conditions that existed in the spring because, as the post surgeon Exra Woodruff stated, "it (the spring) is the resort of pigs…." A stonewall was erected around the spring to alleviate the water-quality problem. Sometime in the 1930's, the spring stopped flowing, possibly as a result of pumping in the vicinity of the fort.

There are approximately 141 springs that have been surveyed in the Davis Mountain area (Chastain-Howley, this volume), with an estimated spring flow of about 1.1 million gallons per day (Hart, 1992). While accurate numbers of springs and spring flow for historic times do not exist, current records show that spring flow has declined as groundwater development has increased in the area.

## Stop 5: Balmorhea State Park—West Texas Springs: Edwards-Trinity Aquifer

The springs at Balmorhea also have a long history of use by all peoples frequenting West Texas. The springs are still quite popular, and they form the main attraction at Balmorhea State Park, providing excellent swimming for visitors. San Solomon Spring is the largest of the springs in the Balmorhea area. Other significant springs include Phantom Lake and Giffin Springs. All are considered artesian and mildly thermal (20º-23º C) (Kreitler and Sharp, 1990). These springs are also home to a number of unique species, the Comanche Pupfish being one that relies on the special habitat of the springs and associated wetlands. Spring flow at San Solomon Spring has been consistent, but nearby springs, such as Phantom Lake, have seen a decline in flow over the past decade. Phantom Lake Spring does not currently flow.

## Stop 6: Clayton Draw—The Rustler Aquifer

The Rustler aquifer, composed of the Rustler Formation, was deposited in Permian times in the Delaware Basin and consists of mostly limestone, dolomite, and gypsum beds. Groundwater occurs in the very permeable solution zones within the upper portion of the formation. Most all groundwater from the Rustler is very high in dissolved solids concentrations and is therefore not potable for human consumption. Heavy pumping in the 1950's resulted in significant drops in water levels (Boghici, this volume). However, a subsequent decline in pumping has allowed water levels to rebound.

## Stop 7: Kent General Store—Rest Stop

The town of Kent was founded in 1892 and was originally known as Antelope because of the great numbers of antelope found in the area. A post office was founded there in 1893 with John Charles Rickli as postmaster (Tyler and others, 1996b). By 1914, the town had four cattle breeders, a general store, and a population of 25. From 1924 until 1965, the population approximately doubled (Tyler and others, 1996b). At its peak in the late 1960's, there were 4 businesses and a population of 65. Currently there is only one business in Kent, the Kent General Store. Feel free to buy something and make a contribution to the economy of Kent.

## Stop 8: Salt Basin—Salt Basin Aquifer

The Salt Basin in Texas extends from the New Mexico-Texas State border in Culberson County to a point about 10 miles west-northwest of Marfa Texas. It is 140 miles long and 25 miles across at its widest point. The basin is subdivided into the Salt, Wild Horse, and Michigan Flats in the north and Lobo and Ryan Flats in the south. Interstate Highway 10 happens to split the basin exactly in half. The freshest water in the Salt Basin is found in Lobo and Ryan Flats, where there may be as much as 4.6 million acre-ft of water available (Gates and others, 1980). Recharge to the Salt Basin is low because of the high evaporation rates and low rainfall rates (Gates and others, 1980). Recharge occurs along the basin margins and as cross-formational flow. Historically, water levels have dropped dramatically in response to heavy pumping, so the viability of this resource is limited.

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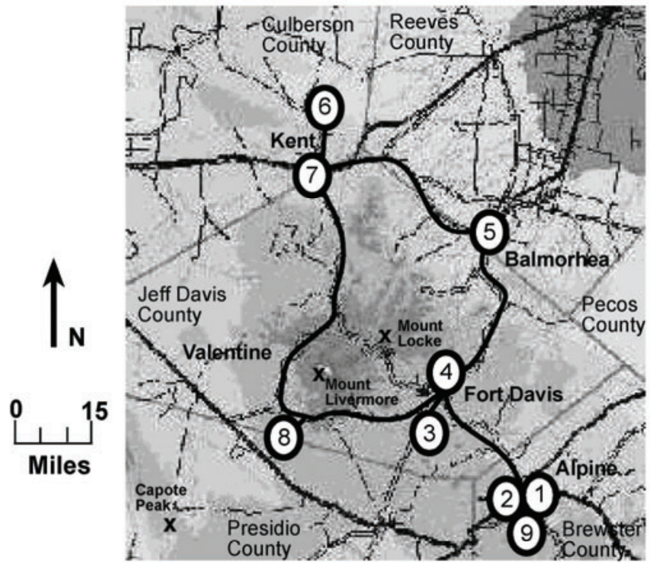
# Aquifers of West Texas
# Field Trip Map




| Stop | Description | Arrive | Depart | Mileage |
|---|---|---|---|---|
| 1 | Sul Ross University Center<br>Introduction | 8:00 | 8:10 | 0 |
| 2 | Kokernot Lodge<br>Visit the Kokernot Spring and discuss spring history, hydrographs of nearby wells, and Igneous aquifer in the Alpine area. | 8:15 | 8:40 | .06 |
| 3 | Village Farms<br>Visit the Village Farms hydroponic tomato farm. | 9:15 | 10:45 | 27.8 |
| 4 | Fort Davis National Park<br>Visit the Fort Davis Spring and discuss spring history, hydrographs of nearby wells, and Igneous aquifer in the Fort Davis area. | 10:55 | 11:20 | 33 |
| 5 | Balmorhea State Park<br>Lunch break | 12:00 | 12:30 | 64.5 |
| | San Solomon Springs<br>Discuss San Solomon Springs, Giffin and Phantom Lake Springs, and Edwards Trinity aquifer. | 12:30 | 12:45 | |
| | Walk Spring Area | 12:45 | 1:00 | |
| 6 | Clayton Draw (end of RR 2424)<br>Observe Rustler outcrops and discuss regional groundwater flow regime. | 1:50 | 2:20 | 105.5 |
| 7 | Kent General Store<br>Rest break. | 2:35 | 3:00 | 119 |
| 8 | Salt Basin (midway on RR 505)<br>Observe southern portion of Salt Basin and discuss Salt Basin hydrogeology. | 4:00 | 4:25 | 167.7 |
| 9 | Sul Ross University Center<br>Concluding comments. | 5:30 | 5:45 | 223.1 |

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| Number of Words: | 78,578 (approx.) |
| Number of Characters: | 447,898 (approx.) |