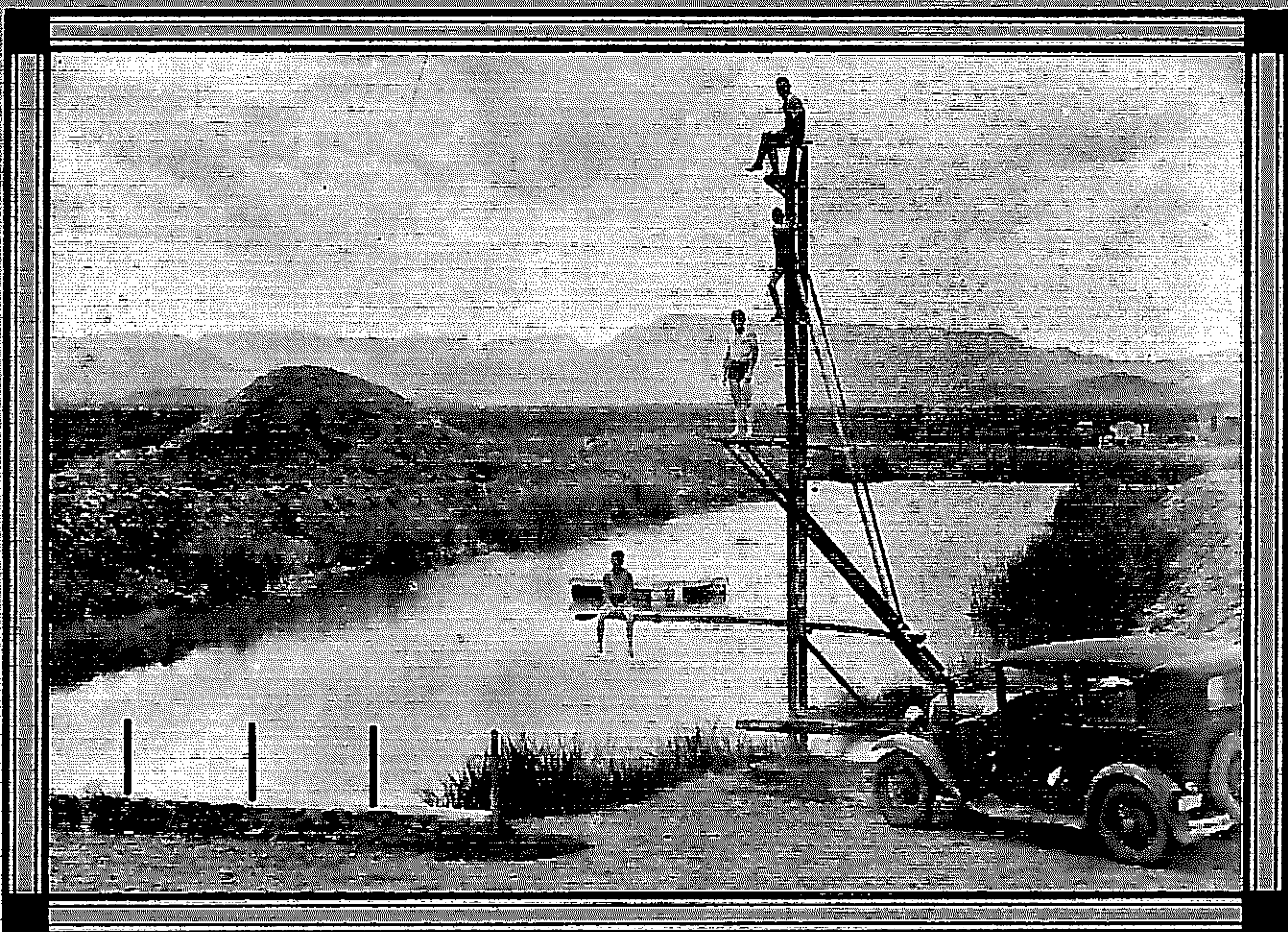
Guidebook 25

# Hydrogeology of Trans-Pecos Texas

Charles W. Kreitler and John M. Sharp, Jr.

Field Trip Leaders and Guidebook Editors



Bureau of Economic Geology • W. L. Fisher, Director

The University of Texas at Austin • Austin, Texas 78713

1990

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# HYDROGEOLOGY OF TRANS-PECOS TEXAS

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Contributors

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with selected reprints by

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Prepared for the 1990 Annual Meeting of the
Geological Society of America
Dallas, Texas
October 29–November 1, 1990

**Bureau of Economic Geology • W. L. Fisher, Director**
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Cover:
One of the five best swimming holes in Texas.
San Solomon Spring with divers, during construction
of Balmorhea State Park, 1930's.
Photograph courtesy of Darrel Rhyne, Park Superintendent,
Balmorhea State Park, 1990.

# Contents

## Selected Reprints on the Hydrogeology of the Trans-Pecos Area

# Preface

This guidebook, *Hydrogeology of Trans-Pecos Texas*, grew out of our interest in the hydrogeology of desert environments. After we participated in Marty Mifflin and Jay Quade's 1988 field trip through eastern Nevada,* we realized that, although there were many similarities, the hydrogeology of Trans-Pecos Texas was quite different from what we had observed in Nevada. Thus the seed was planted for this field trip and guidebook. We hope that those who participate in this trip will likewise become intrigued with the hydrogeology of West Texas and will return to their own hydrologic environments, arid or otherwise, with some new approaches and concepts.

The guidebook is a compendium of information on the hydrogeology of Trans-Pecos Texas. It includes a road log for the 2-1/2-day trip, a list of references cited in the road log, six technical papers, and seven papers previously published in scattered symposia and other guidebooks. These papers provide greater detail than can be contained in the road log.

Much of the information for the road log was generated during four previous field trips (1984, 1986, 1988, and 1989) to the Trans-Pecos region, conducted in conjunction with the graduate course in Geology and Hydrology offered at the Department of Geological Sciences, The University of Texas at Austin. We appreciate the interest and formal and informal contributions of numerous students. Malcolm Ferris assisted in compiling the second- and third-day road logs. The 1989 Geology and Hydrology class served on the rehearsal field trip. Financial and logistical support of the Department of Geological Sciences, The University of Texas at Austin, for these field excursions is gratefully acknowledged.

Four of the technical papers (Gustavson; Mullican and Senger; Fisher and Mullican; Kreitler, Mullican, and Nativ) and the reprinted paper by Scanlon, Richter, Wang, and Mullican resulted from investigations conducted for the Texas Low-Level Radioactive Waste Disposal Authority. Detailed characterization of geologic and hydrologic processes at the principal study area could not have been made without the strong support and commitment of the Authority to conduct a thorough scientific investigation. We also thank the El Paso Geological Society, the New Mexico Geological Society, Incorporated, the West Texas Geological Society, Incorporated, the American Nuclear Society, the American Geophysical Union, the New Mexico Bureau of Mines and Mineral Resources, and the Texas Bureau of Economic Geology for their permission to reissue selected papers or excerpts.

We thank Alan Dutton, Bill Muehlberger, Bernd Richter, Rainer Senger, and Jeff Grigsby for their technical reviews of the guidebook. Manuscript preparation was by Lucille Harrell; word processing and typesetting were by Melissa Snell and Susan Lloyd under the supervision of Susann Doenges. Technical editing was by Tucker Hentz. Production editing was by Amanda Masterson, illustrations were prepared by the Cartography staff under the supervision of Richard Dillon, and publication design was by Margaret Evans.

Many thanks are extended to the Geological Society of America for including this field trip in the agenda for the annual GSA meeting, October 29–November 1, 1990, in Dallas, Texas.

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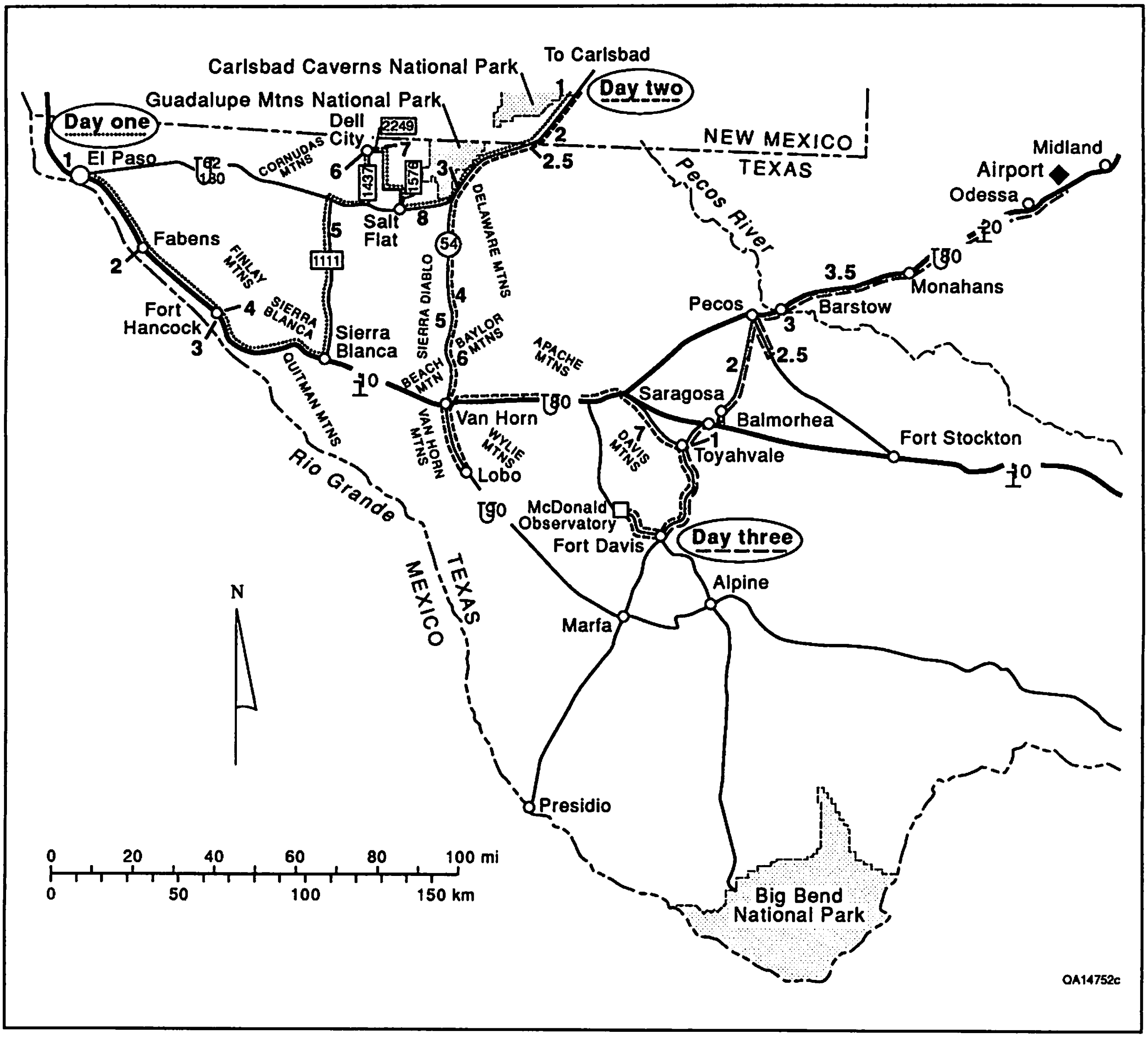
* Mifflin, M. D., and Quade, Jay, 1988, Paleohydrology and hydrology of the Carbonate Rock Province of the Great Basin (East-Central to Southern Nevada), *in* Holden, G. S., and Tafoya, R. E., eds., Geological Society of America Field Trip Guidebook: Colorado School of Mines Professional Contributions No. 12, p. 305–314.




Map of the field trip area, showing location of stops.

# Road Log

## First Day: El Paso, Texas–Rio Grande–Carlsbad, New Mexico

### Friday, October 26, 1990

El Paso, Texas. Assemble in motel parking lot.

### STOP 1.

Discussion of Hueco Bolson hydrogeology, El Paso–New Mexico water resources litigation, and water resources of the El Paso area (see **Ashworth**, this volume, p. 21*). The principal hydrostratigraphic units that we will be seeing during days 1 and 2 are shown in table 1.

Water laws in Texas and New Mexico are significantly different. Whereas New Mexico has, since statehood, always applied the appropriation doctrine to ground water, Texas uses the English doctrine of absolute ownership (riparian) for ground water and the appropriation doctrine for surface water. Texas landowners can capture as much ground water through pumpage as needed as long as water is not "wasted." In Texas the option does exist, however, to create underground water conservation districts that have the power to levy taxes and regulate ground-water withdrawals. Most such districts in Texas have not regulated ground water but have approached water management by voluntary controls and educational programs.

**Mileage**

**0.0** Set odometer to zero at intersection of Airway Boulevard and Interstate Highway 10 near El Paso Airport; drive east on I-10. We are in the Hueco Bolson.

**5.0** The Rio Grande alluvial floodplain can be seen on the right at a lower elevation. I-10 here is on top of old terrace deposits of the Rio Grande. Air pollution in the El Paso–Juarez area is common, as evidenced by the haze to the south and west.

The Hueco Bolson is bounded by the Franklin Mountains on the west, the Hueco Mountains on the east, the Tularosa Basin on the north, the Diablo Plateau, Quitman Mountains, and Malone Mountains on the north and east, and a series of Mexican mountain ranges across the Rio Grande on the south and west. The bolson is approximately 20 to 25 mi (30 to 40 km) wide and 90 mi (140 km) long. In the El Paso area, the Rio Grande has downcut some 250 ft (75 m) deep into the bolson sediments. Most of the fresh water is located in an irregular trough just east of, and subparallel to, the Franklin Mountains. A major recharge area is on the trough's west side, in the alluvial sediments shed from the Franklins. Although the Rio Grande is also an important source of recharge (Ashworth, this volume, p. 21), it is not a factor in the recharge of the bolson aquifers that have heads at elevations higher than the Rio Grande. Water quality in bolson sediments is commonly better than that in the Rio Grande alluvium. Irrigation return flows from farming in New Mexico and Texas have caused deterioration of Rio Grande water quality. In addition, the Rio Grande contains high fecal coliform and biologic oxygen demand (BOD) loads created in large part by sewage inflows from Mexico.

**23.8** Take exit 49 to Fabens.

### STOP 2 (brief) Fabens rest area.

**25.0** An alluvial fan from the arroyo spreads out over Recent alluvium. Such fans are loci for ground-water recharge from surface-water flows in the arroyos.

The undeveloped Fabens artesian system was discovered by two test holes, drilled in 1957 and 1959, which produced brackish water (with total dissolved solids [TDS] of 958–1,540 mg/L). The aquifer consists of alternating clays and sands overlain by a thick clay. The artesian section is between 1,280 and 1,909 ft (390 and 582 m) deep beneath land surface. The recharge area appears to be in the Sierra del Presidios southwest of the Rio Grande in Mexico. Approximately 400,000 to 800,000 acre-ft ($4.9 \times 10^8$ to $9.8 \times 10^8$ $m^3$) of water is estimated to be in storage (Gates and others, 1980), but the water contains excessive amounts of fluoride. Consequently, these

* References in **bold** print are included in this guidebook as either original articles or reprints.

Table 1. Hydrostratigraphic units, days 1 and 2. Modified from Gates and others (1980).

| Erathem | System | Hydrostratigraphic unit | Physical and lithologic characteristics | Water-bearing characteristics |
|---|---|---|---|---|
| CENOZOIC | Quaternary | Rio Grande alluvium and other alluvial systems | Gravel, sand, silt, and clay deposited by the Rio Grande and its tributaries; may be 200 ft (60 m) thick at some locations. | Supply moderate to large quantities of fresh to moderately saline water in the Rio Grande Valley in the Mesilla, Hueco, and Presidio Bolsons. Alluvium of tributary streams is unsaturated in many basins but supplies small amounts of fresh water for domestic and stock use in the Presidio Bolson and near the Rio Grande in other basins. |
| | Quaternary and Tertiary | Bolson deposits | Clay, silt, sand, and gravel deposited by the ancestral Rio Grande or streams in local to individual basins; commonly 1,000 to as much as 9,000 ft (300 to 2,200 m) thick (in the Hueco Bolson). | Principal aquifers in westernmost Texas; supply moderate to large quantities of fresh to slightly saline water in basin areas; contain moderately saline or poorer quality water at depth in the Hueco Bolson and in parts of the Hueco, Mesilla, and Presidio Bolsons and the Salt Basin, mostly in fine-grained lacustrine and alluvial deposits. |
| | Tertiary | Volcaniclastic and volcanic deposits, some filling bolsons | Reworked tuffs and alluvial deposits consisting almost exclusively of volcanic debris (volcanic clastics) interbedded with ash-fall tuffs and volcanic flows or ash-flow tuffs; up to 6,000 ft (1,800 m) thick at Ryan Flat. | Supply small to large quantities of fresh water in Ryan and Lobo Flats; probably occur in southeastern Presidio Bolson; permeable zones are probably most common in the uppermost 1,000 ft (300 m) and may include well-reworked tuff, well-sorted volcaniclastics, weathered zones above and below volcanic flows, and possibly fractured volcanic-flow rocks. |

| Erathem | System | Hydrostratigraphic unit | Physical and lithologic characteristics | Water-bearing characteristics |
|---|---|---|---|---|
| MESOZOIC | Cretaceous | Limestones, undifferentiated, but including the Campagrande Formation, Bluff Mesa Limestone, and Yucca Formation; and the Cox Sandstone | Limestone units include beds of marl, sandstone, conglomerate, siltstone, and shale, more than 5,000 ft (1,500 m) thick; Cox Sandstone is mostly quartz sandstone with some pebble conglomerate and siltstone, shale and limestone; very fine to medium-grained; commonly less than 200 ft (60 m) thick, but can be as thick as 700 ft (210 m). | Limestones supply small to moderate quantities of fresh to moderately saline water in the Sierra Blanca area; Cox Sandstone supplies small to moderate quantities of fresh to moderately saline water locally in the southeastern Hueco Bolson, the Sierra Blanca area, and southeastern Wildhorse Flat. |
| PALEOZOIC | Permian | Limestones, including the Capitan Limestone, the Goat Seep Limestone, and the Bone Spring and Victorio Peak Limestones, undifferentiated; and sandstones, including the Delaware Mountain Group | Capitan and Goat Seep Limestones are massive, thick-bedded reef limestone and dolomite; Capitan is 1,000 to 2,000 ft (300 to 600 m) thick in the Guadalupe Mountains and Beacon Hill area and up to 900 ft (220 m) thick in the Apache Mountains area; the Goat Seep is up to 1,200 ft (360 m) thick in the Guadalupe Mountains area; the Bone Spring and Victorio Peak Limestones are limestone and dolomite with sandstone and siltstone, aggregate thickness 1,800 to more than 3,000 ft; (540 to 900 m) the Delaware Mountain Group is sandstone and limestone with some siltstone; aggregate thickness is on the order of 3,000 ft (900 m). | Capitan and Goat Seep Limestones supply moderate to large quantities of fresh to slightly saline water in the Beacon Hill area, and the Capitan supplies moderate to large quantities of fresh and slightly saline water in the Apache Mountains area; the Bone Spring and Victorio Peak Limestones supply small to large quantities of slightly to moderately saline water in the Dell City area and the northeastern Diablo Plateau; the sandstones and limestone of the Delaware Mountain Group supply small quantities of slightly to moderately saline water along the eastern side of the northern Salt Basin and the foothills of the Delaware Mountains. |
| PRECAMBRIAN | — | Carrizo Mountain Formation and possibly Allamoore Formation | Carrizo Mountain Formation consists of meta-igneous rocks; Allamoore is limestone, conglomerate and metamorphic, volcanic, and igneous rocks. | Supply small quantities of fresh water in the Allamoore area; permeable zones probably are weathered or fractured rock. |

artesian waters will have to be mixed with other sources of low-fluoride water before they can be utilized.

**43.7** Cross Alamo Arroyo. Tributaries of the Rio Grande are important sites for ground-water recharge. Water quality near major arroyos is typically of better quality because of the localized recharge.

**47.0** Fort Hancock exit. Turn right and proceed southward to the Rio Grande.

## STOP 3. Rio Grande.

**50.2** Here we will view the mighty Rio Grande. Its floodplain is an important agricultural resource for both the United States and Mexico. Local crops include cotton, melons, onions, pecans, alfalfa, hay, and in the downstream Gulf of Mexico segment, citrus fruits.

The Rio Grande in this area is a relatively young stream, formed during the late Pliocene (**Gustavson**, this volume, p. 27). The Pecos River, which we will see on the third day of this trip, also underwent major channel shifts during the late Cenozoic. When the Rio Grande began eroding the Hueco and Presidio Bolsons, extensive terraces and pediments were formed through the erosion of tributary streams and stream capture by the Rio Grande. Figure 1 depicts the suggested origin of the terraces.

**0.0** Return to Fort Hancock and cross I-10. Proceed northward on gravel road. Reset odometer to zero at Fort Hancock interchange.

**6.5** Faultline scarp of Campo Grande fault.

The rise in topography across the caliche road is a faultline scarp of the northwest-trending Campo Grande fault, which formed in response to the Basin and Range extension that began 24 m.y. ago. Fault movement has been intermittent to the present. Maximum throws of 28 to 32 ft (8 to 10 m) have been measured across Pleistocene gravels. Maximum vertical offset during the last fault event was about 3 ft (1 m) (Collins and Raney, 1989). This fault may significantly affect regional ground-water flow from the Diablo Plateau (to the north) to the Rio Grande (see fig. 1, **Mullican and Senger,** this volume, p. 37).

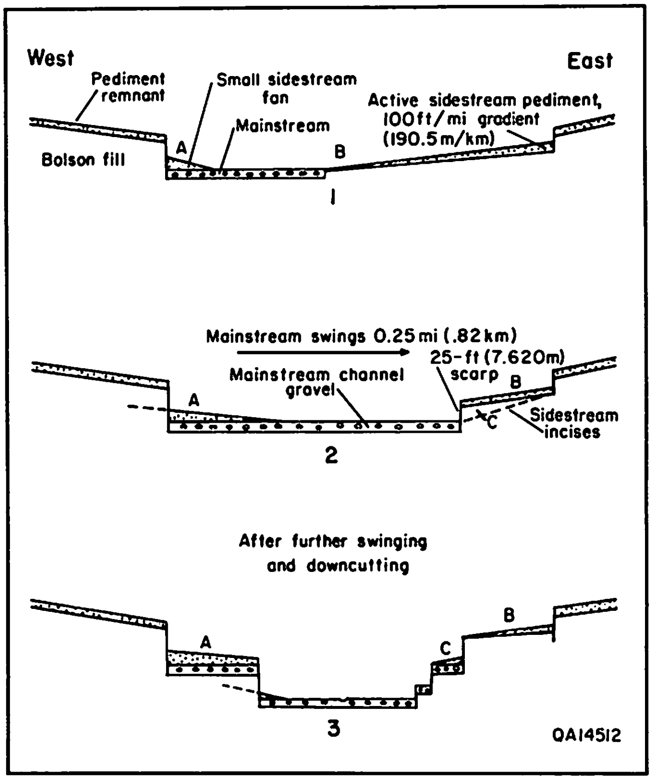



**Figure 1. Diagrams showing the evolution of the Rio Grande terraces (modified from Groat, 1972). Lateral migration of the channel-floodplain complex produced a series of "sidestream" surfaces without downcutting as shown in diagrams 1 and 2. In diagram 3, surfaces A and C are at nearly equal elevations but are not the same age.**

## STOP 4. Principal study area for disposal of low-level radioactive wastes.

**10.0** The Texas Low-Level Radioactive Waste Disposal Authority is investigating the hydrology of this location. The region is hot and arid and has an average rainfall of 10 inches/yr (25 cm/yr). Minimum and maximum average annual temperatures are 7°C (45°F) and 27°C (81°F), respectively. The 363–594-ft- (110–180-m) thick unsaturated section is composed of approximately 50 ft (15 m) of shallow alluvial sands and gravels overlying 545 ft (165 m) of clayey bolson mudstone (lacustrine) and interbedded fluvial sands and silts. (Maximum thickness of bolson sediments in the principal study area is 716 ft [217 m].) The arid climate and the thick, low-permeability, unsaturated section provide favorable hydrologic conditions for this proposed repository. Disposal is being designed to be below grade in shallow gravels to permit well-drained, oxidizing conditions with an engineered cover to inhibit migration of precipitation through the waste. More detailed discussions of the geology and unsaturated and saturated hydrology of this study area and region are in **Gustavson** (this volume, p. 27), **Mullican and Senger** (this volume, p. 37), and **Scanlon and others (1990).**

**0.0** Return to I-10. Reset odometer to zero at the interchange. Proceed eastward on I-10.

**9.7** Using the analogy of clockface times, we can see the Sierra Blanca Peaks at 10:00. Large beryllium deposits were discovered at these two intrusions. Exploration scars can be seen on Little Sierra Blanca Peak.

**17.3** Dissected bolson fill is visible on the right (north) side of I-10.

**19.6** At 2:00 are large, active alluvial fans spreading from the Quitman Mountains. These mark the southeastern corner of the Hueco Bolson. Note that alluvial/lacustrine silt deposits appear at relatively high elevations within the Hueco Bolson. They were deposited prior to the inception of the Rio Grande, which suggests significant downcutting and a different hydrologic setting in the late Cenozoic. See the paper by **Gustavson** (this volume, p. 27).

**33.7** Take the Sierra Blanca exit and proceed northward on FM 1111. The hill northwest of Sierra Blanca is composed of the Cretaceous Finlay (limestone) and Cretaceous Cox (sandstone) Formations. The latter is an important local aquifer. Farther to east, the University of Texas (St. Genevieve) vineyards east of Fort Stockton (approximately 180 mi [300 km] east on I-10) obtain their irrigation water from the Cox or, perhaps, the Maxon (also Cretaceous) sandstone.

**36.0** Surface drainage and the predevelopment ground-water drainage were east-southeast toward Lobo Flat. Water from Lobo Flat flows northward to enter the Salt Graben near Van Horn (35 mi [56 km] to the east-southeast). See **Sharp (1989, fig. 3).**

**43.0** Hills to the right of the road are capped by Cox sandstone.

**50.0** The Guadalupe Mountains can be seen on the skyline at 2:00. The regional ground-water-discharge area for the Diablo Plateau is in the Salt Basin at the foot of the Guadalupe Mountains.

**55.1** The Cornudas Mountains, upper Cenozoic laccoliths, are at 11:30.

## STOP 5 (brief) Diablo Plateau.

**65.8** The Diablo Plateau is an uplifted, east-northeast-dipping homoclinal structure. Maximum elevation of the land surface is more than 5,000 ft (1,500 m). Elevation drops to 3,600 ft (1,100 m) at the salt flats to the east. The vegetation has changed from desert plants, such as creosotebush (Larrea), in the Fort Hancock region, to grasses and yuccas on the plateau, which are indicative of the slightly cooler and wetter climate on the plateau. The plateau is considered the major recharge zone for this flow system. Discharge of the plateau ground water is by evaporation in the gypsum playas in the Salt Basin graben. A more detailed description of the hydrogeology of the Diablo Plateau is in **Kreitler and others** (this volume, p. 49). King (1965) described the area's geology.

**73.5** Precambrian rhyolite forms the Pump Station Hills. The dark-red rhyolite is composed of pink feldspar and quartz phenocrysts and is heavily fractured. The role of these rhyolitic plugs in the regional hydrogeology of the Diablo Plateau is unknown.

**77.5** Turn right on U.S. Highway 62/180. We are driving on Lower Permian rocks, mostly carbonates. A thin veneer of Cretaceous carbonate rocks exists south of U.S. 62/180.

**87.6** Turn left toward Dell City.

**97.0** Turn left to the Soil Conservation Service (SCS) dam. A set of dams with artificial recharge wells for injection of stored flood waters was constructed in the 1980's for flood control by the U.S. Soil Conservation Service.

## STOP 6. SCS Dam.

The Soil Conservation Service constructed the five flood control structures after a devastating flood in 1966 in the farming community of Dell City (Logan, 1984). Twelve wells were constructed to inject captured floodwater. The wells were sited on the basis of photoanalysis of lineaments and are capable of recharging up to 2,000 gpm (120 L/sec). The productive Permian limestones beneath the shallow veneer of alluvium are probably Bone Spring or Victorio Peak, although Ed Reed (personal communication, 1985) suggested that they are Hueco limestones. They produce about 150,000 acre-ft ($1.8 \times 10^9$ $hm^3$) annually for irrigation (Logan, 1984), and water-level declines have been approximately 1 ft (0.3 m) per year (Gates and others, 1980). There is significant return flow of irrigation water to the aquifer, as evidenced by decreased water quality and increased nitrate over time. The early paper by Scalapino (1950) remains an excellent reference on water resources of the Dell City locale.

Leave dam and proceed into Dell City. Go east on FM 2249.

## STOP 7 (brief) Gypsum pit next to feedlot.

The rock exposed in the pit is composed predominantly of gypsum. This gypsum was deposited as either a chemical precipitate or a windblown sediment. Gypsum in the salt flats to the east is now being deposited by evaporation of ground water, as will be discussed at the next stop. It is interesting to note, however, that the elevation of the top of the pit is approximately 3,640 ft (1,110 m), about 30 ft (9 m) above the elevation of the salt flats. This elevation of 3,640 ft (1,110 m) for land surface is characteristic of the land surface elevation of the northern Salt Basin and suggests an earlier depositional surface. The Salt Basin may now be filling up or may have been infilled with gypsum from an earlier hydrogeologic system.

**0.0** Reset odometer to zero at this stop. Go south on FM 1576. Northeast of this turnoff is Round Top, a Tertiary intrusion. Turn left on U.S. 62/180.

**17.4** Enter Salt Basin.

**19.7** Note the whitish color of the gypsiferous soils. Vegetation is mainly saltgrass (Distichlis), a halophyte.

## STOP 8. Salt Flat.

**22.4** There are approximately 40 gypsum playas in the northern Salt Basin. They are characterized by a shallow water table at a depth of approximately 3 ft (1 m) and a capillary fringe as shallow as 8 inches (20 cm). Tensiometer data show upward flow. The saturated and unsaturated sections contain Na-Mg-Cl-$SO_4$ brines. Sediment mineralogy is dominated by gypsum and smaller concentrations of dolomite, calcite, magnesite, and halite and sporadic pockets of native sulfur. Horizontally bedded organic layers are common. King (1948) considered the salt flats to be remnants of a Pleistocene lake that was as deep as 40 ft (12 m). Today, the gypsum playas are formed by ground-water discharge and concomitant evaporation and mineral precipitation. The ground waters are recharged on the Diablo Plateau and along the western flank of the Guadalupe and Delaware Mountains. More detailed discussions of the hydrology, hydrochemistry, and mineralogy of the salt flats are in **Boyd and Kreitler (1987)** and **Chapman and Kreitler** (this volume, p. 59). Evaporite sedimentology was reviewed by Hussain and others (1988), who stressed the transition from alluvial fans (or bajadas) to sand flats to sabkha flats from the Guadalupe Mountains on the east to the salt flats on the west. The locations of the playas are generally adjacent to permeable rocks on either side of the graben.

Continue eastward on U.S. 62/180.

**31.7** Junction with Texas Highway 54. Continue eastward to Carlsbad, New Mexico.

## Second Day: Carlsbad, New Mexico–Fort Davis, Texas
### Saturday, October 27, 1990

Carlsbad gets its drinking water from deep wells in Permian reef facies southwest of town. These are the same rocks that are exposed in the Guadalupe Mountains and in Carlsbad Caverns. An excellent discussion of the control of ground-water resources by these rocks appears in Motts (1968).

Mileage

**0.0** Reset odometer to zero at intersection of U.S. Highways 62 and 285. Drive southward on U.S. 62/180.

**8.2** Pass Dark Canyon Road to the right, which leads to Dog Canyon National Forest and the Algerita escarpment.

**18.2** White City exit. Turn right and drive through Walnut Canyon to Carlsbad Caverns National Park.

### STOP 1. Carlsbad Caverns.

We will examine these spectacular caves, which may have been formed by a combination of normal karst processes and by intensive solutioning by sulfuric acid (Egemeier, 1987) derived from the Delaware Basin to the east. Karstification of this system has apparently been active for at least the last 100,000 yr. Several handouts will be distributed, including a self-guided tour prepared by John Roth, formerly with the park. Ron Kerbo, the National Park Service scientist at the park, will be our guide.

In contrast to the relatively well-understood processes of karstification by meteoric waters charged with carbonic acid, this area undergoes the rather novel sulfuric acid dissolution process diagrammed in figure 2. Natural gas with $H_2S$ migrates because of buoyancy or overpressures from the basin into the permeable Capitan Reef rocks. Here the gas mixes with oxygenated meteoric ground water to form sulfuric acid. Evidence given in support of this hypothesis includes "excessive dissolutioning," formation of endellite (hydrated haloysite), thick gypsum deposits, and light-sulfur isotopic analyses indicative of an $H_2S$ source. It has been estimated that the natural gas from one normal-sized Permian Basin gas field is all that is needed to accomplish the observed dissolution. Another possible source of gypsum might be the bedded gypsum found in the backreef facies (see figure 3).

Return to White City and drive southwestward on U.S. 62/180. The mileage log resumes at 18.2 mi (29.1 km) out of Carlsbad. On the right for many miles we will see the Guadalupe Mountains, which consist of reef and forereef limestones. Also note the large-scale spurren, which are the regular dentate patterns developed during the growth of the reef.

### STOP 2. Castile Formation in roadcut.

**32.2** Varves in the Castile are indicative of deposition in a tranquil marine environment. The permeable evaporite minerals are undergoing intensive surface dissolution at present. Evaporite strata are highly effective confining layers in the subsurface, where they have not undergone dissolution.

The Delaware and Midland Basins. These basins are subdivisions of the Permian Basin, filled with as much as 24,000 ft (7,300 m) of Phanerozoic sediments. They experienced large-scale subsidence during the Mississippian and were deep basins through the Permian. Organic matter was preserved in the basinal facies and with time matured and produced hydrocarbons. These hydrocarbons were trapped by the basinal evaporites, such as the Castile Formation (Hills, 1984), which created the highly productive oil fields of the Permian Basin. Yates field, on the Pecos River at Iraan, Texas, is one of the largest oil fields in the United States.

The permeability of Permian facies is largely controlled by the depositional environments (fig. 3). Basinal facies tend to be very low in permeability, whereas reef and forereef facies, such as those in Carlsbad Caverns, are extremely permeable. Shelf facies, such as those in the Dell City area, are variable in permeability. Permeability in these rocks is controlled by fractures and, to a lesser extent, by dissolution features. This depositional inheritance of hydraulic

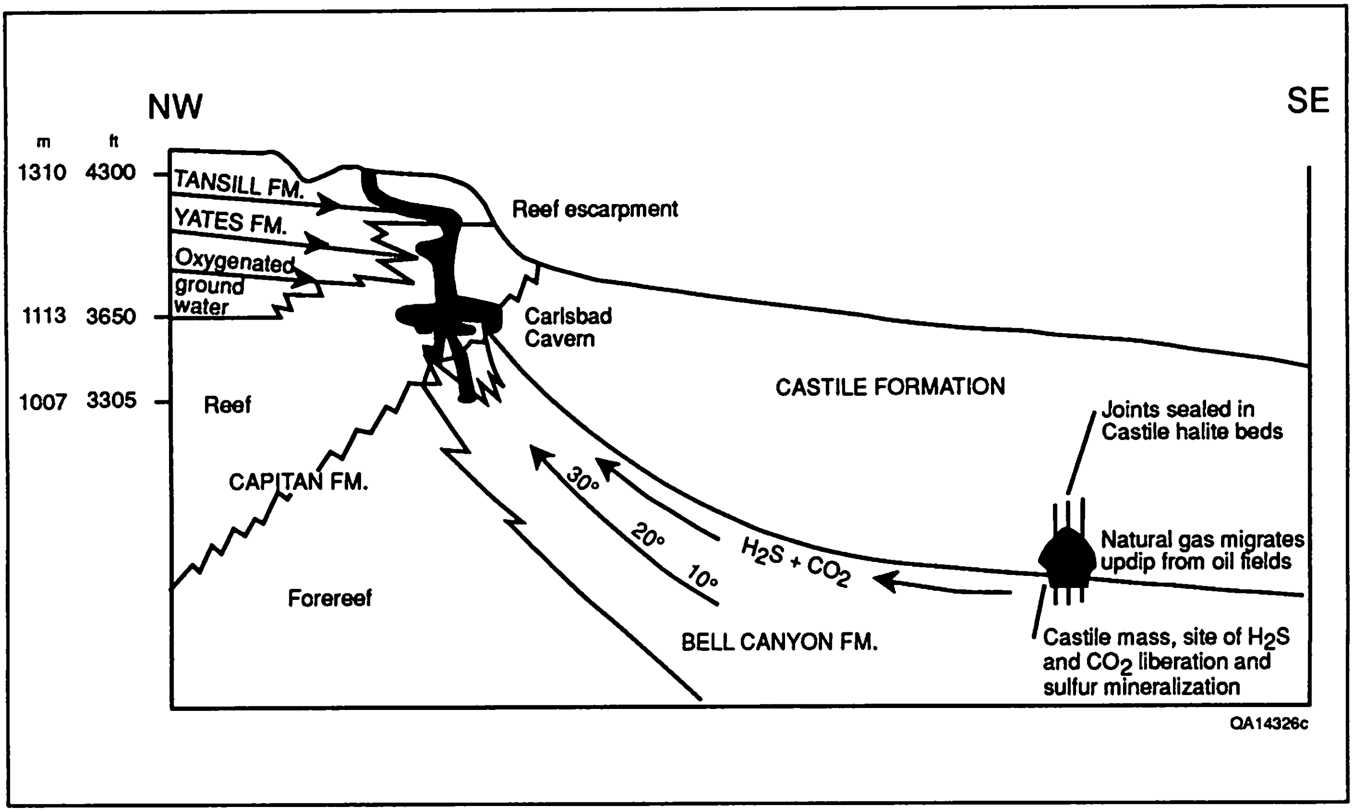



**Figure 2. Proposed model of gas ascension from the Permian Basin into the reef along the Bell Canyon Formation (see figures 3 and 4 for details of the stratigraphy). The model suggests that natural gas migrates updip and encounters anhydrite at the base of the Castile Formation. Reactions between the gas and anhydrite produced $H_2S$ and $CO_2$, as well as diagenetic limestone masses. The gases continued their updip migration into the reef facies. Mixing with oxygenated ground water forms the sulfuric acid that dissolves out the large cave passages in the Guadalupe Mountains. (Figure from Hill, 1987, reprinted by permission of the New Mexico Bureau of Mines and Mineral Resources.)**

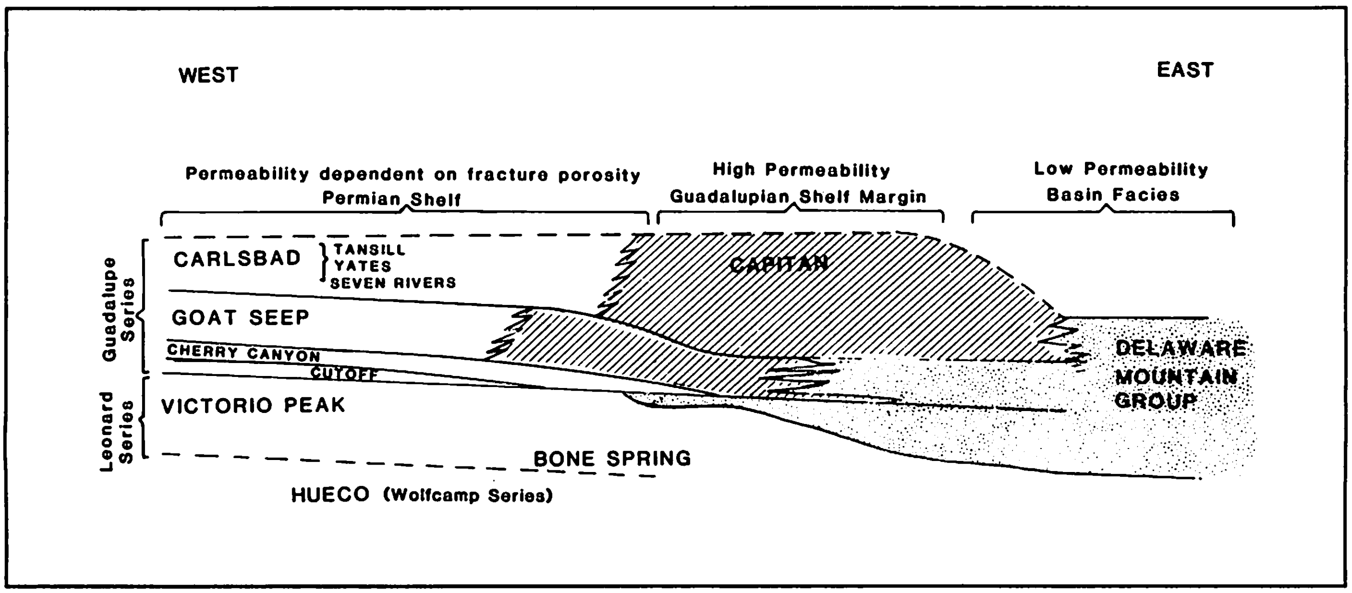



**Figure 3. Relationship of shelf, reef, and basinal Permian facies (from Sharp, 1989, reprinted by permission of the American Geophysical Union).**

properties has a profound influence on the development of flow systems in this region. The low-permeability basin facies make excellent cap rocks for petroleum reservoirs (fig. 3).

**34.4** Texas–New Mexico border. Orla Road exits immediately to the left, leading to an optional stop (Stop 2.5).

Castiles of the Permian Castile Formation (basin facies) can be seen on both sides of the road. Castiles are conical hills that may form by at least three different mechanisms. Some hills may be erosional remnants left by dissolution. Others may be breccia pipes in karst features (Anderson and Kirkland, 1980). The castiles may also be exhumed biogenic limestone cap rock resulting from leakage of deep natural gas and hydrocarbons (Kirkland and Evans, 1976). Numerous dolines attest to the solubility of these rocks. From the air a strong alignment of troughs that trend roughly east-west is also apparent. This area is described in greater detail by Olive (1957).

**44.6** Outcrop is the Lamar Limestone Member of the Bell Canyon Formation (fig. 5).

**47.4** Rader Limestone Member of the Bell Canyon. This is a paleodebris flow off the Permian reef front.

**53.4** Pine Springs store. Guadalupe Mountains National Park is to the right (north).

**54.1** Border fault zone of King (1948). This not only divides the major physiographic provinces but also marks the ground-water and surface-water divides. Flows are east to the Pecos River and west to the Salt Basin.

## STOP 3. El Capitan photo stop.

**58.2** Cherry Canyon sandstone makes up the close outcrops. Salt Flat, the Sierra Diablos, and the Diablo Plateau can be seen to the west, and the Delaware Mountains to the south form cuestas that dip gently to the east toward the center of the Delaware Basin.

**62.9** Junction of U.S. 62/180 and County Road 54. Turn left on 54 and enter the Salt Basin.

**75.1** We are traveling on dissected alluvial fans emanating from the Delaware Mountains.

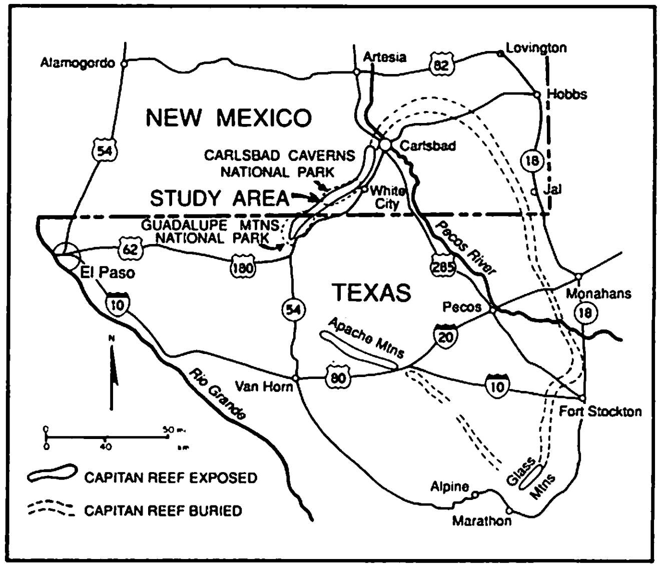



**Figure 4. Map of Permian reef facies (modified from Hill, 1987).**

**79.4** Approximate location of ground-water drainage divide between Salt Flat and the middle portion of the Salt Basin. There is some controversy over the location of this divide among authors who have compared present-day and late 1950's water-table maps. **Nielson and Sharp (1985)** noted the proximity of the Babb Flexure to this divide and suggested that the position of the resulting alluvial fans coming out of Apache Canyon to the west may have been the reason for the location of this divide.

**81.7** The Babb Flexure is visible on the right.

**85.2** Stabilized gypsum-calcite dunes on left side of the road.

## STOP 4. Discharge area for the middle ground-water basin.

**91.0** A small playa is to the east-southeast. Farther to the south-southeast are the Apache Mountains, which are composed of Guadalupian (Capitan) reef-facies limestones. The gentle south-facing slope comprises backreef facies. To the west the Permian Bone Spring Limestone caps Victorio Peak of the Sierra Diablos.

**96.5** Baylor Mountain is visible to the south. This marks the Victorio Flexure.

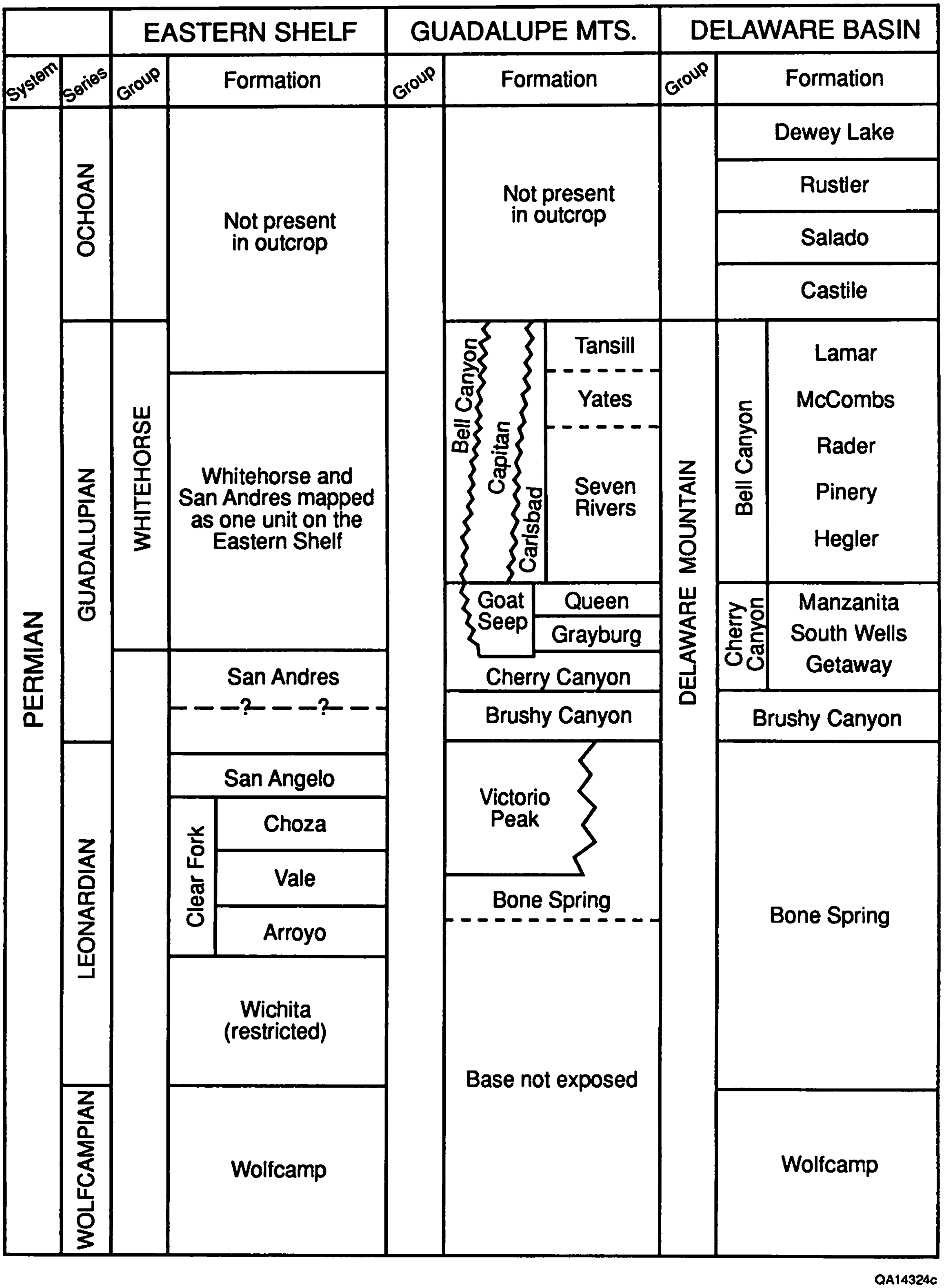



**Figure 5. Stratigraphic column of Permian units of the Eastern Shelf, the Guadalupe Mountains, and the Delaware Basin.**

**STOP 5. Victorio Flexure separates the closed-basin flow systems of the northern Salt Basin from Wildhorse Flat.**

**101.7** A structural feature apparently controls the location of the ground-water divide, either by localizing major alluvial fans or by juxtaposing rocks of varying hydraulic properties. Ranches nearby produce good-quality water from the alluvial fans. The Precambrian Hazel Sandstone (a 1-billion-yr-old, foreland-basin redbed sequence) crops out beneath the Permian Hueco Limestone in the Sierra Diablos to the west.

**107.3** Beach Mountain (ahead) is an isolated fault block.

**STOP 6** (brief) **Panorama of Wildhorse Flat, and the Apache, Wylie, and Davis Mountains.**

**112.5** Wildhorse Flat flow system. Notice the absence of playas in Wildhorse Flat and the soil colors, which are distinctly different from those elsewhere in the basin. Because the southern portion of the Salt Basin is not a closed basin (**Sharp, 1989,** and **Nielson and Sharp, 1985**), predevelopment regional flow was eastward out through Apache Mountain Permian limestones. Computer modeling of the ground-water system in Wildhorse Flat reveals the importance of the underlying Permian limestones (possibly the Bone Spring or Victorio Peak limestones) in controlling the ground-water flow. Only 40 to 70 percent of the flow is in the alluvial fill; the remaining 30 to 60 percent is probably in the limestones underlying the fill. Modeling results imply that the permeability of the fractured limestones is 3 to 5 times greater than that of the basin alluvium.

Water in the bolson is generally unconfined, but it can be locally confined by clay layers. In predevelopment conditions the water table was shallow; transmissivities range between 30,000 and 70,000 gpd/ft (0.0045 and 0.0105 $m^2$/sec). The water quality is good, ranging from 413 mg/L in Van Horn to 2,900 mg/L TDS in the northern portion of Wildhorse Flat. This bolson is the chief water supply for the towns of Van Horn and Sierra Blanca, and the Wildhorse Flat irrigation district; it has also been suggested as a supplemental source for the city of El Paso. Recharge to Wildhorse Flat is from precipitation, irrigation return flow, lateral flow from Lobo Flat to the south, and the alluvial fan deposits fringing the Baylor, Beach, and Wylie Mountains. Although an accurate water budget is unavailable, we know that pumpage for irrigation and loss by evapotranspiration exceed 40,000 acre-ft/yr ($4.9 \times 10^7$ $hm^3$) and that the water table has been dropping by about 1 ft/yr (0.3 m/yr). This depletion has reversed the ancient regional flow to the southeast and suggests the possibility of lateral flow from the more saline systems to the north of Wildhorse Flat.

Giant desiccation polygons. **Goetz (1980** and 1985) described the occurrence of these features east of Baylor Mountain in Wildhorse Flat. These features may have developed as a result of ground shaking during the 1934 Valentine (a city, not the day) earthquake. The polygons are now filling in and disappearing.

**117.0** In the distance to the right is a dam built across the arroyo for flood control. The upper Precambrian Van Horn Sandstone is the reddish formation low on the horizon. Proceed southward to Van Horn.

Two floodwater-diversion dams have been constructed, one on Three Mile Draw and the other on Sulfur Draw. Floods damaging local ranches and parts of Van Horn occur every 2 to 3 yr. An archeological survey conducted prior to construction found abundant sites of early Indians (850–1350 A.D.) in these draws. These data and some pollen data suggest wetter conditions and greater spring discharge during this period.

**128.6** The old county courthouse on the right was built of Van Horn Sandstone. The new courthouse is on the left.

**128.8** Intersection of Texas Highway 54 and U.S. 80 in Van Horn. An interesting history of Van Horn and Van Horn wells appears in Hoffer (1980).

## Optional side trip to Lobo Flat.

**0.0** Reset odometer to zero at intersection of U.S. 90 and I-10. As we drive south on U.S. 90, we leave Wildhorse Flat and enter Lobo Flat, one of the two southern extensions of the Salt Basin. This zone shows a marked steepening of the water-table gradient (flow is to the north) and corresponds to a subsurface fault zone mapped by Hay-Roe (1958) and is shown in the article by **Sharp (1989)**.

**2.4** Eagle Flat surface and subsurface drainage enters from the right. This drains Eagle Flat and the southeast portions of the Diablo Plateau.

**3.4** Enter Lobo Flat irrigation area.

Note the pecan groves that are partially replacing cotton as an irrigated crop. This bolson produced low-TDS waters (150–300 mg/L, Na-$HCO_3$ facies) from over a span of 3,200 ft (980 m) of fill. This is the best water quality in the Trans-Pecos region. The cone of depression produced by pumpage exceeded 120 ft (36 m) between 1951 and 1973 (Hood and Scalapino, 1951; Gates and others, 1980); more recent data have not been compiled. The fill consists largely of sands and gravels derived from the volcanic and crystalline rocks of the Wylie, Davis, Eagle, and Van Horn Mountains.

Proceed southward to a historical marker on the west side of U.S. 90. In the alluvium 0.5 mi (0.8 km) to the west, early settlers dug shallow "seep" wells to produce the only reliable water source between El Paso and Balmorhea (pronounced "Bal′ - mo - ray").

Goetz (1977) evaluated fault activity in the Salt Basin and noted recent motion along the bounding fault on the west side of Lobo Flat, but she was unable to provide quantitative estimates of the rate of fault movement. Causes of this movement are unknown, but it seems doubtful that it is the result of compaction of the bolson fill. Therefore, tectonic activity is the likely cause. There is documented displacement of faults in Trans-Pecos Texas during and immediately after the 1934 Valentine earthquake.

Return to Van Horn. U.S. 90 continues on through Ryan Flat, which is also a tributary of the Salt Basin. The southern fringe of Ryan Flat marks the drainage divide between the Salt Basin and the Rio Grande.

**0.0** Reset odometer to zero at intersection of I-10 and U.S. 90. Proceed eastward on I-10. The Wylie Mountains are on the right.

**4.3** Cross Wildhorse Creek. The creek flows northward after storms.

**6.9** On the left is a talc-processing plant. The talc is mined from the Precambrian Allamoore Formation between Sierra Blanca and Van Horn. Another processing plant is at the mine site. There is a good market for this talc because it is largely asbestos free.

The Apache Mountains are the low hills to the northeast. They are composed of highly permeable Permian reef and backreef limestones, which effectively "drain" Wildhorse Flat (see the papers by **Hiss (1980)**, **Nielson and Sharp (1985)**, and Wood (1968). There is significant local water production from the Cretaceous Cox Sandstone that underlies I-10 at this location. The sandstones are covered by a veneer of alluvium that is considerably thinner than the fill just to the north in Wildhorse Flat, where the alluvium is as much as 1,000 ft (300 m) thick and overlies limestone, probably Permian. Clearly there is a major covered fault north of the Wylie Mountains.

**13.2** Michigan Flat road. To the right is Michigan Flat, the other southern extension of the Salt Basin. Two ranches irrigate extensively from the bolson fill, which is mainly derived from the volcanic and other rocks of the Davis and Wylie Mountains. Water quality is generally good in Michigan Flat. The more productive alluvium/carbonate system in Wildhorse Flat to the northwest is considerably more brackish.

**17.9** Mile marker 157. The low hills to the north are the Permian Munn Formation (Guadalupian) overlain by the Cox Sandstone. In the low hills to the south is the Seven Rivers Formation, also capped by the Cox Sandstone.

**19.8** Plateau ("plah - too′").

**26.4** Boracho ("bore - ah′ - cho") Station exit. This is the approximate location of the surface-water divide between the Salt Basin and the Pecos River flow systems. The Davis

Mountains, composed of Tertiary volcanics, are to the south.

**29.4** The Cretaceous Boracho Limestone, capped by the Cretaceous Buda* ("bew′ - dah") Limestone, is in the roadcuts to the right. The Boracho is characterized by nodular weathering. There are several small road-metal quarries in the Boracho along the highway.

**40.8** Cross Cherry Creek. The roadcuts for the next several miles are in the Boracho Formation.

**45.2** We enter the Rounsaville Syncline. The Spring Hills are to the left. No significant springs are present here today (Brune, 1981).

**LaFave and Sharp (1987)** and LaFave (1987) evaluated the trend of faults and lineations in the region between the Salt and Toyah Basins. There is a strong east-west orientation, which is similar to the strike of the Rounsaville Syncline. There may be an equally strong anisotropy in the carbonate rocks in this area. The direction of maximum permeability should have a similar strike. These trends and the location of highly permeable reef facies are conducive to interbasin flow.

**46.4** Rest area. Gomez Peak is on the right. The area between the road and Gomez Peak is connected hydraulically to the springs at Balmorhea. A quickening of spring flow soon follows after thunderstorms near Gomez Peak.

**48.2** Intersection of I-10 and I-20. Stay right on I-10.

**51.6** More roadcuts through the Boracho Limestone.

**53.8** Take U.S. 290 to the right.

**STOP 7. Take road to the right to the Kingston Ranch and Phantom Lake Spring.**

**66.6** **LaFave and Sharp (1987)** showed a photo (this volume, p. 92) of the orifice of Phantom Lake Spring. The spring issues from an A-shaped cave in Cretaceous limestone. Faulting north of the Davis Mountains has juxtaposed the highly permeable Permian and Cretaceous limestones. Phantom Lake Spring maintains a remarkably steady flow of slightly thermal brackish water, the hallmarks of discharge from a regional flow system as documented by Maxey and Mifflin (1966). Flow quickens in response to local rainfall events, so that both locally and regionally derived flow must be present. Although the gauging record is incomplete on these springs, the discharge is showing a gradual long-term decline.

Return to U.S. 290 and turn right. We cross the dry creekbed, which used to convey flow from Phantom Lake Spring toward the Toyah Basin. Spring flow is now confined to irrigation ditches.

**0.0** Reset odometer to zero at intersection of U.S. 290 and Texas Highway 17. Go south on Texas 17.

**16.6** Aguja Canyon Road. Continue on Texas 17 to Fort Davis. On the right is a shallow alluvial fill. Straight ahead is the eastern extension of the Rounsaville Syncline.

**28.8** Crest of Wild Rose Pass.

**40.7** Fort Davis and the junction of Texas 17 with Texas 118, which goes to the University of Texas' McDonald Observatory.

After dinner, we will drive up to the McDonald Observatory for a view (weather permitting) of the stars.

* Buda is the anglicized Spanish word, "viuda," which means "widow" (Dow Chapman, personal communication, 1990). The first use of the term Buda was for a small Central Texas community between Austin and San Marcos and was so named for a widow who ran the stagecoach station there in the 1800's (Billy Porterfield, personal communication, 1990). The type locality for the Buda Formation, however, is along Shoal Creek in Austin, Texas.

## Third Day: Fort Davis–Balmorhea State Park–Monahans State Park

**Sunday, October 28, 1990**

**Mileage**

**0.0** Reset odometer to zero at the junction of Texas Highway 17 and U.S. 290. Leave Fort Davis on Texas 17 and return to Toyahvale. The road back to Balmorhea winds through the Davis Mountains.

The Davis Mountains cover an irregularly shaped area of approximately 1,400 mi$^2$ (3,626 km$^2$) in Jeff Davis, Brewster, and Presidio Counties. These are a highly dissected remnant of a vast volcanic field that may have covered thousands of square miles of Trans-Pecos Texas and northern Mexico. The thick sequence of Tertiary volcanic rocks is now bounded on the east by the Marathon foldbelt and on the north by the Apache Mountains. The rocks are largely extrusives (tuffs), although both sedimentary units (conglomerates and a few thin sandstones) and intrusive rocks are present. The intrusives form dikes and plugs, many of which form peaks today. The intrusive rocks (basalts, andesites, and trachytes) are presumed to be low in permeability. The volcanic rocks are considered poor aquifer materials, although the town of Alpine, south of the Davis Mountains, obtains its water from fractured basalts. Beneath the volcanic rocks lie sedimentary rocks of Cretaceous and probably Permian age. Igneous activity occurred mostly in the early Tertiary (Eocene–Oligocene) (Price and others, 1986; Parker, 1988).

The geomorphology of the Davis Mountains in this area has not been studied, although the classic work by Albritton and Bryan (1939) identified three cycles of erosion and deposition in the southern Davis Mountains. Albritton and Bryan divided the valley flats into three formations, the Neville, the Calamity, and the Kokernut. Today streams are incising these formations. Perennial streams, although not numerous, do occur in the Davis Mountains. Ground water is produced from alluvium, volcanic rocks, and locally from underlying Cretaceous (?) limestones. Water quality in shallow wells is generally good (Na-$HCO_3$ facies of less than 300 mg/L TDS); deep wells in the limestones produce brackish water that is chemically similar to that flowing from the springs at Balmorhea and from wells in the Toyah Basin.

The presence of windmills and perennial streams in the Davis Mountains indicates the existence of either a perched aquifer or the upper portion of a more regional system. Water-quality data imply the former. Comparison of the shallow water table in the Davis Mountains with the deep water tables in the Diablo Plateau and in the Guadalupe Mountains demonstrates that depth to ground water results from different hydrogeologic settings that may be related to both the hydraulic properties of the rocks and to the differences in rainfall. Annual precipitation rates vary from approximately 12 inches (30 cm) on the Diablo Plateau to 14 inches (36 cm) in the Guadalupe Mountains and as much as 18 inches (46 cm) in the Davis Mountains. Rock types are also different, the Diablo Plateau being composed of fractured, locally permeable back-reef carbonates and the Guadalupe Mountains of highly permeable reef carbonates, in contrast to the volcanics of the Davis Mountains. In outcrop the volcanics show extensive vertical fractures that should result in a high vertical permeability. Nonwelded ash-flow tuffs, interflow units, or occlusion of porosity by weathering must limit vertical percolation rates.

### STOP 1. Entrance to Balmorhea State Park. View of San Solomon Spring and the pupfish canals. Refer to the paper by LaFave and Sharp (1987).

**1.3** San Solomon Spring. This is the largest of a group of springs in the Balmorhea area (White and others, 1941). Phantom Lake, San Solomon, and Giffin Springs (fig. 6) have been classified as artesian; they are mildly thermal (20-23° C) and brackish (about 2,000–2,300 mg/L TDS). The three other springs (Saragosa, East Sandia, and West Sandia) are considered "water-table" springs. San Solomon Spring issues from gravels about 15 ft (5 m) beneath the level of the swimming pool. This is considered one of the five best "swimming holes" in Texas (Dooley, 1988). Its discharge flows in canals to Balmorhea Lake (several miles to the

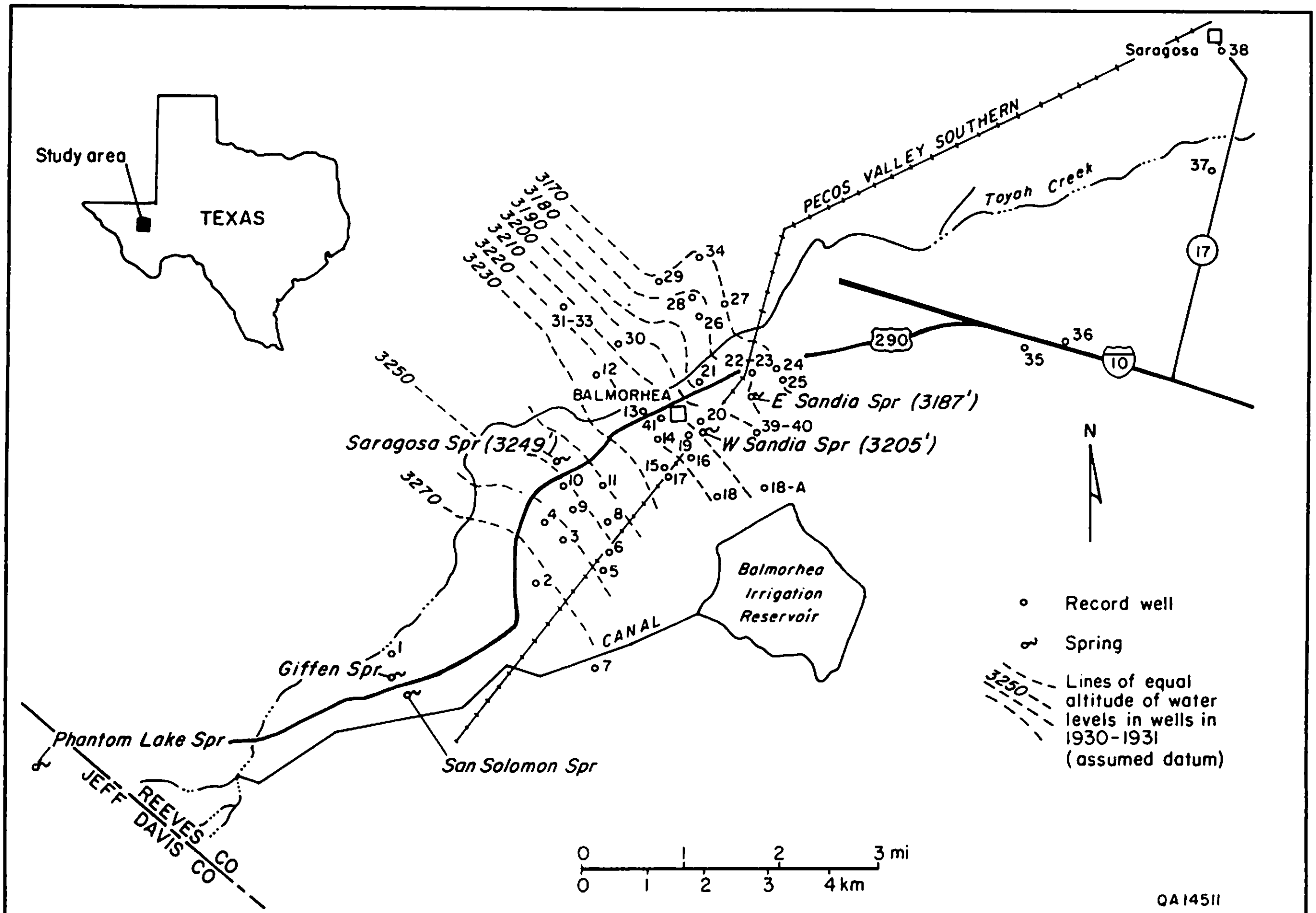



**Figure 6. Location of wells and springs and general water table configuration in the vicinity of Balmorhea, 1930–1931 (from White and others, 1941).**

northeast), which stores water for irrigation and is used for recreation. Saragosa Spring previously issued from the banks of Toyah Creek in the town of Balmorhea. We have not been able to locate it; it has probably ceased to flow because of a regional lowering of the water table. Water chemistry of East and West Sandia Springs is virtually identical to that of the artesian springs, but East and West Sandia Springs are cooler (13–15° C) and flow from the alluvium (about 40 ft [12 m] thick) of Toyah Creek.

The combined discharge of the Balmorhea springs drops to about 23,000 gpm (1,450 L/sec) in dry years but is considerably greater in wet years. There appear to be two sources of recharge for the springs—a regional carbonate system that supplies the brackish water during low-flow conditions and periodic influx from Davis Mountains runoff, which is fresher and cooler. Discharge from the springs has been slowly declining since the advent of irrigation and, perhaps, other ground-water developments in the Davis Mountains, Van Horn, and Balmorhea.

The pupfish canals. The entire population of the Comanche Springs pupfish is located in these small canals. This species originally lived in Comanche Springs in Fort Stockton (Pecos County) and has been successfully transplanted to these canals. Comanche Springs ceased to flow in the 1950's because of irrigation west of Fort Stockton. Consequently, the natural habitat of pupfish has vanished. Another set of springs in Pecos County, Leon Springs, maintained a separate species that is now extinct because of the cessation of natural spring flow. The pupfish genus *Cyprinodon* inhabited more interconnected drainage systems that existed in the southwest during Quaternary periods (10,000 to 30,000 yr B.P.). The population presumably originally existed as one species. Because of increasing aridity since then, however, most of the water bodies dried up; perennial freshwater habitats became restricted to the orifices of large springs. The pupfish adapted to the small and specialized habitats available to them: spring-fed shallow streams and marshes, subthermal springs, and manmade thermal springs from artesian wells. Each of these habitats exhibits characteristic population controls and growth patterns (Brown, 1971; Brown and Feldmeth, 1971).

Spring-fed marshes and streams exhibit the greatest variations in flow, salinity, temperature, and, consequently, in pupfish population. Because of seasonal precipitation, usually greatest in the spring and least in the summer, there can be large die-offs during the summer that permit only a few fish to reproduce for the next season. Subthermal springs that represent discharge from regional aquifer systems are more stable hydrologically; they typically have relatively constant temperature and discharge. Pupfish populations here are small (as few as 20 to 40 fish) but stable. Mexican Spring in Death Valley, California, maintains a stable population with only four or five breeding males (Miller, 1950). Pupfish have also migrated into, and live in, the discharge of artesian thermal wells where they thrive on abundant blue-green algae. The fish can survive in temperatures of as much as 108° F, but they perish quickly at temperatures of over 110° F. These populations sense abrupt temperature gradients and stay in the safe, nutrient-rich waters.

**3.0** In the distance on the right are cabins and trailers at Balmorhea Lake. The lake is fed by San Solomon Spring. The impoundment retains water for irrigation. Approximately 6,000 acres (2,500 hectares) of farm land is irrigated by this spring.

**4.3** Town of Balmorhea. The town is named after its three developers, Balcom, Morrow, and Rhea. Proceed eastward on U.S. 290 through Brogado (Bro - gah′ - do). Brogado is the site of an old Spanish fort, or presidio. East and West Sandia (San - dee′ - yah) Springs originate just southwest of the volcanic hill south of U.S. 290.

**11.1** Junction of Texas 17 and I-10. Exit and drive north on Texas 17 toward Pecos.

**13.6** Saragosa. This little town has been rebuilt twice: first, when Texas Highway 17 was moved to its present location and again in 1986, when the town was virtually destroyed by a tornado. Irrigation from Saragosa north to the Pecos River is strictly by ground water pumped from thick Cenozoic alluvial fill and to a lesser extent, underlying Cretaceous carbonates. These limestones represent the westernmost limit of the Edwards Plateau.

FM 1215 exits to the east at Saragosa. For an optional stop (Stop 1.5) drive approximately 2 mi (3.2 km) to the east, where the State of Texas Land Commission has an experimental farm. Here various climate- and salt-tolerant crops are being evaluated for their use in Trans-Pecos Texas. These crops include the Afghan pine and khanaf (which is used to produce high-quality paper).

**13.9** Cross Toyah Creek. Toyah Creek used to carry springflow through the Toyah Basin. During periods of intense rainfall, Toyah Creek occasionally flowed into the Pecos River.

## STOP 2. Verhalen.

**22.3** A short side trip to the west leads to the heart of the Toyah Basin irrigation district. The main crops being grown are cotton, alfalfa, and feed grains. The famous Pecos cantaloupes, pecans, tomatoes, and onions are also grown. A thesis by LaFave (1987) and the more regional study by Ashworth (1990) highlight the hydrogeology of ground water in the Toyah Basin.

The evolution of the flow system in response to ground-water development is depicted in the paper by **Sharp (1989).** Toyah Basin developed during the Cenozoic in response to dissolution of the underlying Rustler and Castile Formations. More than 1,400 ft (425 m) of alluvial fill has been deposited. The isopach map is shown as figure 7.

The basin fill consists of sands, gravels, silts, and caliche, but the overall stratigraphy of the fill is not well known. It is, however, a highly productive brackish-water aquifer. LaFave (1987) identified

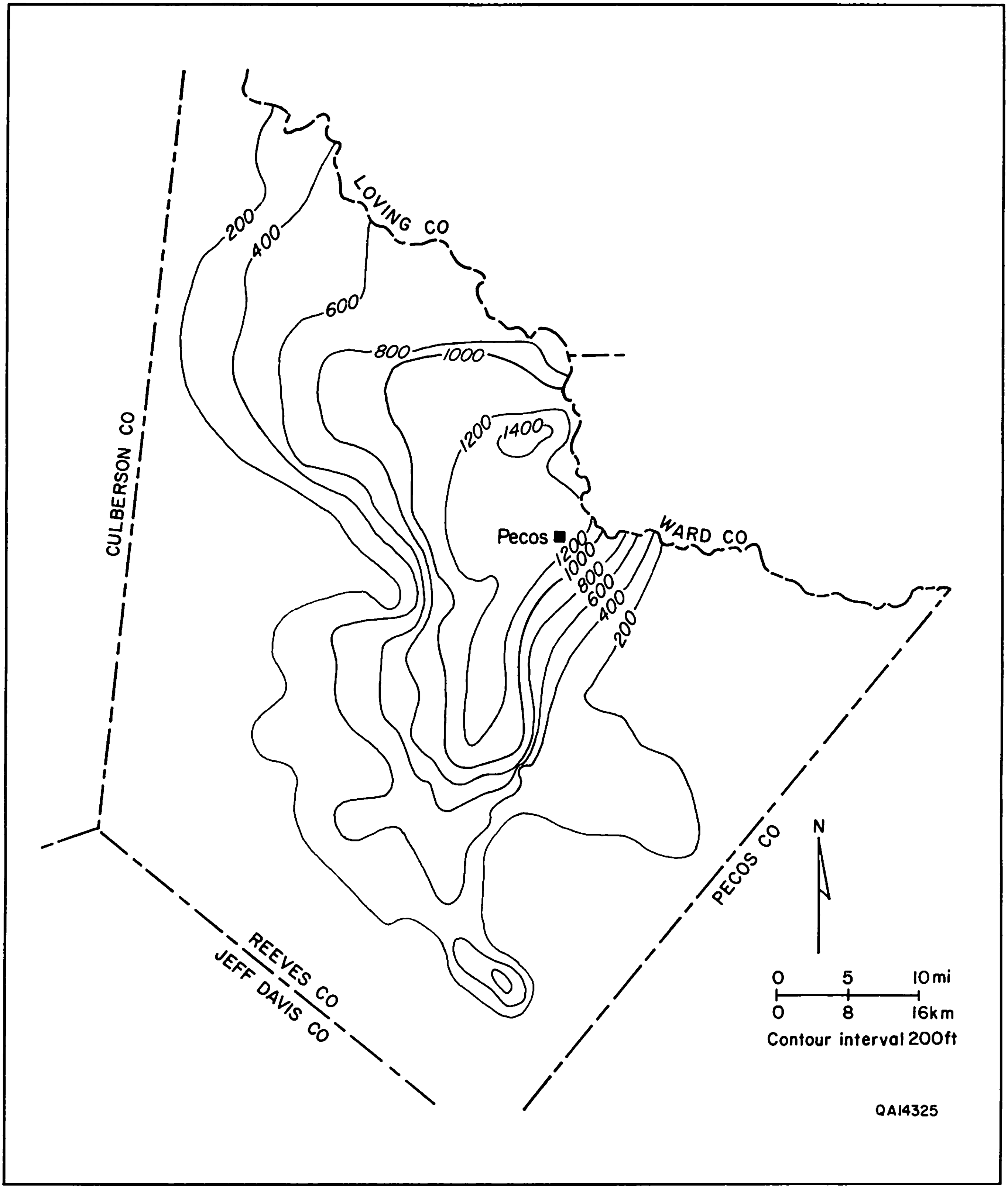



**Figure 7. Isopach map of Cenozoic fill in the Toyah Basin, formed by dissolution of underlying Permian evaporite units.**

two hydrochemical facies: Cl-dominant and $SO_4$-dominant. $SO_4$-dominant waters are found on the western side of the basin's north-south axis; Cl-dominated waters are on the eastern side. He hypothesized that the sulfate was derived either from regional ground-water flow through Permian carbonates and evaporites or from infiltration from the surface draws that funnel surface water eastward into the Toyah Basin. In both cases, however, water chemistry is very similar to that found in the Apache Mountains to the west, supporting the concept of regional west-to-east flow in the northern Trans-Pecos. Although LaFave did not observe changes in water quality from irrigation return flows, recent geostatistical analyses (Lou Zachos, personal communication, 1989), and the presence of nitrates in the center of the Toyah Basin irrigation area (Ashworth, 1990) indicate some effect, but there has been no significant deterioration of ground-water quality in the Toyah Basin during its development as an irrigation area.

Several methods of irrigation are used in Trans-Pecos Texas. Operating costs, initial capital costs, water availability and salinity, crop requirements, soil permeability, and field configuration determine the optimal type of irrigation (Bernstein and Francois, 1973). The most common practice is furrow irrigation, which entails using irrigation ditches, some of which are lined with concrete. The furrow method works best with row crops, such as cotton. Sprinkler irrigation is used for grain and forage crops such as alfalfa. The most efficient method is drip or trickle irrigation, in which water drips out of perforated metal or plastic pipe. This method is commonly used in orchards. Pumpage for irrigation has been declining since the 1950's because of the increased costs of natural gas and electricity, legal actions against the Anderson-Clayton cotton cartel, and fluctuations in the price of cotton, the major crop.

Return to Verhalen and continue north to Pecos.

**40.2** Pecos city limits. Turn right on I-20.

**42.7** Junction of I-20 and U.S. 285. Turn right for optional stop (Stop 2.5). Approximately 4 mi east on U.S. 285, stop at Toyah Lake.

Toyah Lake represents a type of playa that is different from those in the Salt Basin. Before ground-water development in the Toyah Basin, Toyah Lake was fed by a number of springs. With the advent of irrigation, water levels dropped beneath the elevation of the lakebed. It now receives most of its seasonal waters from direct precipitation and surface runoff, although the presence of saltcedar near its shoreline indicates that the water table may have risen significantly in recent years because of the decline in pumpage of water for irrigation.

Farther east, U.S. 285 crosses the Santa Rosa aquifer of Triassic age. The Santa Rosa yields good-quality water, but its yield is much more limited than that of the Toyah Basin aquifer. Flow is generally from northeast to southwest under the Pecos River (Ashworth, 1990).

**0.0** Return to I-20. Reset odometer to zero. Drive east on I-20. Mosquito Lake, a small playa lake, can be seen southeast of the intersection. Its origin is similar to that of Toyah Lake, unlike that of playas in the Salt Basin.

**3.8** Barstow, Texas, on the left, is locally famous for its peach orchards. This town once depended on Pecos River water for irrigation of its fields, but the Pecos is now too salty for irrigation use except during a few times of great discharge.

## STOP 3. Cross the Pecos River.

**5.0** Evolution of the Pecos River has been controlled by a sequence of local and regional tectonic events (Thomas, 1972; Bachman, 1980). Near its confluence with the Rio Grande near Langtry, Texas, the Pecos cuts through deep, spectacular canyons. The "pre-Ogallala" Pecos River flowed from near the San Juan Mountains and joined the Concho River, a predecessor to the Rio Grande. The lower Pecos started entrenching its present valley at that time. At the time of Ogallala deposition headwaters of the Pecos were probably diverted from the San Juans to the Tularosa Basin. The river then began exhuming a paleovalley formed by dissolution of Permian evaporites and filled by

Ogallala sediments. By the late Pleistocene the Pecos had eroded down to the resistant Permian San Andres Formation and had begun eastward lateral and headward migration to cut the small western Caprock Escarpment of the High Plains. Headward erosion captured the drainage of the Brazos and Canadian Rivers that flowed east-southeastward over the Llano Estacado, although the Canadian River later recaptured its headwaters when it also breached the Caprock.

The phreatophytes along the banks of the Pecos are mostly tamarisk, or saltcedar (Robinson, 1965), a plant that was introduced to Texas about 150 yr ago. It is a major water consumer, transpiring between 5 and 9 acre-ft (12 and 23 $hm^3$) of water per acre per year in stands along the Pecos River. In the late 1960's the U.S. Bureau of Reclamation removed 54,000 acres (22,000 $hm^3$) of saltcedar along the Pecos between Lake Sumner, New Mexico, and Pecos, Texas. It was hoped that this would conserve 47,000 to 70,000 acre-ft (19,000 to 28,000 $hm^3$) of water per year. The success of this clearing is debatable: saltcedars send roots down as much as 33 ft (10 m) and the roots send up new shoots rapidly. Suggestions to use potent pesticides have not been well received. In addition, the saltcedars provide excellent wildlife habitat, and further clearing has been opposed for these reasons.

**7.3** Some fields visible on the left are irrigated with waters from the Allurosa aquifer, the hydrostratigraphically undifferentiated alluvium, and Santa Rosa aquifer systems.

**26.0** Exit left on Texas Highway 115 to Wink for an optional stop. The famed Wink Sink (Stop 3.5) is several miles north of Wink. This collapse structure was apparently the result of evaporite dissolution along improperly cased oil wells (Baumgardner and others, 1982; Johnson, 1989).

**39.4** We are entering an area of extensive sand dunes. The partially vegetated sand dunes extend for several miles.

**42.8** Exit to Monahans Sand Hills State Park. Dunes within the park can be seen to the left. A detailed description of dune geology is in Machenberg (1984).

**62.4** Odessa Meteor Crater Road.

**85.0** Exit I-20 to Midland-Odessa Airport.

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# Water Resources of the El Paso Area, Texas

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## Abstract

Fresh to slightly saline ground water in the El Paso area occurs in deposits of the Hueco and Mesilla Bolsons and the Rio Grande alluvium. Pumpage, primarily for public supply use, is in excess of recharge and has resulted in significant water-level declines and quality deterioration. Surface water is restricted to the Rio Grande and is currently 90 percent allocated for irrigation. Increasing water demands caused by a rapidly growing population will result in a depleted fresh-water source in the early to middle 21st century unless alternate sources can be acquired.

## Introduction

Water has historically played a key role in the development of the El Paso region. The first irrigation practiced in the area was near the Spanish mission Our Lady of Guadalupe of El Paso, which was founded in 1659 in what is now downtown Ciudad Juarez, Mexico (U.S. Bureau of Reclamation, 1973). Today, water needed to support a rapidly expanding population and economy on both the United States and Mexico sides of the border is primarily supplied from a combination of ground water that occurs in bolson and alluvial deposits and surface water from the Rio Grande. Increasing water requirements, declining water levels, and deteriorating water quality make water-resource management a necessity for this arid region.

## Structure

The geologic framework of the El Paso area, which lies within the Basin and Range province, is primarily controlled by the Rio Grande Rift, which is characterized by a series of grabens, or downdropped basins. The Hueco and Mesilla Basins in El Paso County were formed by normal block faulting and are asymmetrical, downward displacement being greater on one side of the basin than the other. Upland areas flanking the basins on the west and east in Texas are outcrops of rocks ranging in age from Precambrian to Tertiary, whereas west- and southwest-flanking uplands in Chihuahua, Mexico, are outcrops of Cretaceous rocks.

## Hueco Bolson

East of the Franklin Mountains, the Hueco Bolson extends northward into New Mexico, where it merges with the Tularosa Bolson, and southeastward to about Fort Quitman, where it lies between several mountain ranges in Texas and Mexico. South of the City of El Paso, the El Paso–Juarez Valley of the Rio Grande occupies the southwestern edge of the bolson (fig. 1).

The Hueco Bolson, the principal aquifer in the El Paso area, consists of an upper fluvial zone (Camp Rice Formation) of mostly stream-channel and floodplain deposits composed of silt, sand, gravel, and caliche, and a lower lacustrine zone (Fort Hancock Formation) containing mostly clay and silt (Strain, 1966). Maximum thickness of the bolson, according to Mattick (1967) and Gates and others (1978), is about 9,000 ft (2,700 m) and occurs within a deep structural trough paralleling the east side of the Franklin Mountains. Thickness and sediment grain size generally decrease in an easterly direction across the basin.

Recharge to the Hueco Bolson is about 39,300 acre-ft (48 $hm^3$) per year, 6,000 acre-ft (7 $hm^3$) per year principally from runoff along the base of the surrounding mountains, and about 33,300 acre-ft (41 $hm^3$) per year leakage from the overlying Rio Grande alluvium (Meyer, 1976). In 1985, the City of El Paso put into operation the Fred Hervey Water Reclamation Plant. This facility is currently restoring 5 to 6 million gallons (19 to 23 million liters) of sewage a day to drinking water standards and injecting most of it back into the aquifer through 10 recharge wells.

Ground water in the bolson outside the valley is under water-table conditions, whereas in the valley the aquifer is under leaky artesian conditions. Depth to water in the aquifer ranges from about 350 ft (107 m) near pumping centers to less than 100 ft (30 m) elsewhere.

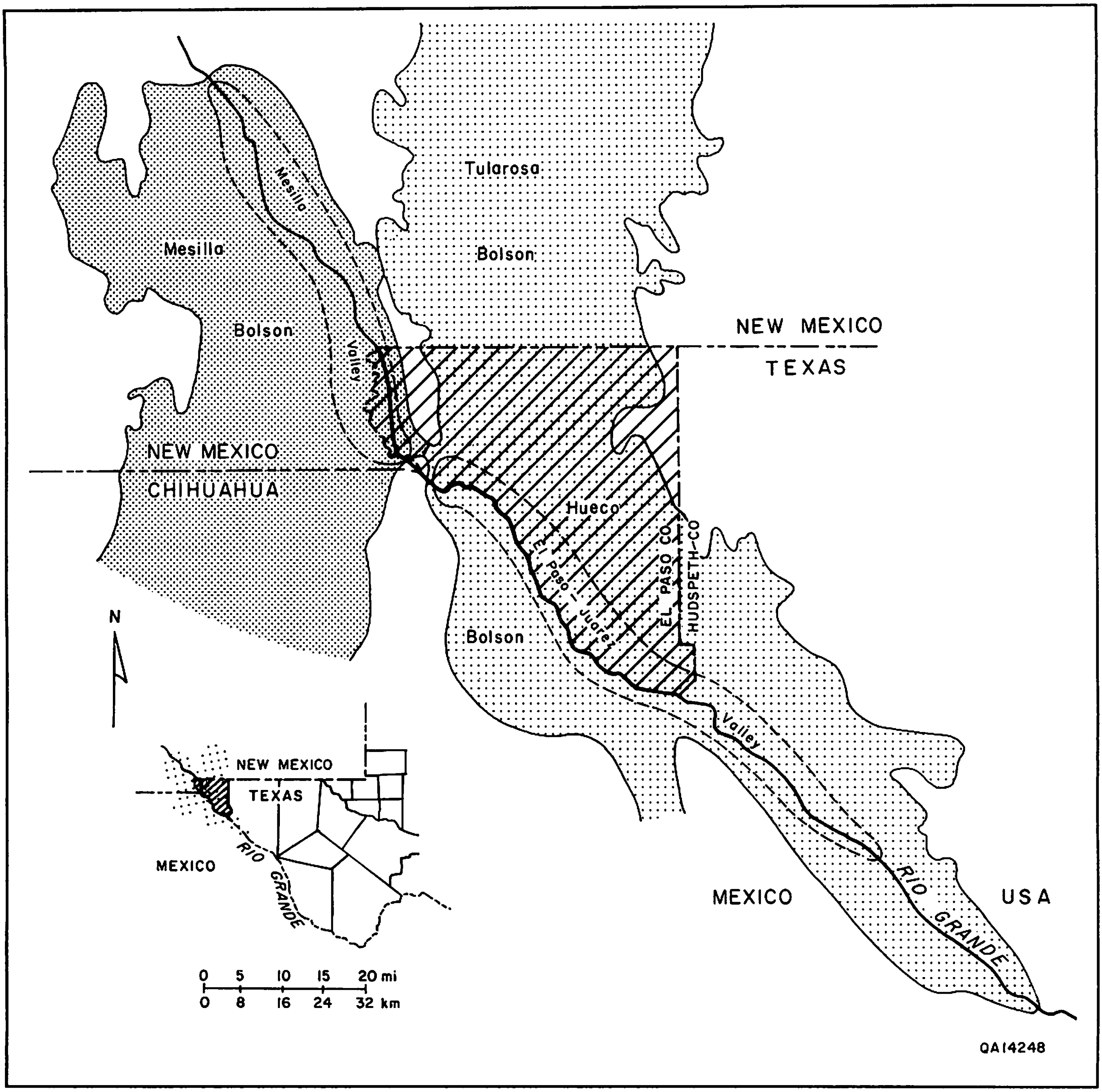


**Figure 1. Physiographic map of the El Paso area.**

Large-scale ground-water withdrawals from the Hueco Bolson, especially from municipal well fields located in the downtown areas of El Paso and Ciudad Juarez, have caused major water-level declines, which have significantly changed the direction and rate of flow and chemical quality of ground water in the aquifer. Declining water levels, primarily in the shallow water table, have also resulted in a minor amount of land-surface subsidence due to the dewatering of clay beds (Land and Armstrong, 1985). Historically (1903–1989), the greatest declines—as much as 150 ft (46 m)—occurred in the downtown areas of El Paso and Ciudad Juarez (fig. 2).

The chemical quality of Hueco Bolson ground water varies, both areally and with depth. Dissolved-solids concentrations range from less than 500 to more than 1,500 mg/L. Increasing dissolved-solids

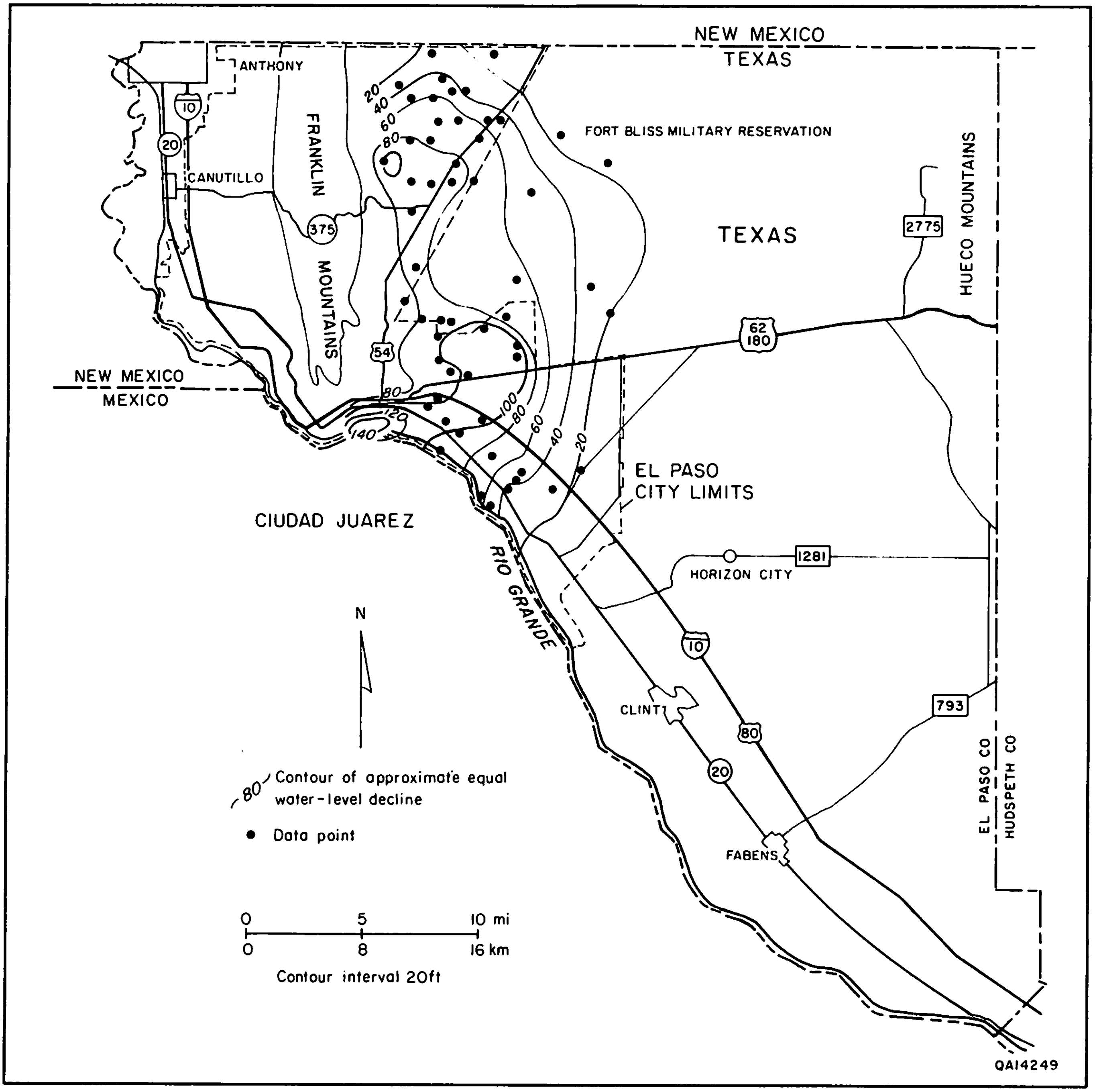


**Figure 2. Water-level decline in the Hueco Bolson aquifer from 1903 to 1989.**

concentrations in fresh-water zones are attributed mainly to downward leakage of brackish water from shallow zones and possibly upconing of brackish water from below. Analyses of water samples show an average annual increase in dissolved solids of about 10 mg/L since the 1950's and 1960's in the United States and about 30 mg/L since the 1970's in Ciudad Juarez (White, 1983).

# Mesilla Bolson

The Mesilla Bolson occupies a basin that extends from Las Cruces, New Mexico, southward into Mexico (fig. 1). On the surface, this area is characterized by a broad, nearly level plain called La Mesa and an incised river valley of the Rio Grande known as the Mesilla Valley. In the El Paso area, the lower Mesilla Valley extends southward from the New Mexico–Texas state line at Anthony to the "El Paso del Norte," the narrow gap between the southern end of the Franklin Mountains and the northern end of the Sierra de Juarez in Mexico.

The bolson consists of alluvial basin-fill deposits composed of clay, silt, sand, and gravel of the Santa Fe Group and is overlain by the Rio Grande alluvium in the vicinity of the river (Hawley and others, 1969). In the lower Mesilla Valley near the city of Canutillo, ground water is pumped for municipal, industrial, and irrigation supply from three water-bearing zones differentiated by their lithology, water levels, and chemical quality. These zones are locally referred to as the shallow, intermediate, and deep aquifers. The shallow aquifer includes the overlying Rio Grande alluvium.

Leggat and others (1962) estimated that about 18,000 acre-ft (22 $hm^3$) of water per year recharges the aquifers in the lower Mesilla Valley. Recharge occurs by precipitation in the valley and runoff from the Franklin Mountains; by seepage from canals, laterals, the Rio Grande, and applied irrigation water; and by ground-water underflow from uplands in New Mexico (Alvarez and Buckner, 1980).

Depth to water in the shallow aquifer is generally 5 to 15 ft (1.5 to 4.6 m) below land surface, although during periods of heavy withdrawals for irrigation, water levels may decline an additional 10 to 15 ft (3 to 4.6 m). Water levels in wells completed in the deeper aquifers range from slightly higher to several feet lower than those in the shallow aquifer, especially in the more heavily pumped area between Anthony and Canutillo. Declines of 10 to 30 ft (3 to 9 m) in the intermediate and deep aquifers have reversed the vertical flow of ground water, causing leakage of inferior-quality water from the shallow zone.

The chemical quality of the ground water in the lower Mesilla Valley ranges from very fresh to saline, salinity generally increasing to the south along the valley. The water is also freshest in the deep aquifer and contains progressively higher concentrations of dissolved solids in the shallower zones. Dissolved-solids concentration has increased in ground water produced from the intermediate aquifer at an average rate of about 9 mg/L per year (White, 1983).

# Rio Grande Alluvium

The Rio Grande alluvium consists of stream-channel and floodplain deposits composed of poorly sorted clay, silt, sand, and gravel, which are in part derived from the erosion and redeposition of adjacent bolson deposits. These alluvial sediments, which reach an average maximum thickness of about 200 ft (60 m), are hydraulically connected to underlying bolson deposits and form a part of the shallow aquifer. The alluvium is an important source of shallow ground water for supplemental irrigation when the surface-water flow in the Rio Grande is insufficient to meet the total agricultural water needs of the valley farmers. Recharge occurs from infiltration of precipitation that falls directly on the surface and runoff from the adjoining bolson surfaces; leakage from the Rio Grande and numerous canals that traverse the heavily cultivated and irrigated floodplain; and excess irrigation water applied to the cultivated land (Alvarez and Buckner, 1980). Ground water in the Rio Grande alluvium is under water-table conditions and is generally only a few feet below land surface, except in areas where the water level has declined because of direct pumpage from the alluvium or because of downward leakage into underlying heavily pumped aquifers. The alluvium has been completely drained in parts of downtown El Paso and Ciudad Juarez. Water in the Rio Grande alluvium ranges from slightly to moderately saline (1,000–10,000 mg/L), the freshest water occurring near the river where the alluvium is being recharged. Downward leakage of poor-quality water from the alluvium has caused problems in areas where the underlying bolson aquifers are being heavily pumped.

# Rio Grande

El Paso County is one of the leading agricultural counties in Texas. Virtually all agricultural production depends upon irrigation, primarily from the Rio Grande, augmented by pumpage from the Rio Grande alluvium during periods of surface-water shortage.

In the 1890's, Mexico filed a claim against the United States for damages due to river-water shortages in the Mesilla and El Paso–Juarez Valleys. The 1896 Embargo and subsequent Mexican Treaty of 1906 placed restrictions on the unlimited use of water from the upper reaches of the Rio Grande and eventually resulted in the 1938 Rio Grande Compact. The compact provides for scheduled deliveries of water to the Rio Grande by the states of Colorado and New Mexico.

To deliver the required water downstream, the U.S. Bureau of Reclamation constructed Elephant Butte Reservoir, diversion dams, canals, and open drains. This water delivery and recovery system, completed in 1925, was called the Rio Grande

Project. Landowners on the project are represented by the Elephant Butte Irrigation District of New Mexico and the El Paso County Water Improvement District No. 1 of Texas. In 1985, approximately 162,000 acre-ft (200 $hm^3$) of river water was diverted for irrigation use in El Paso County. In addition, about 17,000 acre-ft (21 $hm^3$) was diverted under contract to the City of El Paso for municipal use.

# Ground-Water Availability

Approximately 9.7 million acre-ft (11,960 $hm^3$) of theoretically recoverable fresh water is calculated to exist in Hueco Bolson deposits on the Texas side. The Mesilla Bolson deposits and Rio Grande alluvium together contain about 560,000 acre-ft (690 $hm^3$) of fresh water in storage under the Texas part of the lower Mesilla Valley. Even though this fresh water in storage is potentially recoverable, the proximity of poor-quality ground water to it requires the constraint that only 50 to 75 percent of the fresh water be pumped, to prevent the degradation of its chemical quality.

The fresh-water section of the Mesilla Valley thickens to the north and west in New Mexico to as much as 2,400 ft (732 m) and contains about 20 million acre-ft (24,660 $hm^3$) of fresh water. In the La Mesa area in New Mexico, as much as 34 million acre-ft (41,922 $hm^3$) of fresh water is theoretically available (Wilson and others, 1981).

The amount of ground water needed to supply projected demands in El Paso County exceeds the estimated annual effective recharge to the aquifers; therefore, a large portion will continue to be drawn from storage. Increasing water demands caused by a rapidly growing population will hasten the rate of water withdrawal from the aquifers, resulting in a depleted fresh-water source in the early to middle 21st century unless alternate sources can be acquired.

# Acknowledgments

This paper is summarized from a more extensive report by the author published as Texas Water Development Board Report 324. Appreciation is extended to Tom Cliett and Don White of El Paso for providing current ground-water use and availability data and for conducting a critical review of the TWDB report.

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# Regional Stratigraphy and Geomorphic Evolution of the Southern Hueco Bolson, West Texas, and Chihuahua, Mexico

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## Abstract

The Hueco Bolson is a segment of the Rio Grande Rift that formed as a result of late Tertiary Basin and Range deformation. The upper Tertiary Fort Hancock Formation and the upper Tertiary–Quaternary Camp Rice Formation compose most of the basin fill. Fort Hancock lithofacies include gravel and sand, deposited as proximal to distal alluvial fans near the basin margin, and clay and sandy clay, deposited in ephemeral lakes and saline playas near the basin center. The Fort Hancock Formation is separated from the overlying Camp Rice Formation by a regional unconformity. The unconformity records a period of extensive erosion that marks the integration of the ancestral southern and northern segments of the Rio Grande approximately 2.25 m.y. ago. Camp Rice Formation lithofacies include sand and gravel deposited by the ancestral through-flowing Rio Grande and its tributaries, sand deposited as a dune complex, sand and silt deposited as loess, and clay, sandy clay, and gypsum deposited in ephemeral lakes.

Paleoclimatic conditions can be inferred from both buried soils and from depositional environments. Calcic soils and the widespread development of ephemeral lakes suggest a semiarid environment during the late Tertiary–Quaternary.

## Introduction

The upper Tertiary Fort Hancock and the upper Tertiary–Quaternary Camp Rice Formation partly fill the Hueco Bolson (basin), West Texas, and Chihuahua, Mexico (fig. 1). The Hueco Bolson is underlain primarily by Lower Cretaceous limestone and sandstone (Albritton and Smith, 1965). As a result of early Tertiary Laramide deformation, these rocks were folded and thrusted northeastward toward the relatively undeformed Diablo Plateau. Extension related to formation of the Rio Grande Rift initiated during the late Oligocene–Miocene and continuing to the present further disturbed these rocks and resulted in a series of basins including the Hueco Bolson (Seager and others, 1984).

The Hueco Bolson extends from about 20 mi (32 km) northeast of El Paso, Texas, south and southeast for approximately 113 mi (180 km) to the Quitman Mountains, Texas. The structure of the basin is not well understood, although Mattick (1967) showed that near the New Mexico–Texas border the Hueco Bolson contains as much as 9,000 ft (2,728 m) of bolson sediments. Basin subsidence has continued into the Quaternary, as shown by normal faults displacing Quaternary units along the northeastern margin in the United States (Collins and Raney, 1989), along the southwestern margin of the basin in Mexico (Muehlberger and others, 1978), and along the eastern base of the Franklin Mountains north of El Paso, Texas (Machette, 1987).

A variety of fluvial and lacustrine sediments partly fill basins, such as the Hueco Bolson that formed during Tertiary to Recent extension. These Tertiary–Quaternary sediments make up the Santa Fe Group throughout much of New Mexico and West Texas (see Gile and others, 1981, and Seager and others, 1984, for discussions of this unit). The Fort Hancock and Camp Rice Formations compose the Sante Fe Group in the Hueco Bolson (fig. 2). For a detailed discussion of the stratigraphy and paleosols of the Santa Fe Group in the southern Hueco Bolson, including extensive references, see Gustavson (1989a; in press).

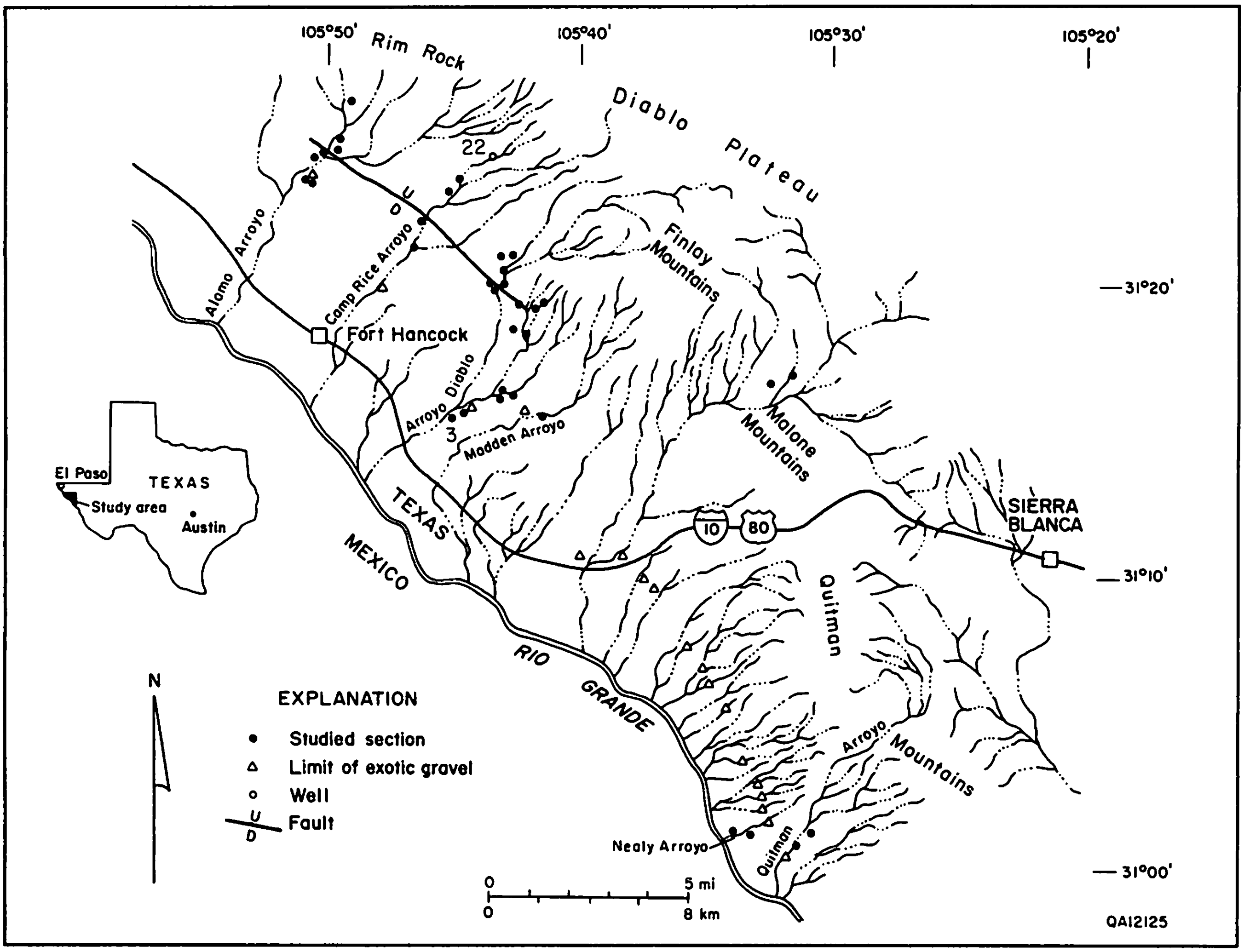


**Figure 1. Location map showing described sections, Texas Low-Level Radioactive Waste Disposal Authority hydrologic and stratigraphic test well no. 22, and the northeastern limit of exotic gravel in the Camp Rice Formation (limits of exotic gravel modified from Albritton and Smith, 1965). Numbered section locates figure 3 in this report.**

# Fort Hancock Formation

The Fort Hancock Formation was first described and named by Strain (1966) for outcrops in the Hueco Bolson near Fort Hancock, Texas. Albritton and Smith (1965) described similar "older" basin sediments from the southern third of the Hueco Bolson. Strain (1966, 1971) interpreted the Fort Hancock Formation as lacustrine clay, silty clay, and crossbedded silt that were periodically exposed during periods of aridity. Calcic paleosols also formed during periods of exposure. Stuart and Willingham (1984) recognized both lacustrine and fluvial sediments in the Fort Hancock Formation. Clay facies, including gypsum beds, were thought to be playa-lake deposits. Fluvial facies, which consist of mudstone and sandstone, were thought to have been deposited at playa margins or on levees or floodplains. Channelized sandstone facies were interpreted as having been deposited in low-sinuosity braided channels, and conglomerates were interpreted as forming lags, alluvial fans, or alluvial aprons along the bolson margin. Riley (1984) described a channel sand with epsilon cross stratification that led him to interpret that sandy facies in the type area of the Fort Hancock Formation were deposited by a meandering stream system. Although Riley (1984, p. 25) attributed clayey facies in the type section of the Fort Hancock to deposition

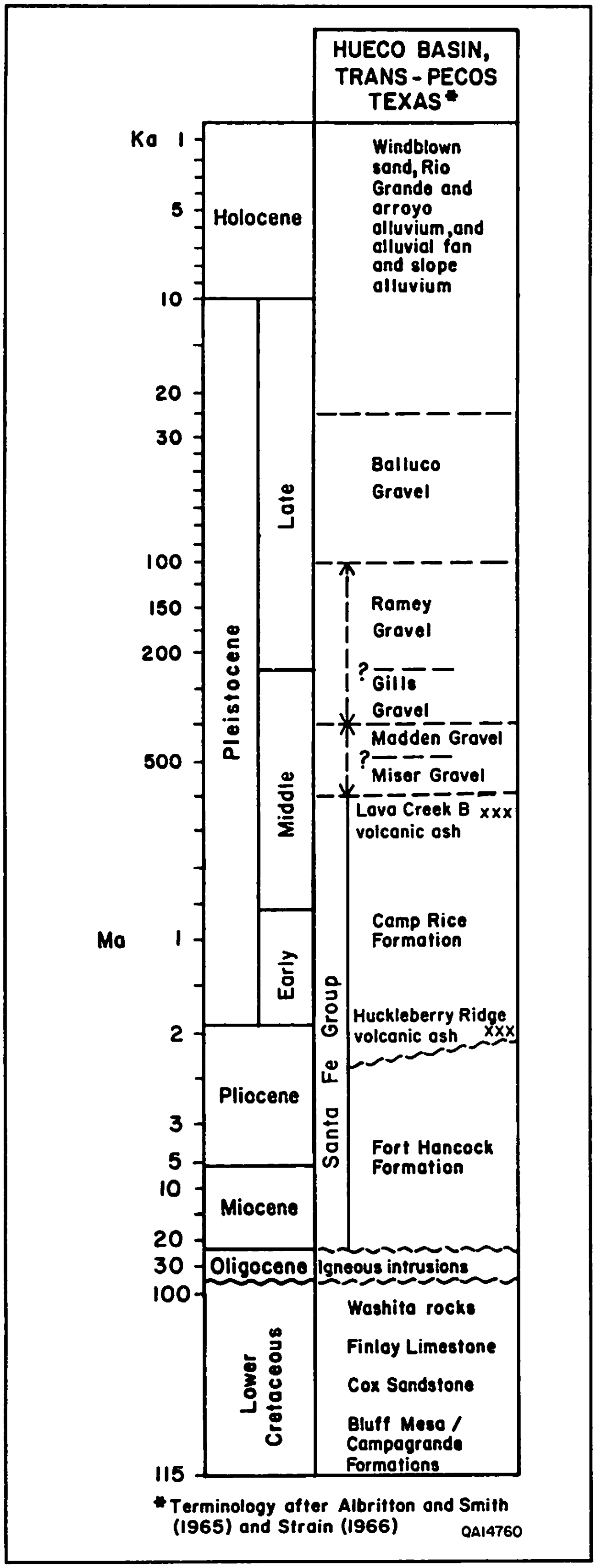


**Figure 2. Stratigraphic correlation chart for south-central New Mexico and the Hueco Bolson, Texas.**

by overbank flooding from a meandering stream, he did not completely reject the possibility of deposition of the clayey facies by lacustrine processes. Caliche nodules were recognized as evidence of subaerial exposure and the development of paleosols (Riley, 1984).

## Stratigraphy

Four lithofacies are present in exposures of the Fort Hancock Formation: (I) gravel; (II) gravel and sandy mud or sandy silt; (III) sandy mud and sandy silt; and (IV) clay, sandy clay, and gypsum (fig. 3). These facies reflect the textural gradation from basin margin to basin center (proximal- to distal-alluvial-fan to playa lake) (fig. 4). These lithofacies are also present in cores as 578-ft-(175-m-) thick upward-fining sequences recording lacustrine expansion over basin-margin alluvial fans. Flat-bedded, imbricated, clast-supported, pebble-to-boulder gravel lithofacies overlies Cretaceous bedrock and was probably deposited by ephemeral, locally braided streams as proximal-alluvial-fan sediments. Lithofacies composed of interbedded gravel and sandy mud and sandy silt were most likely deposited in settings that were transitional between proximal and distal alluvial fans. Thin, laterally extensive sandy silt and sandy mud beds are horizontally laminated and ripple and ripple-drift cross-laminated, suggesting that they are distal-alluvial-fan or fan-delta sediments that were deposited by shallow braided streams.

Pedogenesis has destroyed most sedimentary structures in the smectite-rich clay and sandy clay lithofacies (fig. 3), but locally preserved horizontally laminated clay and sandy clay with common mudcracks and local interbeds of gypsum suggest that these sediments were deposited in playa lakes. Lacustrine clays and sandy clays are commonly interbedded with fluvial sandy mud and silt. Transitional zones between lacustrine facies and overlying fluvial facies mostly coarsen upward, reflecting progradation of distal alluvial fans into the playa lakes. Contacts between fluvial facies and overlying lacustrine facies are commonly sharp, reflecting an abrupt change from fluvial to lacustrine sedimentation due to avulsion of fan distributaries.

## Age of the Fort Hancock Formation

Fossil vertebrate remains preserved in the type sections of the Fort Hancock Formation and the overlying Camp Rice Formation compose the Hudspeth local fauna of Blancan Age (Pliocene–Pleistocene) (Strain, 1966; see Riley, 1984). The

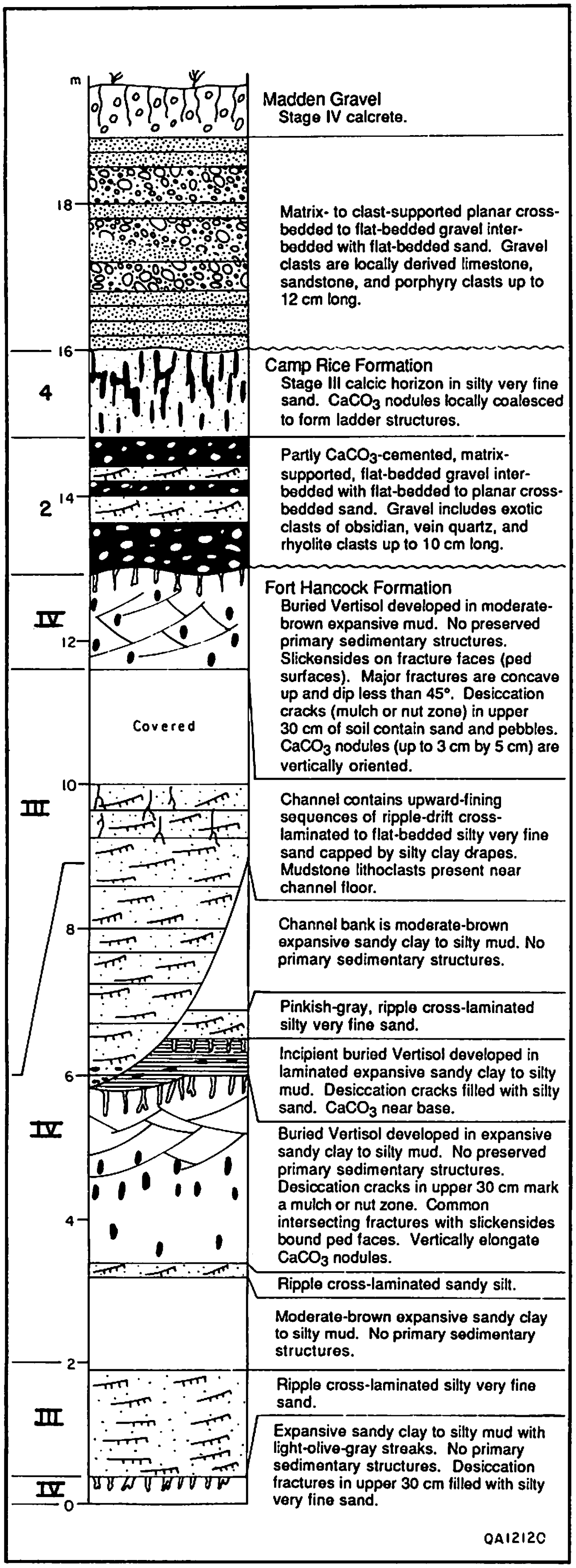


**Figure 3. Stratigraphic section of the upper Fort Hancock Formation and Camp Rice Formation exposed in Diablo Arroyo (see fig. 1 for location). Elevation of the base of the section is approximately 3,730 ft (1,137 m). Elevation of the top of the Fort Hancock Formation is 3,773 ft (1,155 m). Roman numerals identify lithofacies in the Fort Hancock Formation: III, sand, sandy mud, and sandy silt (distal alluvial fan); IV, clay and sandy clay (ephemeral lake). Arabic numerals identify lithofacies of the Camp Rice Formation: 2, sand and exotic gravel (braided stream); 4, coarse silt to fine sand (loess).**

Camp Rice Formation contains the Huckleberry Ridge volcanic ash (2.01 m.y. old) (Izett and Wilcox, 1982). Vanderhill (1986) obtained paleomagnetic data from the upper Fort Hancock Formation in the Hueco Bolson and placed the Fort Hancock in the late Gauss epoch. Consequently, the Fort Hancock Formation is middle Pliocene in age where it is exposed in the Hueco Bolson (fig. 2). Deposition of the Fort Hancock Formation ceased when through-flowing drainage of the Rio Grande developed during the late Pliocene about 2.25 m.y. ago (Gustavson, 1989a; in press).

# Camp Rice Formation

Strain (1966) described the Camp Rice Formation as fluvial sediments consisting mainly of channel gravel deposited by a through-flowing stream and sand, silt, and clay deposited as alluvial fans. Volcanic ash beds are preserved locally. Riley (1984) and Stuart and Willingham (1984) suggested that the Camp Rice Formation was deposited primarily by a braided stream carrying mostly bedload. They did not distinguish between axial- or through-flowing stream deposits and deposits that make up basin-margin alluvial fans or tributary streams.

## Stratigraphy

Lithofacies groups of the Camp Rice Formation comprise (1) sand and gravel, (2) sand and exotic gravel, (3) sand, (4) coarse silt and very fine sand, and (5) clay, sandy clay, and gypsum. Sand and gravel and sand and exotic gravel are volumetrically the most important lithofacies (fig. 3). Collectively, these lithofacies represent deposition by the ancestral Rio Grande and one or more tributaries (lithofacies 1 and 2), by eolian processes (lithofacies 3 and 4), and in an ephemeral lake (lithofacies 5) (fig. 5).

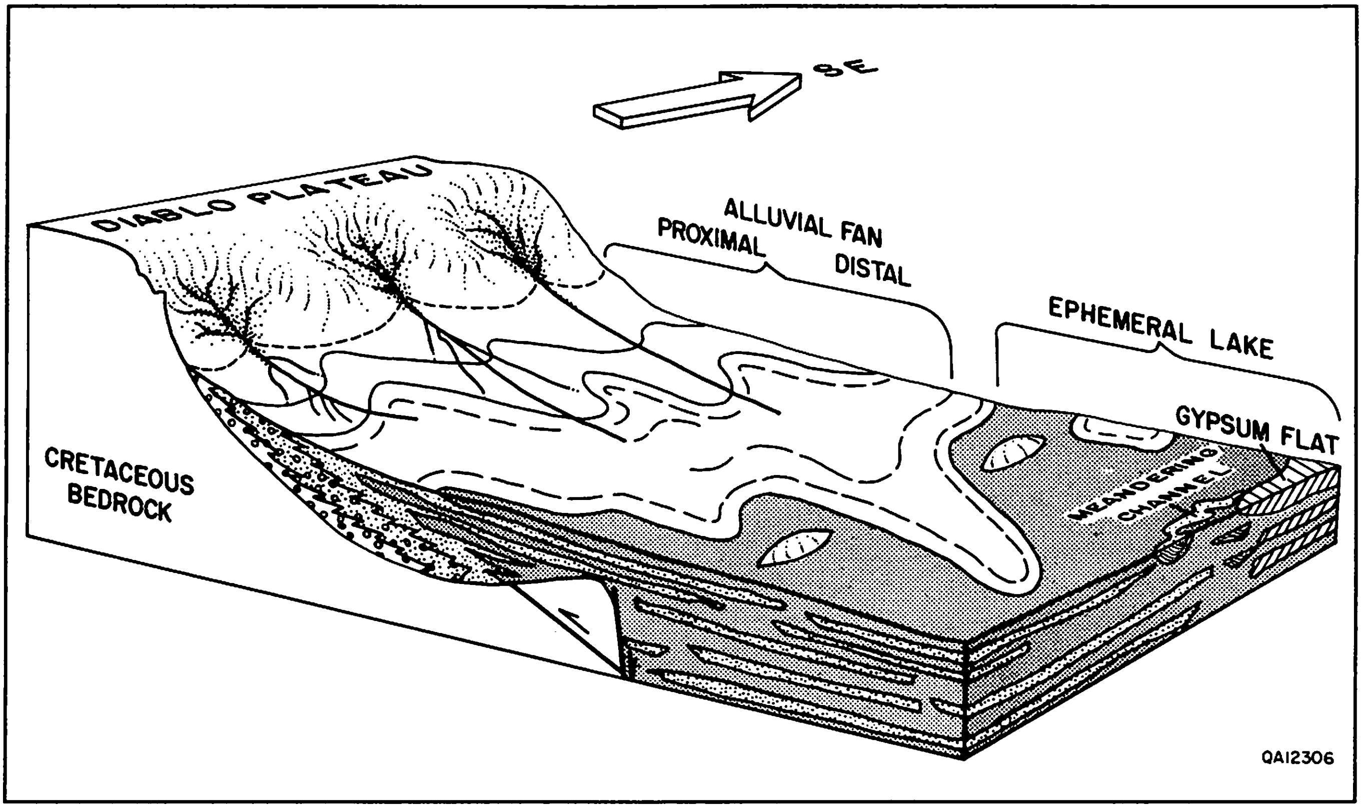


**Figure 4. Block diagram showing interpreted depositional environments of the Fort Hancock Formation in the southern Hueco Bolson. Primary lithofacies include alluvial fan and ephemeral lake sediments.**

Sand and gravel lithofacies comprise primarily flat-bedded to trough cross-bedded sand and gravel including $CaCO_3$ nodules and numerous mudstone lithoclasts derived from the Fort Hancock Formation. Channel fills fine upward. This facies fills channels that are oriented at a high angle to the Rio Grande and contain primarily sediment that was eroded from Fort Hancock strata. These sediments were laid down by short braided tributaries of the axial drainage of the Hueco Bolson, the ancestral Rio Grande.

The sand and exotic gravel lithofacies consists primarily of flat-bedded to trough crossbedded sand and gravel with secondary amounts of interbedded sand (fig. 3). Channel fills fine upward. Gravel, which is primarily composed of locally derived limestone, includes exotic clasts of obsidian, vein quartz, rhyolite, and other igneous, volcanic, and metamorphic clasts. These rock types, which are absent in the the southern Hueco Bolson area, were derived from the Rio Grande drainage northwest of the study area and probably largely outside of the Hueco Bolson. This facies is confined to the axial part of the Hueco Bolson. Sand and exotic gravel lithofacies were probably deposited by braided streams of the through-flowing ancestral Rio Grande along the axis of the Hueco Bolson.

Sand, very fine sand and coarse silt, and clay, sandy clay, and gypsum (lithofacies 3, 4, and 5) compose only a minor part of the Camp Rice Formation. The sand lithofacies consists of well-sorted planar crossbedded medium sand deposited as an eolian dune. The coarse silt to very fine sand lithofacies consists of several cycles of very fine sand to coarse silt with rare to common $CaCO_3$ nodules overlain by light-brown, angular, blocky to prismatic-fracturing muddy sand with common $CaCO_3$ nodules (buried B soil horizon). Calcium carbonate-filled root tubules (rhizocretions) are rare to common. No primary sedimentary structures are preserved in this lithofacies. Each cycle of fine sand and coarse silt to muddy sand represents an episode of eolian loess sedimentation on a stable vegetated surface followed by soil development. The clay, sandy clay, and gypsum lithofacies are very similar to the clay and sandy clay lithofacies of the Fort Hancock Formation. The similarities in color, texture, stratigraphy, and pedogenic structure between these lithofacies strongly suggest that, like the clay and sandy clay lithofacies of the

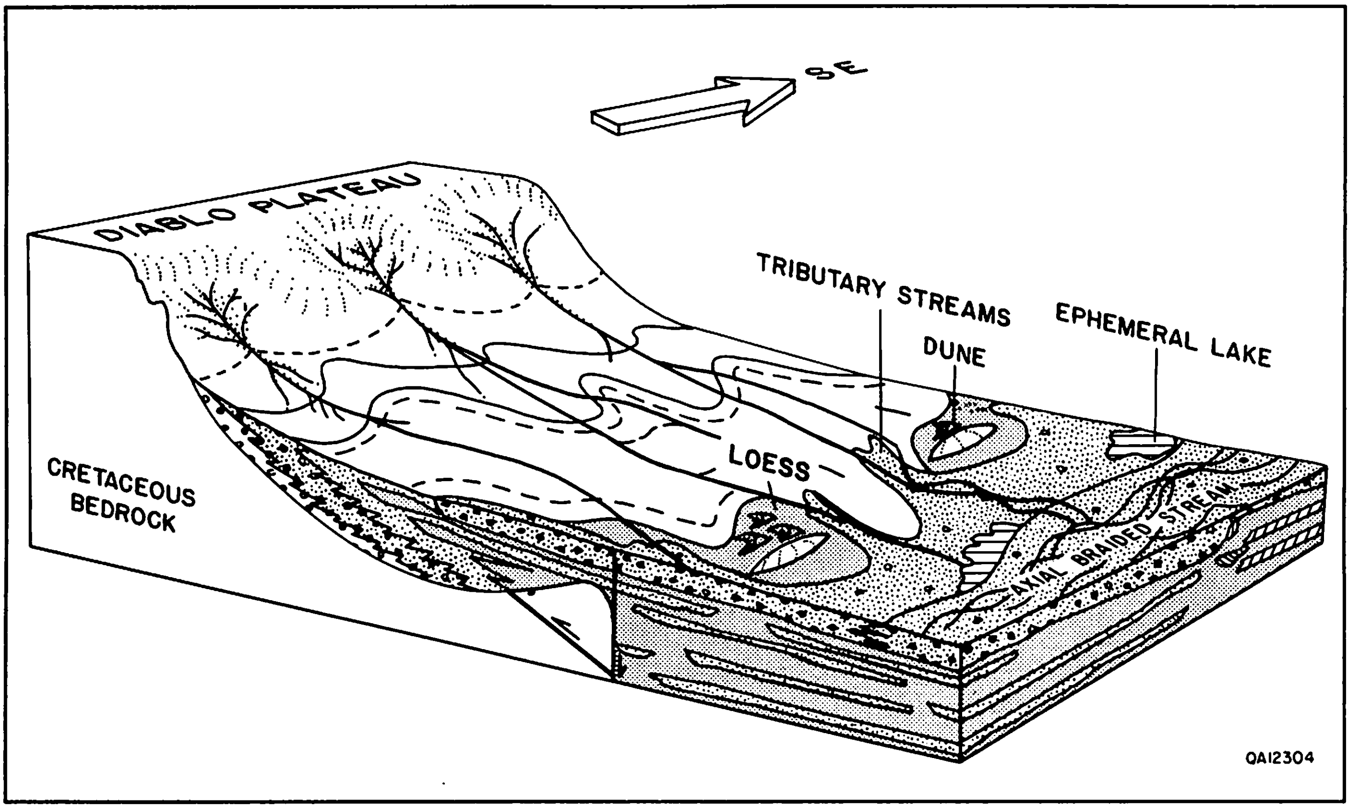


**Figure 5. Block diagram showing interpreted depositional environments and lithofacies of the Camp Rice Formation in the southern Hueco Bolson. Primary lithofacies include axial braided stream, gravel-bearing tributaries of the axial stream, and local dune and lacustrine sediments.**

Fort Hancock Formation, the clay and sandy clay lithofacies of the Camp Rice was deposited in an ephemeral lake. The thin gypsum bed suggests that this basin held a saline playa for a brief time.

## Age of the Camp Rice Formation

Strain (1966) thought that the lower part of the Camp Rice Formation, which also contains a Blancan vertebrate fauna, was Aftonian (early Pleistocene) and that the middle section, which contains an ash bed of the Pearlette family of volcanic ashes, was Kansan (middle Pleistocene). Strain (1966) did not speculate on the age of upper Camp Rice sediments other than to recognize them as Pleistocene. In its present usage the Blancan Land Mammal Age extends from the late Pliocene (4 m.y. ago) to earliest Pleistocene (1.5 m.y. ago) (Tedford, 1981). The Pearlette ash reported by Strain (1966) is now recognized as the Huckleberry Ridge Ash of the family of Pearlette ashes and has been dated at 2.01 m.y. (Izett and Wilcox, 1982). Consequently, the Camp Rice Formation is late Pliocene to early Quaternary in age (fig. 2).

## Paleovertisols

Recent vertisols are clayey soils that develop one or more of the following characteristics: (1) gilgai (surface microtopography); (2) deep, wide desiccation cracks (≥4 inches [10 cm] wide at a depth of 20 inches [50 cm]); (3) high bulk density when dry; (4) very slow hydraulic conductivity when moist; (5) slickensides on ped faces close enough to intersect at some depth between 10 and 40 inches (25 and 100 cm); or (6) wedge-shaped structural aggregates whose long axes dip between 10° and 60° at some depth between 10 inches (25 cm) and 40 inches (100 cm) (Soil Survey Staff, 1975). Montmorillonite, which is a member of the smectite family of clay minerals with a high coefficient of linear extensibility, usually composes at least 30 percent of these soils (Dudal and Eswaran, 1988).

Paleovertisols, which are rich in smectite clay, are commonly preserved in outcrops and core of lacustrine facies of the Fort Hancock and Camp Rice Formations (Gustavson, 1989b; in press b) (fig. 3). Soil structures in buried vertisols of the Fort Hancock Formation normally include (1) mulch zones that mark the former surface layer of buried vertisols and comprise compacted, angular, blocky granules, (2) near-vertical desiccation cracks (polygonal in plan view) as much as 60 inches (150 cm) deep and 0.6 inch (1.5 cm) wide, which may be filled with sand from an overlying unit or with granules from the mulch zone, (3) common intersecting fractures with slickensides that bound small (<0.7-ft [<0.2-m]), blocky soil aggregates and large (1- to 10-ft [0.3- to 3-m] long), wedge-shaped aggregates, and (4) $CaCO_3$ filaments and nodules, which characterize buried calcic vertisols. Modern vertisols develop largely as the result of repeated cycles of swelling and shrinking of expansive clays due to hydration and desiccation (Soil Survey Staff, 1975; Wilding and Tessier, 1988). Thus, paleovertisols in the Fort Hancock and Camp Rice probably also are the result of cyclic swelling and shrinking of lacustrine facies during episodes of rainfall or flooding and subsequent desiccation of the sediments in ephemeral lakes that occupied much of the floor of the Hueco Bolson. Pedogenic $CaCO_3$ filaments and nodules in paleovertisols in lacustrine and other lithofacies are similar to $CaCO_3$ structures in modern Stage I and II calcic soils, which develop mainly in an arid to semiarid climate (Machette, 1985).

# Tertiary–Quaternary Geomorphic Evolution of the Hueco Bolson

Major events in the depositional history of the Hueco Bolson have been closely tied to its physiographic evolution since formation of the basin during the late Oligocene–Miocene. Sedimentation into a closed basin, but without the contribution of the northern Rio Grande, persisted through most of the earlier (Miocene) history of the basin. This is reflected in the lower part of 8,250 ft (2,500 m) of sediments preserved near the northern limit of the basin west of the Hueco Mountains (Mattick, 1967; Seager and others, 1984). In Pliocene and possibly latest Miocene time the northern ancestral Rio Grande discharged into the Hueco Bolson, but the basin remained internally drained, as indicated in the predominance of fine-grained clastic sediments near the basin center and coarser clastics near the basin periphery (Fort Hancock Formation of Strain, 1966; Gustavson, 1989a; in press a).

Through-flowing drainage in the Hueco Bolson, which is now a segment of the Rio Grande, developed during the late Pliocene prior to 2.01 m.y. ago (Gile and others, 1981; Seager and others, 1984). Through-flowing streams incised older sediments and deposited coarse sand and gravel, including "exotic" clasts derived from crystalline rocks that crop out only in areas northwest of the Hueco Bolson (Camp Rice Formation of Strain, 1966; Hawley and others, 1969; Gustavson, 1989a; in press a).

Although many authors have recognized that the lithologic differences between the Camp Rice and Fort Hancock Formations in the Hueco Bolson reflect different depositional environments and, in part, different sediment source areas, precisely how this change in depositional environment came about is poorly understood. Strain (1966) did not elaborate on the transition from lacustrine (Fort Hancock) to fluvial (Camp Rice) deposition other than to say that it represented the change from lacustrine sedimentation in a closed basin to sedimentation by a through-flowing stream. Strain (1971) suggested that through-flowing drainage of the Hueco Bolson developed by (1) overflow of Hueco Bolson lake waters, which were part of the larger Lake Cabeza de Vaca, southward into the Red Light and Presidio Bolsons, (2) headward erosion from the Presidio Bolson, or (3) a combination of the two. Hawley and others (1969) suggested that the change from lacustrine sedimentation to fluvial sedimentation developed progressively from the Palomas Basin on the north to the Hueco Bolson on the south. However, that before integration with the lower Rio Grande system south of the Quitman Mountains the upper Rio Grande fed large lakes in several basins in the border region of Texas and New Mexico (Hueco, Mesilla, and Tularosa Bolsons = Lake Cabeza de Vaca of Strain [1971] and the Red Light Bolson). For additional discussions see Hawley (1975, 1981).

Integration of the northern and southern segments of the ancestral Rio Grande to form a through-flowing stream in the Hueco Bolson is marked by an abrupt upsection change in depositional style across a regional disconformity from low-energy ephemeral lake, saline playa, and distal alluvial fan facies, which were deposited in a closed desert basin as the Pliocene Fort Hancock Formation, to high-energy fluvial facies deposited by the through-flowing Rio Grande and its tributaries as the Plio-Pleistocene Camp Rice Formation. The Hueco Bolson formed during Tertiary Basin and Range deformation and was a closed basin throughout its early history. Fossil and paleomagnetic data indicate that upper Fort Hancock strata are 2.48 to 3.4 m.y. old. These sediments consist mostly of fine sand and silt deposited as distal alluvial fans and silty clay and clay and local gypsum deposited in ephemeral lakes and saline playas, respectively.

Basin infilling ended as the southern Hueco Bolson drainage divide was overtopped, joining the northern and southern segments of the Rio Grande. Exposures of Fort Hancock lacustrine strata as high as 4,050 ft (1,234 m) indicate the minimum elevation of the drainage divide. The Rio Grande incised the divide area and the southern Hueco Bolson by as much as 400 ft (130 m) prior to deposition of Camp Rice fluvial sediments. Incision produced a disconformity with local elevation differences of up to 120 ft (35 m) and a regional southeast slope of about 8 ft/mi (1.5 m/km). Camp Rice lithofacies overlie the unconformity and include mostly fluvial sand and gravel, which contain the Huckleberry Ridge ash (2.01 m.y.), and locally eolian sand and silt, which contain the Lava Creek B ash (0.62 m.y.). Sand and gravel lithofacies with exotic gravel derived from rock types that occur only northwest of the study area are confined to the axial part of the basin and were deposited by the through-flowing Rio Grande. Sand and gravel lithofacies, with gravel composed primarily of mudballs derived from the Fort Hancock Formation and lacking exotic gravel clasts, were deposited by ephemeral tributaries of the Rio Grande.

Unification of segments of the Rio Grande post-dates the Fort Hancock Formation, which may be as young as 2.48 m.y. Basin incision and deposition of nearly 130 ft (46 m) of Camp Rice Formation sediments occurred prior to deposition of the Huckleberry Ridge Ash 2.01 m.y. ago. Thus, integration of Rio Grande drainage probably occurred between 2.48 m.y. and 2.01 m.y. ago or about 2.25 m.y. ago.

# Acknowledgments

I thank Bob Blodgett, John Hawley, Tucker Hentz, Amanda Masterson, Jay Raney, and Charles Kreitler for constructive reviews of this manuscript. Illustrations were prepared by Yves Oberlin and Richard Dillon. This work was supported by the Texas Low-Level Radioactive Waste Disposal Authority under Interagency Contract Number IAC (90-91)-0268.

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# Saturated-Zone Hydrology of South-Central Hudspeth County, Texas

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## Abstract

Hydrologic investigations of the saturated zone have been conducted in south-central Hudspeth County, Trans-Pecos Texas. The principal study area is in the Hueco Bolson, a fault-bounded desert basin that developed in the late Tertiary. Ground water in the principal study area is found in Hueco Bolson silts and sands at depths of 361 ft (110 m) and 478 ft (146 m) and at depths of 592 ft (180 m) in Cretaceous limestones. The unsaturated zone consists of approximately 50 ft (15 m) of alluvial silt, sand, and gravel underlain by 300 to 500 ft (91 to 152 m) of lacustrine and fluvial clay, silt, and fine sand. The scope of this investigation included (1) evaluating ground-water resources in the area, (2) determining ground-water flow paths and velocities, and (3) testing hydrologic hypotheses using ground-water flow models.

Transmissivities of aquifers in bolson and Cretaceous strata, based on nine aquifer tests and two slug tests, range from approximately 0.35 to 81.4 $ft^2/d$ (0.033 to 7.56 $m^2/d$); corresponding permeabilities range from 0.0056 to 0.79 ft/d (0.0017 to 0.24 m/d). A composite potentiometric surface based on water levels measured in all available wells and hydrologic interconnection of the Diablo Plateau aquifer, Hueco Bolson silt and sand aquifer, and Rio Grande alluvium aquifer indicates that ground water is recharged on the Diablo Plateau and flows to the south and southwest toward the Rio Grande beneath the bolson pediment. Information on water chemistry, particularly tritium, carbon-14, and total dissolved solids, generally supports the interpreted flow pattern; some discrepancies can be related to paleohydrologic effects associated with the incision of the Rio Grande during Quaternary time.

## Introduction

Investigations have been conducted in Trans-Pecos Texas to characterize both the regional and local hydrology of an area selected by the Texas Low-Level Radioactive Waste Disposal Authority as a proposed site for a low-level radioactive waste repository. This paper summarizes findings of regional and local hydrologic studies of the saturated zone.

The principal study area is approximately 40 mi (65 km) southeast of El Paso in the Hueco Bolson, a fault-bounded desert basin that developed in the late Tertiary. Fine-grained lacustrine and fluvial sediments of the Hueco Bolson were deposited over a basement of mostly Cretaceous shallow-marine strata. Ground water in the regional study area is found in three aquifers: the Diablo Plateau aquifer (predominantly Cretaceous limestones and sandstones), the Hueco Bolson silt and sand aquifer, and the Rio Grande alluvial aquifer (Mullican and others, 1989). Figure 1 depicts the general geometry of the different hydrostratigraphic units from the Diablo Plateau to the Rio Grande. Cretaceous strata crop out locally near the northwest-oriented Campo Grande fault trend. Along this fault trend Cretaceous strata are displaced against bolson deposits southwest of the fault. The southwestern edge of the Diablo Plateau shows a flexure of Cretaceous strata that dip beneath bolson deposits in the central part of the area.

Ground-water resources in the vicinity of the principal study area are limited by three key factors: (1) costs of drilling and completing wells, (2) costs of producing water at depths typically greater than 400 ft (122 m), and (3) very low productivity of aquifers.

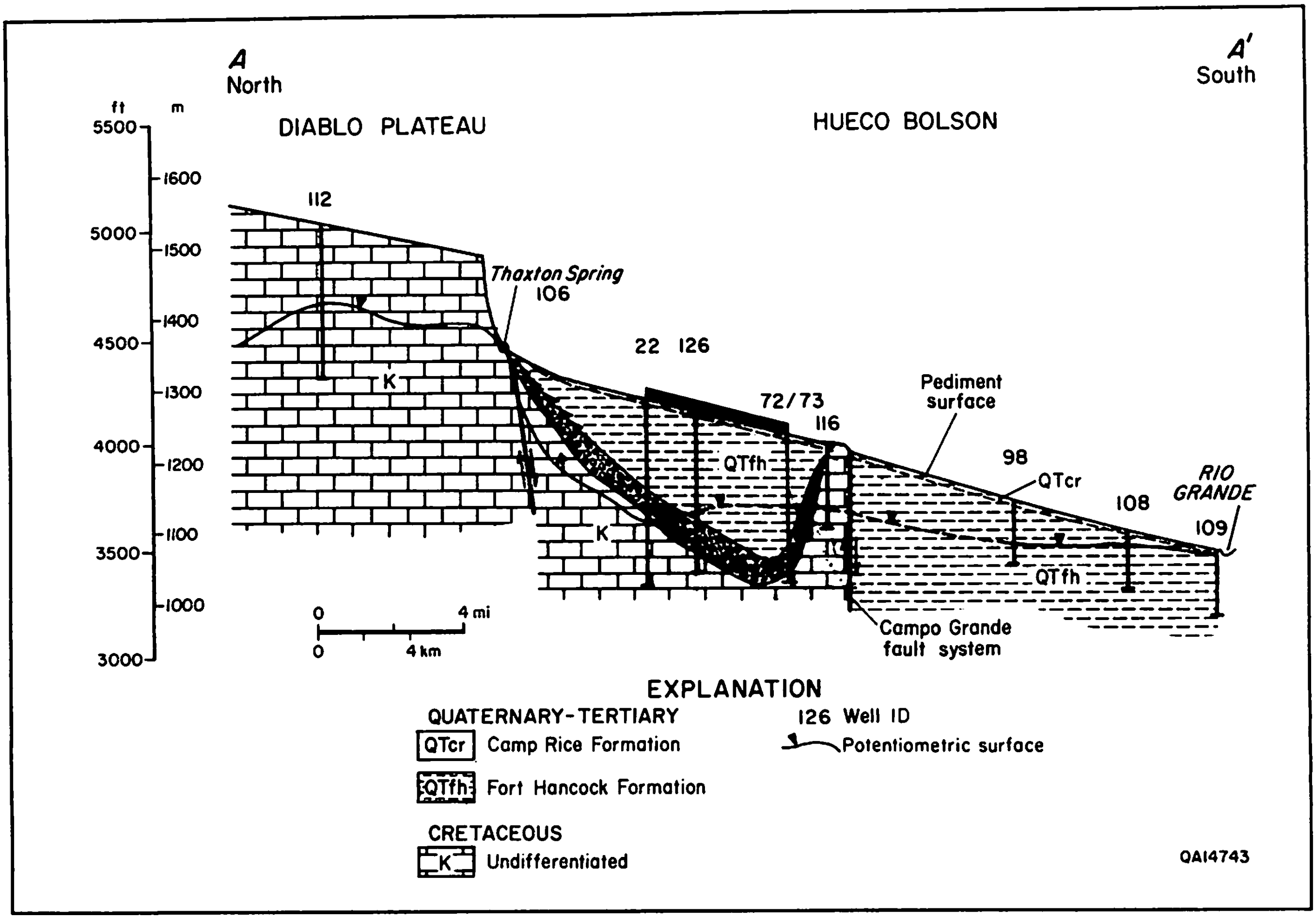


**Figure 1. Hydrologic cross section illustrating relationship between potentiometric surface and stratigraphy from the Diablo Plateau to the Rio Grande. Location of cross section shown in figure 2.**

# Hydrologic Characteristics

Transmissivities measured in wells producing from the Diablo Plateau aquifer and Hueco Bolson silt and sand aquifer, based on nine pumping tests and two slug tests, range from approximately 0.35 to 81.4 $ft^2/d$ (0.033 to 7.56 $m^2/d$); corresponding hydraulic conductivities range from 0.0056 to 0.791 ft/d (0.0017 to 0.241 m/d). All wells tested in the Diablo Plateau aquifer and Hueco Bolson silt and sand aquifer show semidiurnal water-level fluctuations reflecting barometric pressure variations that are typical of confined and semiconfined aquifers.

At the principal study area, the water table lies at depths ranging from 361.5 to 592 ft (110 to 180 m) below land surface. Ground-water flow, inferred from a potentiometric-surface map constructed from measurements of static water levels in the Diablo Plateau aquifer, Hueco Bolson silts and sand aquifer, and Rio Grande alluvium aquifer (fig. 2), is generally from northeast to southwest, despite deviations created by lateral permeability changes. It is assumed that the three hydrostratigraphic units are hydrologically continuous from the Diablo Plateau to the Rio Grande, representing regional recharge and discharge areas, respectively. A ground-water divide is oriented northwest just north of the Diablo Plateau escarpment (fig. 2).

Although the regional gradient in the study area is toward the Rio Grande, water-level elevations in several wells producing from the Diablo Plateau aquifer and Hueco Bolson silt and sand aquifer located near the Campo Grande fault trend are higher than those measured in wells within the principal study area (fig. 2). The apparently low hydraulic heads near the principal study area create a relatively steep southwestward gradient between the Diablo Plateau and the principal study area, and a relatively low, northward gradient from the area along the Campo Grande fault trend toward the principal study area (fig. 2).

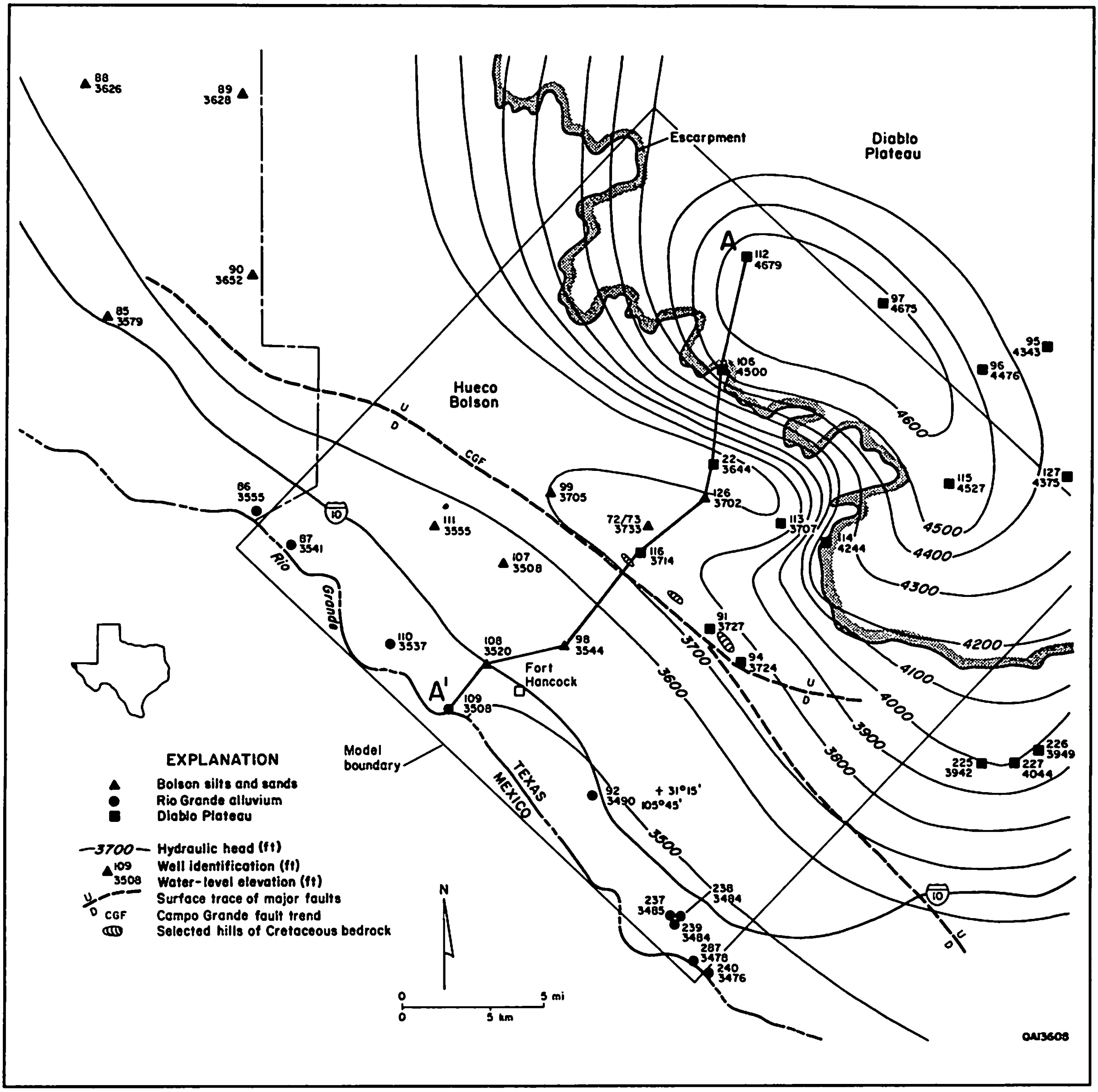


**Figure 2. Map of potentiometric surface in the study area. Efficient hydrologic connection between the different aquifer systems was assumed in the preparation of this map. A–A' is line of cross section illustrated in figure 1.**

There are two possible explanations for observed hydraulic-head distributions. The potentiometric high near the Campo Grande fault (figs. 1 and 2) might suggest that recharge occurs along the Cretaceous outcrop near the Campo Grande fault. However, because of the high potential for evapotranspiration and the great depth to the water table below land surface in the Hueco Bolson, the probability that infiltrating water would reach the water table, even in the Cretaceous outcrops along the fault, is low. Alternatively, the relatively high hydraulic heads near the fault might indicate that ground water flows westward from the Diablo Plateau east-northeast of the principal study area through a high-transmissivity facies that lies south and east of the principal study areas. This latter interpretation of hydraulic-head data is supported by the spatial distribution of total dissolved solids (TDS) and tritium concentrations measured in different wells in the area (fig. 3). Well 114 along the Diablo Plateau escarpment shows relatively low TDS and high tritium concentration, and wells 91, 94, 113, and 116 south and southeast of the principal study area show much lower TDS than wells 22 and 126 at the principal study area. Wells 91, 94, 113, and 114 in the Hueco Bolson and wells 96 and 112 on the Diablo Plateau show tritium concentrations above detection limit (0.8 tritium units [TU]), indicating some recent recharge water, whereas well 116 and those at the principal study area do not. The high tritium concentrations in well 114 indicate rapid recharge, possibly along faults and fractures in the vicinity of the escarpment. Cretaceous strata in this area are associated with the dome structure of the Finlay Mountains and are extremely fractured. The spatial distributions of TDS and tritium concentrations (fig. 3) suggest increasing ground-water ages along a preferential flow path from the eastern Diablo Plateau toward the Campo Grande fault.

On the basis of a series of steady-state simulations using a two-dimensional, planar flow model (Mullican and Senger, 1990), it was indicated that the main controls on the regional flow pattern (fig. 1) are (1) preferential recharge in the eastern part of the study area on the Diablo Plateau (Finlay Mountains), (2) relatively high permeability of Cretaceous strata along the Campo Grande fault, (3) distribution of bolson deposits with high and low permeabilities along east-west trends north of the Campo Grande fault, and (4) displacement of Cretaceous strata north of the central part of the Campo Grande fault against bolson deposits to the south of the fault (fig. 1) forming a low-permeability barrier to ground-water flow toward the Rio Grande.

Late Quaternary evolution of the Rio Grande valley resulted in the incision of the stream bed between 10,000 and 25,000 years ago by as much as 200 ft (61 m) (Gile and others, 1981). Within the last 10,000 years, the base level of the Rio Grande has remained stationary. This suggests that water levels were significantly higher in the Hueco Bolson in the recent geologic past. If the potentiometric surface were 200 ft (61 m) higher than the present surface in the area of Rio Grande incision, then the potentiometric surface would be relatively flat over much of the Hueco Bolson study area, perhaps even intersecting land surface. It can be assumed, however, that water levels on the Diablo Plateau may not have been significantly higher than those of today because additional ground water on the plateau would have been discharged through springs and seeps along the escarpment. The observed hydraulic head distribution, however, is assumed to represent steady-state equilibrium between recharge and discharge rates; that is, hydraulic heads in the modern ground-water flow system are assumed to have fully adjusted to the change in base level of the Rio Grande.

Carbon-14 ages of water samples from wells south of the Campo Grande fault (well no. 98: 6,000 yr; well no. 107: 15,000 yr; well no. 111: 7,000 yr) are noticeably younger than those at the principal study area (well no. 126: 29,000 yr) (Fisher and Mullican, 1990). Prior to the Rio Grande incision, water levels might have been much closer to land surface than those observed today, particularly near the river. In addition, recharge rates in the past were probably higher than those of today because of a wetter climate (Baumgardner, 1990). The lower evapotranspiration potential and the shorter travel time for infiltrating surface water to reach the water table in the recent geologic past may explain these relatively young ground waters south of the Campo Grande fault. Lateral ground-water flow in the Hueco Bolson was probably slower in the past because of the lower hydraulic-head gradient associated with the higher base level of the Rio Grande.

# Summary

Ground water in the principal study area in the Hueco Bolson is found at depths of 361.5 ft (110 m) and 478 ft (146 m) in Hueco Bolson silts and sands and 592 ft (180 m) in Cretaceous limestones.

In matching the composite potentiometric surface of the three aquifer units with the numerical results of a series of steady-state simulations, using a two-dimensional, planar, flow model, we found that the main controls of the regional flow pattern are (1) preferential recharge from the eastern part of the regional study area on the Diablo Plateau (Finlay Mountains), (2) relatively high permeability of Cretaceous strata along the Campo Grande fault, (3) relatively high permeability of bolson deposits north of the Campo Grande fault, separated from Cretaceous strata along the Campo Grande fault by a low-permeability zone, and (4) a low-permeability zone north of the central part of the Campo Grande fault where displacement between outcropping Cretaceous strata and bolson deposits to the south acts as a low-permeability zone for ground-water flow toward the Rio Grande.

The inferred distribution of permeability zones causes preferential flow from the eastern Diablo

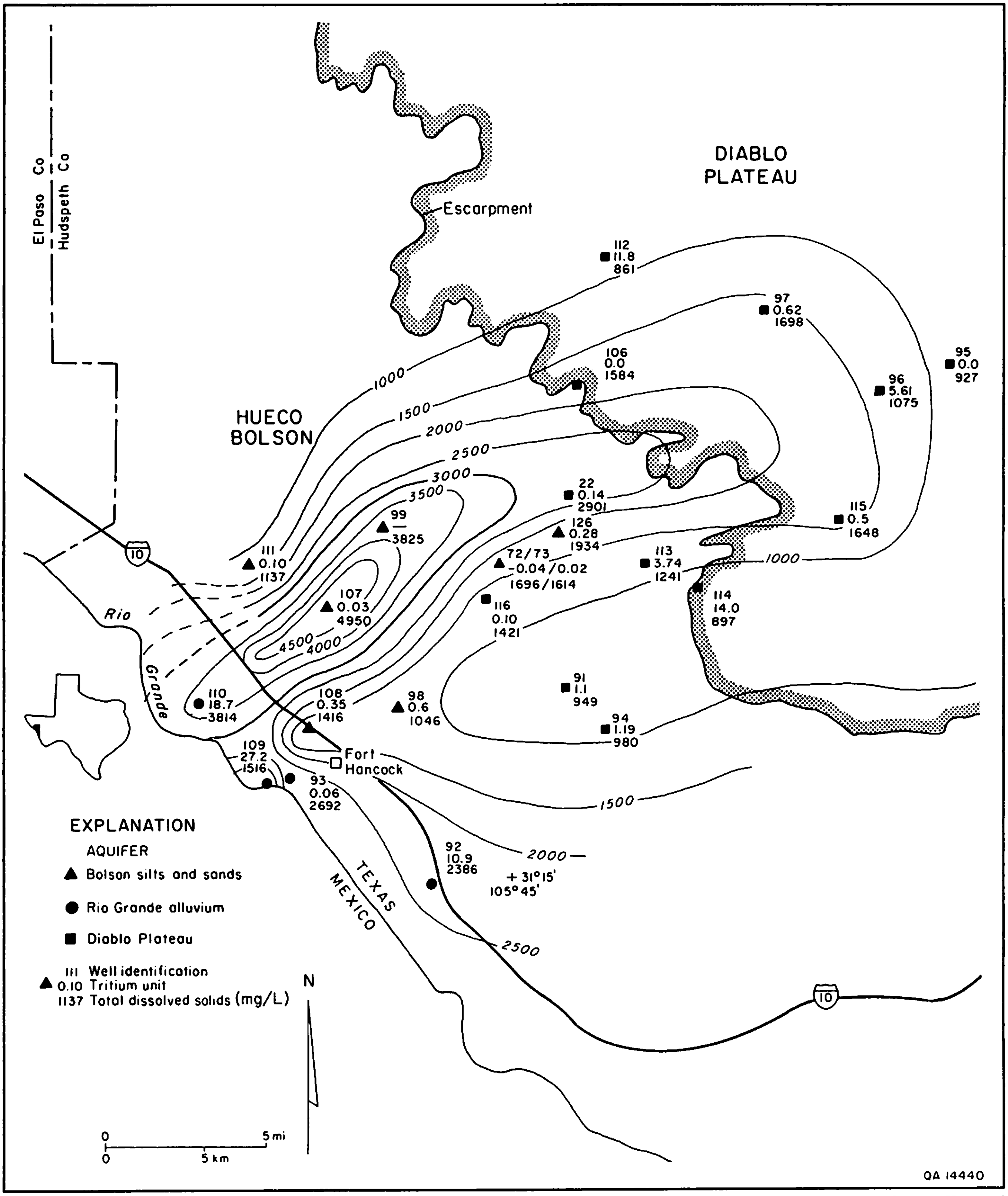


**Figure 3. Distribution of total dissolved solids (TDS) and tritium concentrations of water samples collected at the different wells in the area.**

Plateau toward Cretaceous outcrops along the Campo Grande fault, creating the observed nose of high hydraulic heads south of the principal study area. The relatively low hydraulic heads near the principal study area are caused by preferential drainage along permeable bolson deposits to the west and southwest toward the Rio Grande. Data on water chemistry, particularly tritium, total dissolved solids, and to a lesser extent carbon-14 age data, generally support the interpreted flow pattern; some discrepancies between flow interpretation based on hydrochemical data and hydrologic data may be related to paleohydrologic phenomena associated with the incision of the Rio Grande during Quaternary time.

# Acknowledgments

This research was funded by the Texas Low-Level Radioactive Waste Disposal Authority under Interagency Contract Number IAC(90-91)0268. The conclusions of the authors are not necessarily approved or endorsed by the Authority.

Special acknowledgments are given to the local residents of the Fort Hancock community for their help in locating, testing, and sampling water wells in the area. Dennis Walker, Joe Galvan, and the late Scott Wilkey have continually assisted in various ground-water investigations, and to them we extend our heartfelt thanks. Byrl Binkley and R. C. Corona were our drillers and well pullers, and we appreciate their patience in dealing with our demands.

A. R. Dutton, T. F. Hentz, C. W. Kreitler, and W. R. Muehlberger provided helpful reviews of this manuscript. Illustrations were drafted under the supervision of R. L. Dillon.

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# Origin of Solutes in Ground Waters from the Diablo Plateau, Hueco Bolson, and Rio Grande Aquifers, Trans-Pecos Texas

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## Abstract

Ground-waters from three major aquifers, water extracted from unsaturated sediments, soil leachates, and soil and core samples were analyzed to investigate the hydrochemical history of ground water in the southeastern Hueco Bolson and southwestern Diablo Plateau. The Hueco Bolson and Diablo Plateau aquifers contain mainly sodium-sulfate water that derived solutes by calcite, dolomite, and gypsum dissolution coupled with exchange of aqueous calcium for sodium on ion exchange sites; these waters evolved from calcium-bicarbonate to sodium-bicarbonate to sodium-sulfate ground-water types. Rio Grande ground water is dominated by sodium and chloride derived from dissolution of salts that precipitate on irrigated fields during times of high evaporation and from return flow of irrigation water. Major compositional characteristics of these ground waters are apparently acquired early in their flow history, largely as a result of reactions in the unsaturated zone.


## Introduction

The southeastern Hueco Bolson and southwestern Diablo Plateau, Trans-Pecos Texas, lie within the northern Chihuahuan Desert and have a subtropical arid climate characterized by high mean annual temperature, low mean annual precipitation, and marked daily and seasonal fluctuations in both temperature and precipitation. The Hueco Bolson is a fault-bounded basin that developed at approximately 24 m.y. in response to east-northeast extension and normal faulting (Henry and Price, 1985). The Bolson is filled with coarse to fine siliciclastic sediment derived from local highlands and imported by the ancestral Rio Grande. The Diablo Plateau consists primarily of Cretaceous limestone and sandstone. The Plateau was not deformed during basin development and now forms a topographic high several hundred feet above the bolson surface and is separated from it by a sharp escarpment.

Water-bearing strata in the area have been divided into three principal aquifers (**Mullican and Senger**, this volume, p. 37). The Diablo Plateau aquifer within our study area comprises Cretaceous limestones and sandstones that are exposed on the Diablo Plateau and extend underneath bolson sediments. The Hueco Bolson silt and sand aquifer includes all saturated bolson fill; water production is mainly from sand and silt lenses between mudstones and clay layers. A third unit, the Rio Grande alluvial aquifer, forms a narrow band along the Rio Grande. Ground-water flow, inferred from water level measurements, is generally from the southwestern Diablo Plateau toward the Rio Grande (**Mullican and Senger**, this volume, p. 37).

## Information Sources

Ground-water samples were collected from all available wells and springs and analyzed to augment hydrologic studies. In addition, water extracted from unsaturated bolson sediments and leachates from soil samples were analyzed for major and minor dissolved constituents, and the mineralogy and ion exchange capacity of soil and sediment samples were measured. Sample sources, analytical methods, and results are presented in Fisher and Mullican (1990).

# Major Features of Ground-Water Chemistry

Salinity varies widely within each major aquifer; however, the most saline ground waters occur in a northeast-trending band that is approximately parallel to the regional potentiometric gradient (compare fig. 1 with fig. 2 of **Mullican and Senger**, this volume, p. 39). The distribution of predominant cations and anions (fig. 1) shows that samples from the Diablo Plateau aquifer collected from wells away from the escarpment have sodium and bicarbonate as the major cation and anion, respectively. Most samples from wells or springs on the bolson pediment or near the edge of the Diablo Plateau, whether from the Hueco Bolson or the Diablo Plateau aquifers, have sodium and sulfate as the predominant cation and anion, respectively. One well from the Diablo Plateau aquifer near the toe of the escarpment produces calcium-sulfate water. Samples from Rio Grande alluvium have sodium and chloride as the predominant cation and anion, respectively. All water samples from the unsaturated zone have sodium and bicarbonate as the major cation and anion, respectively. Calcium and bicarbonate are the most abundant ions in all soil leachates, with one exception; sodium and chloride predominate in the leachate from soil on a field irrigated by Rio Grande river water.

Mineral saturation states computed by the geochemical modeling program SOLMNEQ (Kharaka and Barnes, 1973) show that ground water from the Hueco Bolson and Diablo Plateau aquifers is saturated with calcite and dolomite, whereas water from the Rio Grande alluvium is slightly oversaturated with both of these minerals (Fisher and Mullican, 1990). Despite high sulfate concentrations in many of the waters, none is saturated with gypsum. Most ground waters, as well as soil leachates and water from the unsaturated zone, contain high dissolved silica concentrations that result in oversaturation of ground water with respect to all common silicate and aluminosilicate minerals. Such high concentrations of dissolved silica are attributed not to mineral-water reactions but to dissolution of the amorphous silica phytoliths that can constitute several weight percent of the plants that grow in arid climates (Iler, 1979).

Hueco Bolson and Diablo Plateau ground waters show an excess of sodium relative to chloride (fig. 2) and a direct relation between excess sodium (sodium minus chloride) and sulfate (fig. 3). A sodium/chloride molal ratio approximately equal to one is typically attributed to halite dissolution, whereas a ratio greater than unity is commonly interpreted as reflecting sodium added from silicate weathering reactions (Mackenzie and Garrels, 1966; Meybeck, 1987). Silicate weathering is unlikely to significantly affect Hueco Bolson and Diablo Plateau ground waters because (1) water that derives solutes primarily by silicate weathering should have bicarbonate as the most abundant anion, whereas bicarbonate is minor in all but one sample analyzed in this study, (2) water involved in typical silicate weathering reactions should have silica concentration controlled by saturation with quartz or clay minerals (Mackenzie and Garrels, 1966), whereas all of the ground-water samples analyzed in this study are highly oversaturated with respect to all common silicate phases (Fisher and Mullican, 1990), and (3) the amount of unstable silicate minerals in surface sediments or bolson fill is insufficient to significantly affect ground-water chemistry.

# Hydrochemical Evolution of Ground Waters

Distributions of major cations and anions with respect to the regional potentiometric surface, results of mineral-water equilibria evaluations, and observed relations among concentrations of major dissolved species suggest that a simple set of chemical reactions controls ground-water compositions. Dissolution of calcite and dolomite to saturation is indicated by ground-water saturation indices for these minerals, whereas dissolution of gypsum is indicated by high dissolved sulfate concentrations. Ion exchange is suggested by the excess of sodium relative to chloride and by the relation between excess sodium and sulfate.

The hypothesis that these reactions largely control ground-water chemistry can be readily tested. Dissolution of calcite and dolomite releases calcium, magnesium, and bicarbonate to solution. If exchangeable sodium is present in unsaturated sediments or in the aquifer matrix, some will be replaced by aqueous calcium or magnesium, thereby adding sodium to the ground water and removing an equivalent amount of calcium or magnesium. Dissolution of gypsum will provide additional calcium as well as sulfate, and the increased aqueous calcium concentration will lead to further exchange of sodium for calcium. Quantitatively, the stoichiometry of these reactions requires that a plot of $2Na^{+}+Ca^{+2}+Mg^{+2}$ versus $SO_4^{-2}$ for all water samples defines a straight line (Fisher and Mullican, 1990).

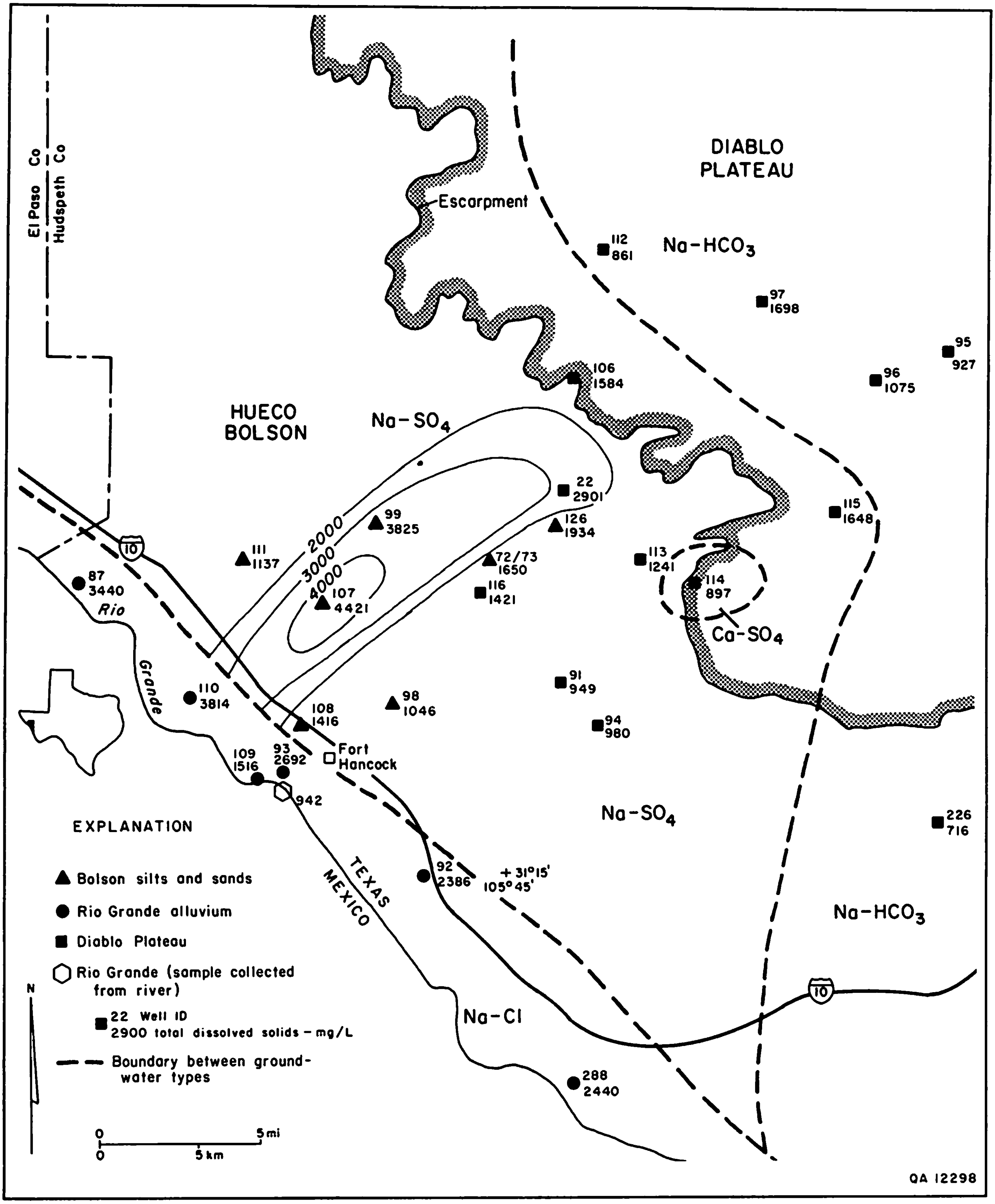


**Figure 1. Map showing representative total dissolved solids (TDS) concentrations and predominant cations and anions in ground-water samples. Contour interval for TDS concentrations is 1,000 mg/L: contours are not extended into the Rio Grande aquifer because the salinity of Rio Grande ground waters varies with rates of irrigation and evaporation.**

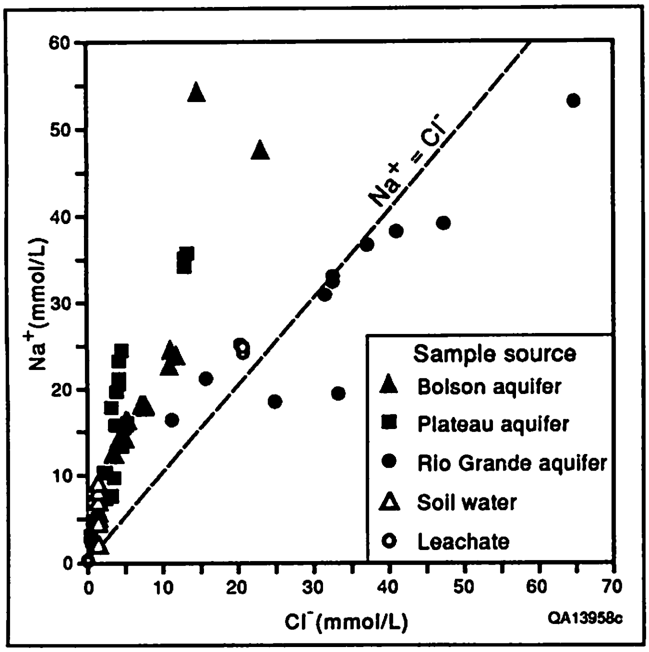


**Figure 2. Plot of $Na^+$ versus $Cl^-$ in ground waters, soil-moisture samples, and soil leachates.**

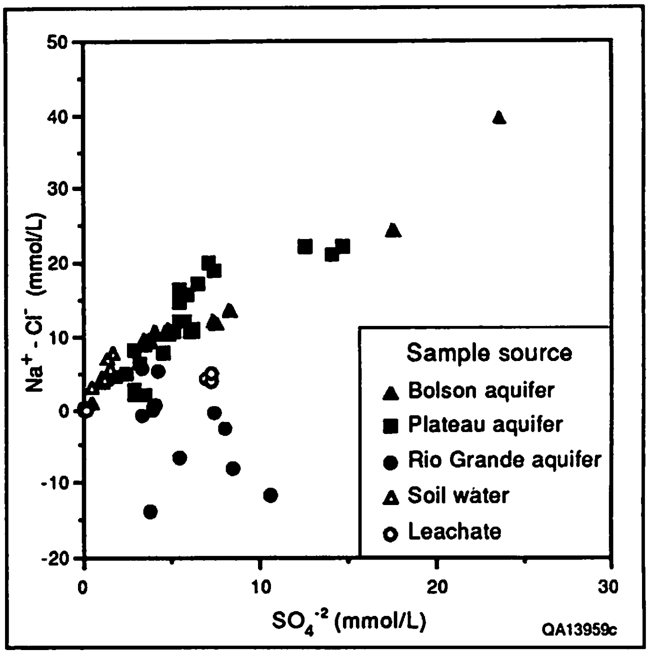


**Figure 3. Plot of $Na^+$-$Cl^-$ versus $SO_4^{-2}$ in ground waters, soil-moisture samples, and soil leachates.**

Figure 4 shows that ground-water samples, water from the unsaturated zone, and soil leachates define two parallel trends in this coordinate system. Figure 4 also provides a convenient basis for interpreting the hydrochemical evolution of the ground waters. The main data trend (fig. 4) includes all soil leachates, all water samples from the unsaturated zone, and all Hueco Bolson and Diablo Plateau ground waters. The least saline waters of this group are leachates from bolson, plateau, and unirrigated alluvial soils. These are calcium-bicarbonate waters and represent the composition of recharge as it first enters and dissolves minerals in the unsaturated section. The next most saline group of samples consists of all waters extracted from the unsaturated zone (soil waters) and several ground waters from the Diablo Plateau aquifer. These are sodium-bicarbonate water and calcium-sulfate waters in which ion exchange reactions have occurred but not extensively enough to produce a sodium-sulfate water type. These samples represent recharge water that has dissolved readily soluble material in the soil zone and has also begun to react either with the sediments of the unsaturated zone or with the aquifer matrix. The main body of this data trend (fig. 4) consists of Hueco Bolson and Diablo Plateau ground waters that have extensively exchanged calcium for sodium, to the point that they have become sodium-sulfate ground waters. These represent the final stage of hydrochemical evolution for Hueco Bolson and Diablo Plateau ground waters.

A second group of samples diverges from the main data (fig. 4); these are all sodium-chloride waters collected from the Rio Grande aquifer or leached from soil irrigated with Rio Grande water. Major compositional differences between the two groups of samples result from the history of the Rio Grande and upper Rio Grande valley. Intensive irrigation in the upper Rio Grande basin began in the early 1880's. Post-1880 irrigation, combined with a series of droughts in the 1940's and 1950's, resulted in serious degradation of river quality (Young, 1981); the Rio Grande in Trans-Pecos Texas is now a brackish, sodium-chloride type water. Heavy irrigation carries large amounts of dissolved salts to cultivated fields; for example, Young (1981) estimated that solutes in irrigation water added approximately 10 tons of salt per acre in the lower El Paso valley and Hudspeth valley in 1955 alone. Solutes in irrigation water become concentrated by evaporation on fields and in the shallow subsurface. The resulting saline water either eventually returns to the shallow alluvial aquifer or to the river, where it is subsequently used to irrigate fields downstream, or the salts are precipitated in the soil zone during times of high evaporation.

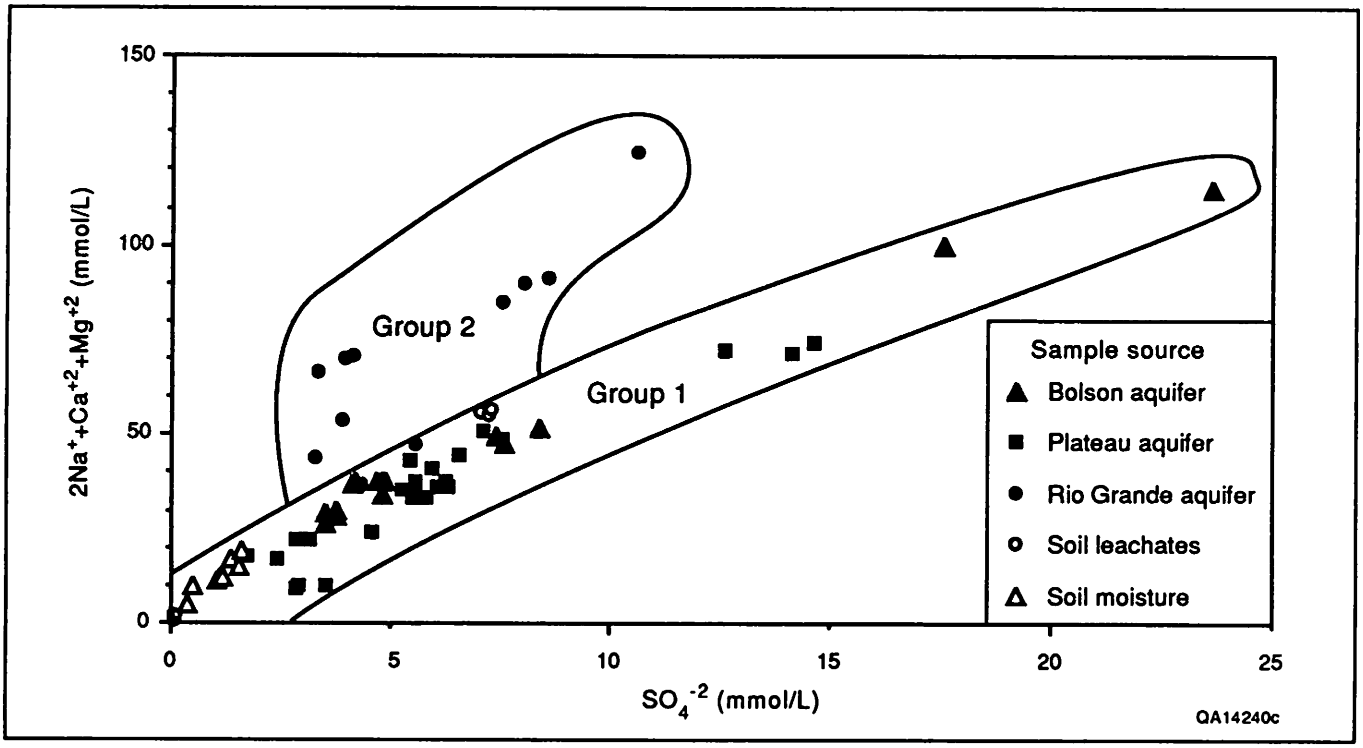


**Figure 4. Plot of $2Na^{+}+Ca^{+2}+Mg^{+2}$ versus $SO_4^{-2}$ in ground waters, soil-moisture samples, and soil leachates. Groups 1 and 2 are discussed in text.**

# Summary and Conclusions

Two distinctly different ground-water systems are present in the southeastern Hueco Bolson and southwestern Diablo Plateau region of Trans-Pecos Texas. Hueco Bolson and Diablo Plateau ground water is generally controlled by simple mineral dissolution and ion exchange. Calcite and dolomite are dissolved to saturation in most ground waters. Gypsum is also dissolved but is too scarce to cause saturation. Dissolution of gypsum raises the calcium concentration in ground-water and soil-water samples, which drives exchange of aqueous calcium for sodium on ion exchange sites. This coupled process results in water compositions dominated by sodium and sulfate in most samples. The hydrochemical evolution of these waters can be traced from recharge to final state by examination of soluble ions that are readily leached from soil, water extracted from the unsaturated zone, and ground waters collected from wells on the plateau, near the escarpment, and on the bolson.

Ground water from the Rio Grande alluvial aquifer is predominantly a sodium-chloride type that reflects the effects of irrigation rather than natural processes in an arid environment. Extensive irrigation upstream and dissolution of salts that precipitate in irrigated fields during times of high evaporation rates largely control dissolved solutes in this aquifer.

The major ionic characteristics of the ground waters are apparently established early in the flow history by processes that occur in the unsaturated zone; these same processes continue in the aquifers. Evidence to support this conclusion includes the observation that the composition of soil leachate and soil-moisture samples falls on the same compositional evolution trends as do those of Diablo Plateau, Hueco Bolson, and Rio Grande ground waters. The fact that ground-water samples from carbonate strata sampled at wells on the Diablo Plateau where soil thickness ranges from zero to only a few centimeters is compositionally similar to soil moisture and ground water from bolson siliciclastic strata is further evidence that geochemical processes in the vadose zone establish the major features of ground-water chemistry.

# Acknowledgments

This investigation could not have been completed without the kind help of many residents of Hudspeth County. Funding for this research was provided by the Texas Low-Level Radioactive Waste Disposal Authority under Interagency Contract Number IAC(88-89)0932. The conclusions of the authors are not necessarily approved or endorsed by the Authority. Bernd Richter and Rainer Senger assisted with sample collection; Bernd Richter provided unpublished chemical analyses of soil-moisture samples. Discussions with Alan Dutton and reviews by Alan Dutton and Tucker Hentz improved draft versions of this report. Figures were drafted under the direction of Richard L. Dillon.

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# Hydrogeology of the Diablo Plateau, Trans-Pecos Texas

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William F. Mullican III,[1] and Ronit Nativ[2]

## Abstract

In Trans-Pecos Texas, ground water flows from recharge points in high mountains or broad monoclinal plateaus to desert basins and regional rivers. The plateaus appear to be more effective recharge zones than discrete mountains because of their large lateral extent and large rainfall catchment areas. Ground water recharged in the Diablo Plateau partially discharges in the adjacent Salt Basin by evaporation, resulting in the precipitation of gypsum deposits.

## Introduction

Hydrogeologic studies of the arid Western United States traditionally have focused on the Basin and Range province, which is characterized by porous-media basin-fill aquifers located between high mountain ranges. The mountains are significantly wetter and cooler than the intervening desert basin floors. Spring snowmelt from the mountains provides an annual source of recharge in this hydrogeologic setting. Many of the arid parts of the world, however, do not benefit from such contrasting topographic and hydrologic regimes. In arid Trans-Pecos Texas, the southern extent of the Basin and Range province, mountains are neither as high nor as numerous as in most of the Basin and Range province. The general climate is warmer, and rarely is there winter snow. The hydrogeology of the arid Trans-Pecos region represents a model different from the Basin and Range setting and may be more representative of many other arid parts of the world.

This study of the hydrogeology of Trans-Pecos Texas and more specifically of the Culberson and Diablo Plateaus was initiated at the request of the Texas Low-Level Radioactive Waste Disposal Authority to evaluate these plateaus for a low-level radioactive waste disposal site for the State of Texas. Through the process of site investigation, regional hydrogeologic characterization of both plateaus was conducted. Greater emphasis is placed on the Diablo Plateau in this paper because of its hydrogeologic relationship to the origin of the gypsum flats in the Salt Basin. Details of the hydrogeology and hydrochemistry of the Diablo Plateau and Culberson Plateau are contained in Kreitler and others (1987).

### Regional Geologic Setting

The Trans-Pecos Texas region lies in the southeastern part of the Basin and Range province (fig. 1) and consists of topographically high plateaus and mountains separated by major normal faults from adjacent topographically low desert basins. Structural development of the province began about 24 m.y. ago during east-northeast-oriented extension (Henry and Price, 1985) and continues to the present. The basins were progressively filled by detritus eroded from the adjacent ranges and from Colorado and New Mexico by the ancestral Rio Grande (**Gustavson**, this volume, p. 27).

The northern and western parts of Trans-Pecos Texas have well-developed northwest-trending basins and ranges, with as much as 5,000 ft (1,500 m) of relief. Quaternary fault scarps occur throughout much of Trans-Pecos Texas (Muehlberger and others, 1978; Henry and Price, 1985) and are abundant in the Salt Basin, a large north-oriented Basin and Range graben aligned along the Culberson/Hudspeth county line.

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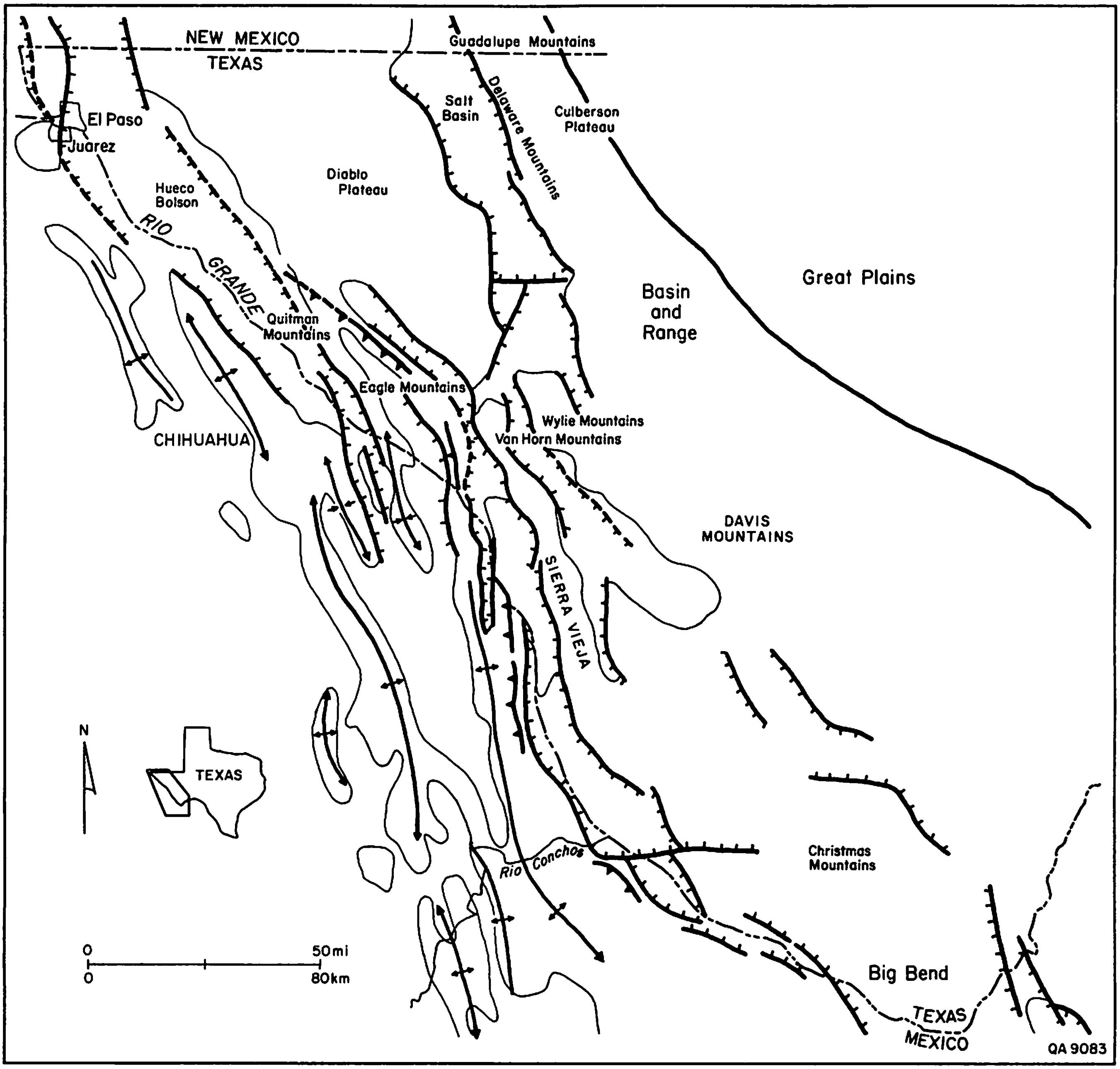


**Figure 1. Tectonic and physiographic map of Trans-Pecos Texas and adjacent northern Mexico (modified from Henry and Price, 1984).**

## Regional Climatic Setting

Trans-Pecos Texas lies within the northern Chihuahuan Desert (King, 1948). The region has a subtropical arid climate (classification of Thornthwaite, 1931) characterized by (1) high mean temperatures with marked fluctuations over broad diurnal and annual ranges (minimum and maximum average annual temperatures are 7° and 27°C [45° and 81°F], respectively) and (2) low mean annual precipitation (10 inches/yr [25 cm/yr]) with widely separated annual extremes. For example, total annual precipitation at Carlsbad, New Mexico, has ranged from 2.7 to 33.8 inches (7 to 86 cm) (Orton, 1964). Precipitation occurs primarily during late summer and early autumn rainfall from thundershowers. These storms occur when moist air from the Gulf of Mexico penetrates northwestward and rises to higher elevations as it approaches the mountains (Carr, 1967). 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 areas is approximately 83 inches (211 cm) (Larkin and Bomar, 1983). High evaporation and low, highly localized rainfall combine to form arid conditions during all or part of most years. For 19 of the 31 yr from 1951 to 1981, Hudspeth, Culberson, and adjacent counties recorded the lowest annual precipitation of any reporting stations in Texas (Bomar, 1983).

# Hydrogeology of Culberson Plateau

Ground-water flow in the Culberson Plateau is to the east toward the Pecos River, down the basinward structural dip of the Permian limestone, sandstone, gypsum, and halite formations as they dip into the Delaware Basin. The potentiometric surface of the regional aquifer is shallow and generally less than 100 ft (30 m) below land surface. Large areas of the Culberson Plateau are dominated by evaporites and carbonates, which control ground-water flow and chemistry. For example, the Permian Castile Formation, composed predominantly of gypsum in the near surface, displays evidence of extensive gypsum solution and local collapse and contains a complex system of karst features and underground solution channels. High tritium concentrations (up to 28 TU) in the ground water indicate active recharge through the thin unsaturated zone and short residence time of ground water in the aquifers; numerous springs discharge from the shallow ground-water table. Dissolution of Permian evaporites has a major effect on ground-water chemistry, resulting in high salinities, with Ca and $SO_4$ being the dominant ions and $\delta^{34}S$ values typical of Permian rocks.

# Hydrogeology of the Diablo Plateau

The Diablo Plateau is underlain by two hydrologic components: a regional water table that not only underlies the Diablo Plateau but also extends underneath part of the Hueco Bolson (**Mullican and Senger**, this volume, p. 37), and a shallower, locally perched, confined to semiconfined aquifer located in the southwestern portion of the study area (fig. 2). Measured depth to the regional water table has been as great as 800 ft (244 m), whereas depths to the shallower aquifer are generally less than 200 ft (61 m). Ground-water quality is fresh to brackish. The chemical composition of the ground waters is generally of a Ca-$SO_4$ type, total dissolved solids ranging from 715 to 3,803 mg/L. The fresh water section may be extremely thick. The U.S. Soil Conservation Service (SCS) drilled a hydrologic test hole to 1,800 ft (549 m) in the Dell City irrigation district on the northeastern side of the plateau and never crossed the base of fresh/brackish water. The limestones are extremely transmissive, as evidenced by the extensive ground-water production from the Dell City region, where approximately 98,500 acre-ft/yr ($10^8$ $m^3$/yr) has been pumped for 30 yr with only 33 ft (10 m) of drawdown. Individual wells when properly located (by lineation analysis approaches, for example) can produce 2,000 to 3,000 gpm (125 to 190 L/sec). Using aerial photography to locate areas of intense fractures, the SCS has successfully located 11 of 12 flood-water injection wells.

Three out of seven pumping tests conducted during recent investigations by the authors in the Diablo Plateau showed no measurable drawdown of the aquifer during extended periods of production (pumping tests with discharges less than 20 gpm (1.2 L/s), typically lasting 48 hr or longer). The Diablo Plateau is made up of zones of extensively fractured bedrock that formed because of a complex tectonic history. These fracture systems may significantly impact ground-water flow and ground-water production because much of this transmissivity is solution and fracture controlled. Video logs run by the SCS in the hydrologic test hole recorded continuous vertical fractures and grapefruit-size dissolution cavities. Fractures were also identified during aquifer pumping tests by the authors (fig. 3). Scalapino (1950) reported that during the development phase of drilling in Dell City, only 44 percent of the wells were successful. On numerous occasions, where wells were drilled as close as 100 ft (30 m), one well would be capable of rates greater than 2,000 gpm (125 L/s), whereas the adjacent well would produce less than 100 gpm (6 L/s), suggesting that highly productive wells intersected fractures, whereas others did not.

Ground-water flow on the Diablo Plateau is predominantly from southwest to northeast. The ground-water divide is close to the southern edge of the Diablo Plateau (fig. 2). Most ground water flows down the structural dip of the monocline, and only a minor amount of ground water flows southwestward into the Hueco Bolson. The hydraulic gradient for the Diablo Plateau aquifer is approximately 5 ft/mi (1 m/km), a relatively low gradient considering that the plateau has approximately 1,300 ft (400 m) of relief in the same region. This low gradient also suggests at least locally high

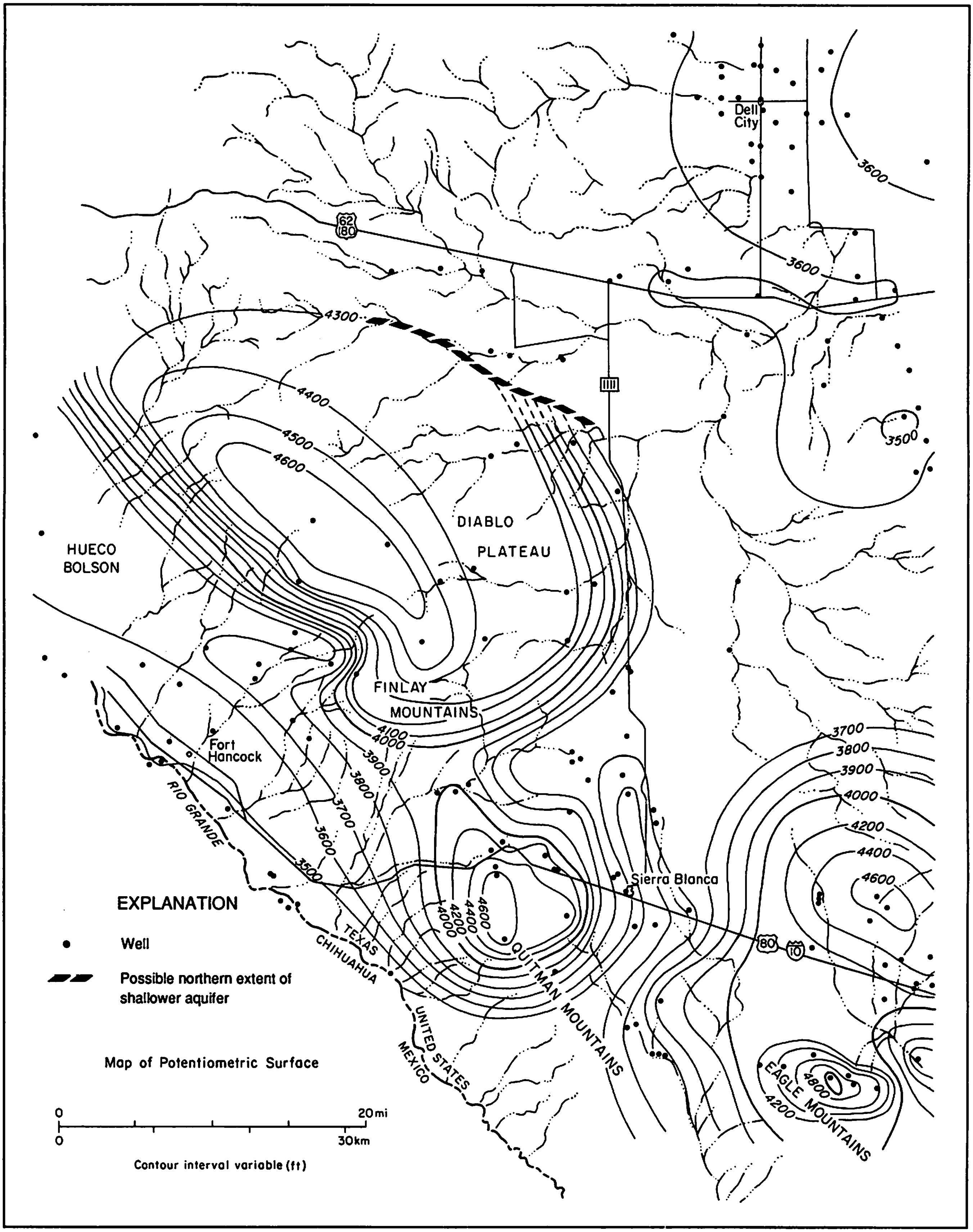


**Figure 2. Potentiometric surface map of the Diablo Plateau, Salt Basin, southeastern region of the Hueco Bolson, and surrounding areas.**

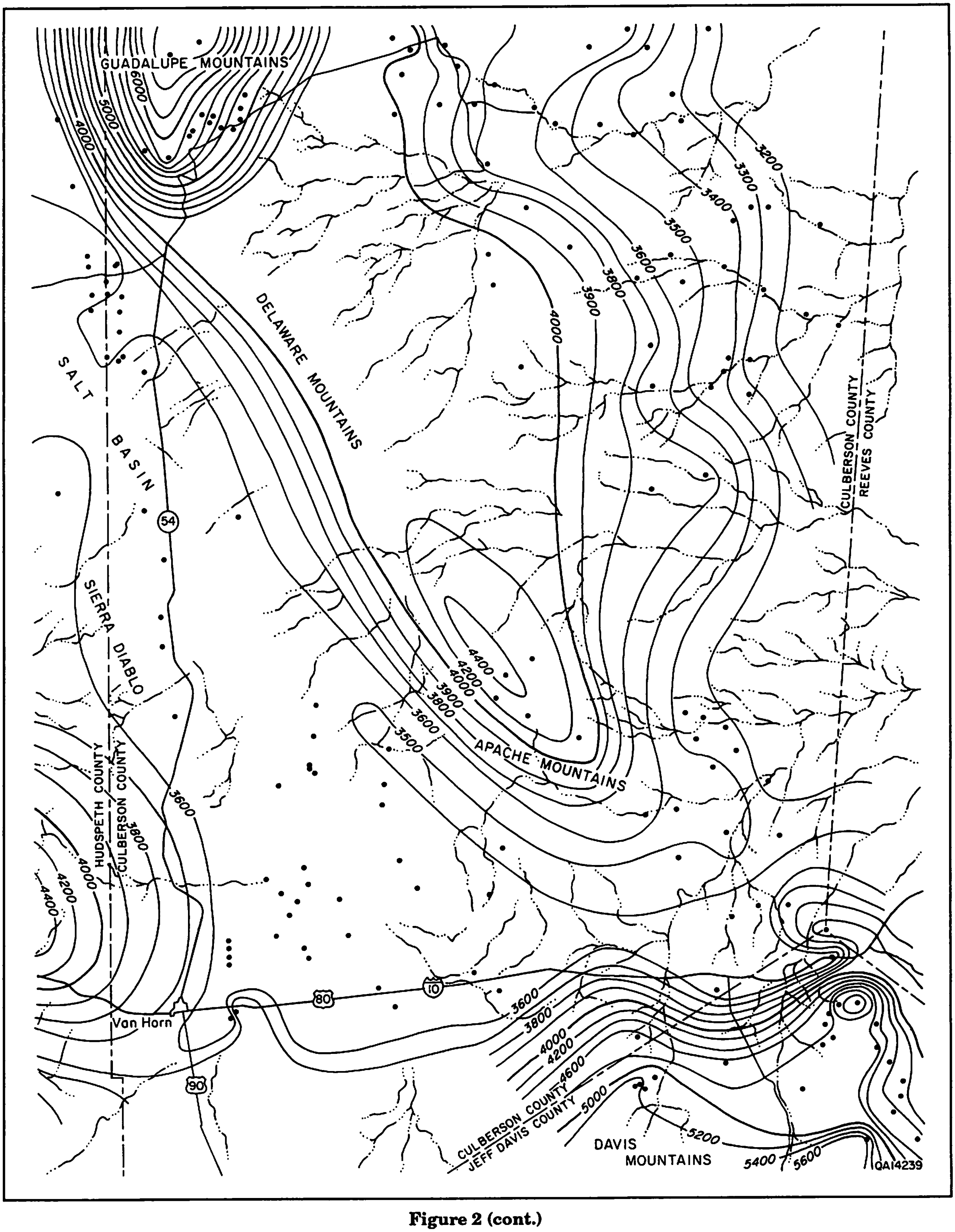


**Figure 2 (cont.)**

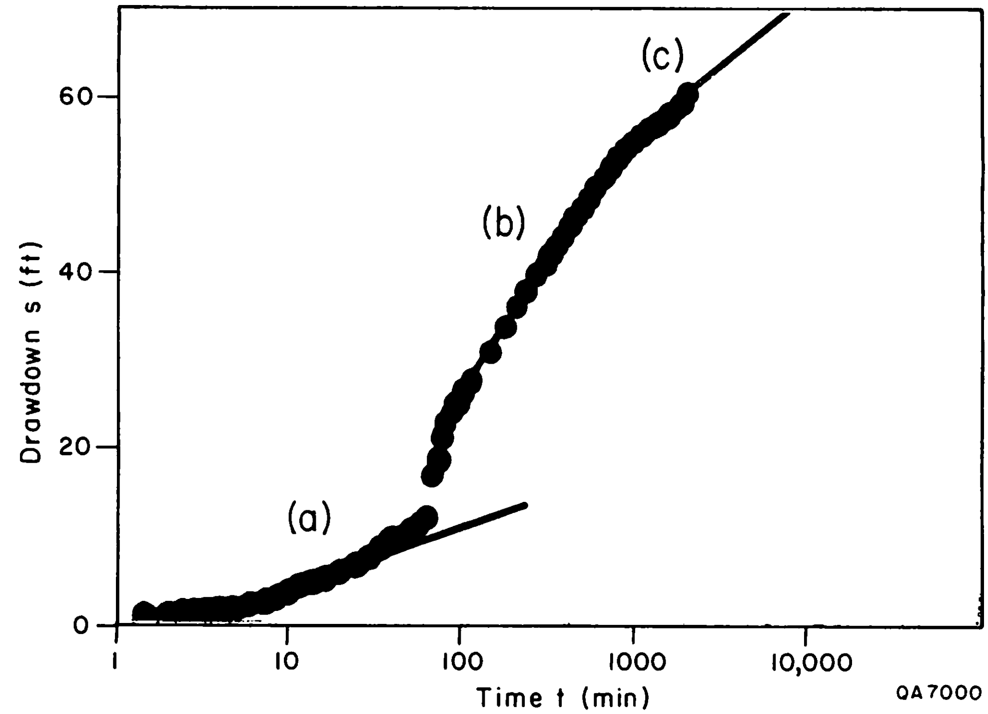


**Figure 3. Aquifer test from LL162 well (Williams Ranch) showing characteristic drawdown curve for fractured media. Location of LL162 shown on figure 4 in west-central part of map area (from Kreitler and others, 1987).**

transmissivities. (Regional ground-water flow is not considered to be affected by extensive ground-water usage in the Dell City region because of the very high transmissivity in the region and in the rest of the aquifer.)

Recharge occurs over the entire study area and is not restricted to the updip part of the potentiometric surface in the areas of higher elevation. The catchment area for recharge therefore is the area of the plateau (approximately 2,900 $mi^2$ [7,500 $km^2$]). Tritium occurs in nearly all wells regardless of their location within the regional water table (fig. 4). Most recharge probably occurs during flooding of the arroyos that traverse the plateau. Fractures, typically concentrated in arroyos, permit surface water to move rapidly through the thick unsaturated section. Analysis of soil-water chloride indicates low concentrations in the arroyo soils (less than 500 mg/L) and significantly higher Cl concentrations (greater than 5,000 mg/L) in the inter-arroyo soils (fig. 5), indicating recharge in the arroyos and minimal recharge between the arroyos.

Ground water discharges by two mechanisms: ground-water evaporation and interbasinal ground-water flow. The potentiometric surface approaches land surface beneath the gypsum flats that occupy the topographic low between the Diablo Plateau and the Guadalupe and Delaware Mountains. Extensive evaporation results in precipitation of gypsum, halite, and carbonates (**Boyd and Kreitler, 1988**). The gypsum sediments at land surface typically are moist all year. The presence of a shallow water table (depth to water approximately 3 ft [1 m]), a thick capillary fringe and upward hydraulic gradients in the unsaturated section, and the Na-Mg-Cl-$SO_4$ chemical composition of brines from both the saturated and unsaturated section beneath the salt flats document ground-water discharge by evaporation and demonstrate that ground-water evaporation is the primary process for formation of the gypsum flats (Chapman, 1984; Boyd and Kreitler, 1986). (These processes are summarized in more detail by Boyd and Kreitler, 1986, and **Chapman and Kreitler**, this volume, p. 59). Occasionally very heavy rains and runoff on the Diablo Plateau will cause flooding of the gypsum flats, although the lake waters quickly percolate into the underlying sediments and the lakes are

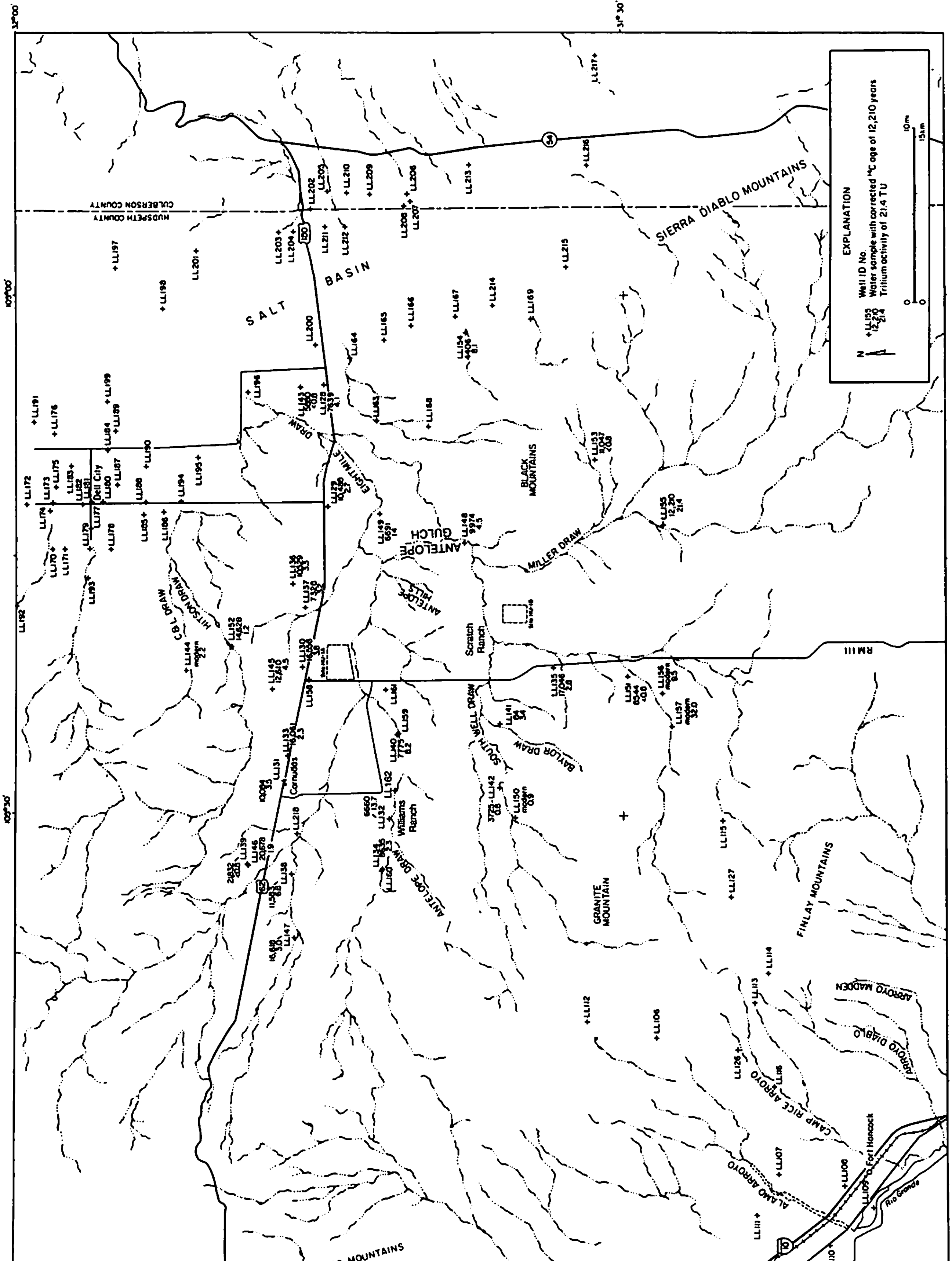


**Figure 4. Areal distribution of tritium in water wells that produce from the Diablo Plateau (from Kreitler and others, 1987).**

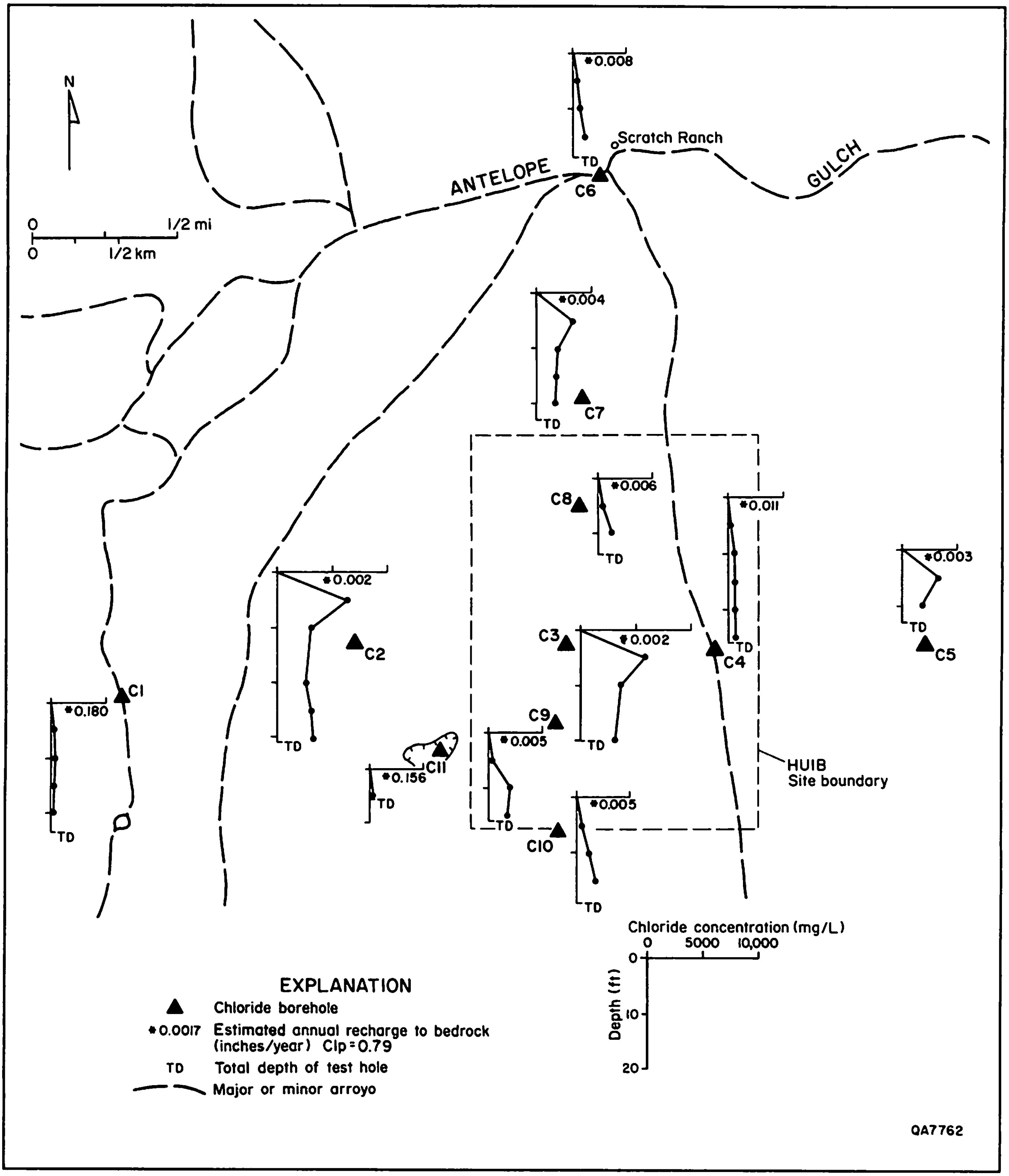


**Figure 5. Cl distribution in arroyo and interarroyo regions. Location of Antelope Gulch and Scratch Ranch is shown in figure 4. Annual recharge estimates using the technique developed by Allison and Hughes (1978) (from Kreitler and others, 1987).**

"dry" within a few weeks. Gypsum deposition may be occurring much more extensively than just in the salt flats. Much of the Quaternary fill within the basin is very gypsiferous, and much of the basin may be infilling or has previously infilled with gypsum as a chemical precipitate. Ground water from the Diablo Plateau is the major source of this water in the Salt Basin. Neither the Guadalupe Mountains nor the Delaware Mountains appear to contribute much ground water to the Salt Basin. Most ground water recharged in the Guadalupe or Delaware Mountains is considered to flow to the east beneath the Culberson Plateau.

The second method of discharge of Diablo Plateau ground water may be regional flow from the Diablo Plateau beneath the gypsum flats to the south through Permian carbonates with ultimate discharge in Balmorhea Springs and the thick Cenozoic alluvium in Pecos County. If the Salt Basin is predominantly a low-permeability ground water barrier of evaporites and continental sediments and thus the final discharge point for Diablo Plateau ground water, then ground water flow within the Diablo Plateau should be forced to land surface as spring flow, but there are no springs along the western edge of the salt flats. Considering the thick fresh-water section in the Diablo Plateau and the relatively thin section of only about 3,280 ft (1,000 m) of evaporites and continental sediments in the Salt Basin (Veldhuis and Keller, 1980), there may be good hydrologic communication between limestones within the Diablo Plateau and limestone beneath continental sediments in the Salt Basin graben. The gypsum flats may indicate wicking by evaporation of shallow ground water rather than the effects of a closed ground-water flow system. The potentiometric surface map for the Salt Basin (fig. 2) suggests that ground water flows to the south and then possibly to the east toward Balmorhea Springs. The Ca-$SO_4$ composition of the spring water supports the existence of such a large regional flow system. (See La Fave and Sharp, 1987, for spring location and more details).

# Conclusions

Ground water flows from recharge points in high mountains or broad monoclinal plateaus to discharge points in topographically low desert basins and regional rivers. Diablo and Culberson Plateaus dominate recharge, appearing to be more effective recharge zones than discrete mountains because of the large areal extent and therefore large rainfall catchment areas. Ground water in the Diablo Plateau and the Culberson Plateau (east of the Delaware Mountains) flows toward the east; flow directions appear to be controlled by the eastward structural and topographic inclination of these large plateaus. There is limited ground-water movement toward the west and southwest down the steep escarpments of these plateaus.

Recharge on the plateaus occurs when runoff from sporadic but torrential rainfalls is focused into the arroyos. Even though the water table beneath the Diablo Plateau may be very deep (up to 800 ft [240 m] below land surface), recharge appears to be rapid and probably occurs along fractures beneath the arroyos. Direct recharge from precipitation on the desert soils is considered minor because evapotranspiration rates are so much higher than annual precipitation rates. High chloride concentrations in interarroyo soils and low chloride concentrations in arroyo soils support this conclusion.

The Salt Basin receives its recharge from the adjacent Diablo Plateau and from adjacent mountainous areas to the east. The basin is dominated by gypsum deposits, which result from ground-water evaporation and evaporite mineral precipitation, and therefore functions as a discharge zone for the Diablo Plateau.

# Acknowledgments

This research was funded by the Texas Low-Level Radioactive Waste Disposal Authority under Interagency Contract Number IAC(86-87)1061. We would like to thank Skeet and Jay Williams of the Williams Ranch for their hospitality, time, information, and use of their water wells during this study. We owe special thanks to George and Ethyl Temple of Salt Flat, Texas, for their always open door after a hot day in the field.

Drilling and well workover services were provided by Byrl Binkley Drilling Contractor of El Paso and James Doss of the Bureau of Economic Geology. Homer Logan of the Soil Conservation Service was a valuable resource through his recharge studies in the Dell City area. Appreciation is extended to Kenneth Moore and Syd Sullenger of the University of Texas Lands Office for their help in locating water wells in the area. Alan Dutton of the Bureau of Economic Geology provided technical assistance during pumping test no. 1.

We appreciate reviews by R. K. Senger, B. C. Richter, W. R. Muehlberger, and J. M. Sharp. Technical editing was by T. F. Hentz, and illustrations were drafted by Annie Kubert Kearns and Maria Saenz. Word processing was by Melissa Snell.

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# The Unsaturated Zone of the Salt Flats of West Texas

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and Charles W. Kreitler[2]

## Introduction

Many playas in the Western United States are inactive hydrologic features, relics of a former pluvial period. In classic Great Basin playas, water tables are often at great depth and the playas themselves are at low elevations, where little, if any, recharge occurs. Although located in the eastern margin of the Great Basin, the playas in the Salt Basin of West Texas differ from the classic model in that they are active hydrologic systems. Because the water table is located within 3 ft (1 m) of the playa surface, the salt flats are important areas of evaporative discharge. A study of the processes in the shallow subsurface, including fluid-pressure distribution and water chemistry, indicated that the most active zone for mineral precipitation is the unsaturated zone (Chapman, 1984).

## Hydraulics

At the study site, located on the east side of the flats north of U.S. 62/180, the water table is at a depth of 40.6 inches (105 cm). Tensiometer and neutron probe measurements identified a thick capillary fringe extending approximately 18 inches (45 cm) above the water table. Seasonal drying of the soil at shallow depths causes the capillary fringe to change from a thicker fringe during the winter to a thinner one in the summer. Hydraulic gradients indicate that above the capillary fringe, water movement is usually upward to the playa surface (fig. 1). Daily temperature cycles cause fluctuations in soil suction: the highest suctions occur during the heat of the day, and the lowest suctions occur just before dawn. A 0.08-inch (2-mm) rainfall on the playa one evening reversed the hydraulic gradient, indicating downward-directed flow during the following morning. Within 24 hr, flow was again toward the playa surface, indicating that the infiltrated rainfall was quickly returned to the atmosphere.

## Hydrochemistry

Water chemistry supports the hydrodynamic finding that the capillary fringe extends from the water table to approximately 24 inches (60 cm) below the surface, as the water in the fringe is similar in chemical composition to the ground water. The effect of evaporative discharge on the chemistry of the unsaturated zone water is revealed by a substantial increase in the total dissolved solids content of the shallower samples (fig. 2) and an enrichment in deuterium and oxygen-18. The $\delta^{18}O$ of Diablo Plateau ground water characteristically ranges from −6 to −10‰ (Kreitler and others, 1987), whereas $\delta^{18}O$ of waters from the unsaturated section of the gypsum playa ranged from -2.3 to -1.1‰. Chloride contents indicate that almost 50 percent of the original water volume has been removed by evaporation at 12 inches (30 cm) below the playa surface (sample E, fig. 3). Another 10 percent is lost between 12 and 8 inches (30 and 20 cm) (sample F, fig. 3). Using the computer code AQ/SALT (Bassett and Griffin, 1981), which was designed for high-ionic strength waters, we calculated that calcite, dolomite, and gypsum were at or above saturation. Although approached, halite saturation was not achieved. Ion ratios show a depletion in calcium and sulfate (relative to chloride) in the shallower samples, indicating that gypsum precipitation occurs in the unsaturated section. Dolomitization may promote gypsum precipitation by releasing additional calcium into solution.

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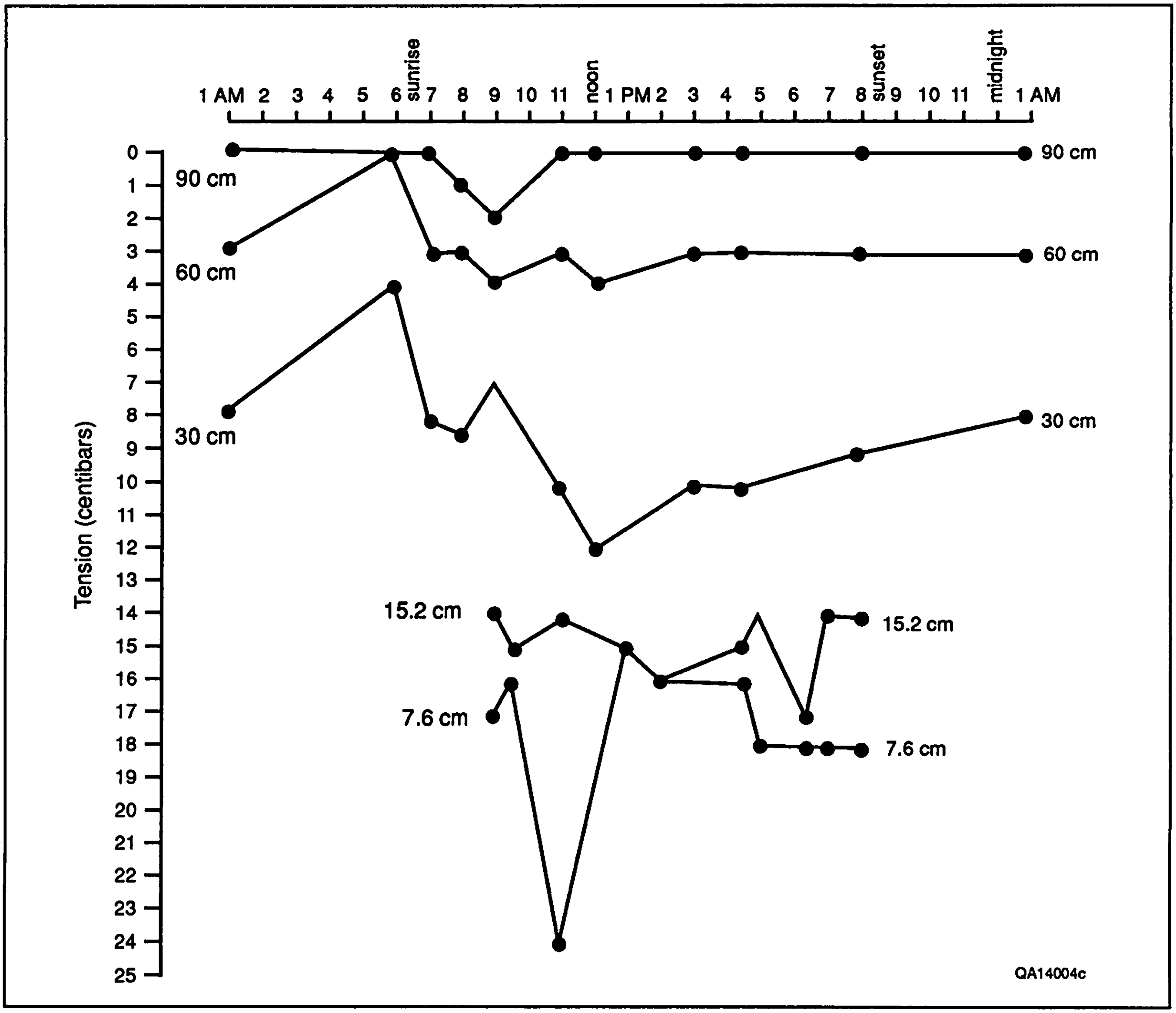


**Figure 1. Diurnal suction measurements made with tensiometer at 12, 24, and 36 inches (30, 60, and 90 cm) deep during June 1982 and by tensiometers at 3 and 6 inches (7.6 and 15.2 cm) deep during May 1983.**

# Lithology

The shallow sediments in the study area consist almost entirely of gypsum and dolomite and minor amounts of quartz (calcite has been found by workers in other parts of the flats). The sediments are massive below the water table and laminated throughout the unsaturated zone (fig. 4). The laminations may be the result of periodic flooding of the playas with water and detrital material, or adhesion of eolian material onto the moist surface of the flat (perhaps aided by algal mats). Evidence of the gypsum precipitation predicted by the water chemistry is found in the form of enterolithic bands, nodules, and discontinuous lenses of gypsum. The dolomite is a poorly ordered mud, probably formed by the alteration of calcium carbonate in the high-Mg pore water.

# Conclusions

The notion of relict systems containing an unaltered record of past lacustrine environments is inapplicable to the salt flats of western Texas. The unsaturated zone in the salt flats is an active system of evaporative discharge, brine evolution, and mineral precipitation.

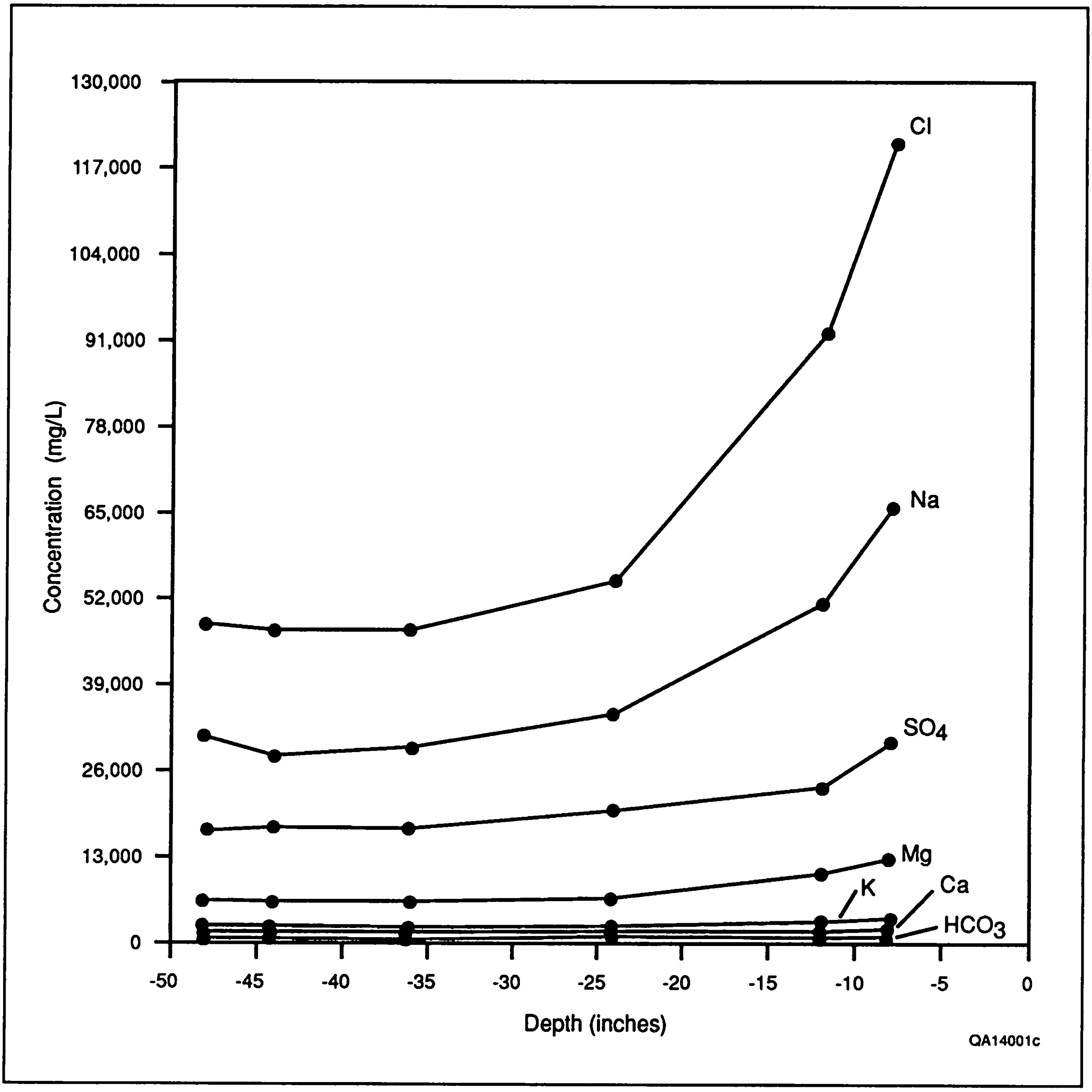


**Figure 2. Concentration of major ions plotted against depth. Samples collected in May 1983.**

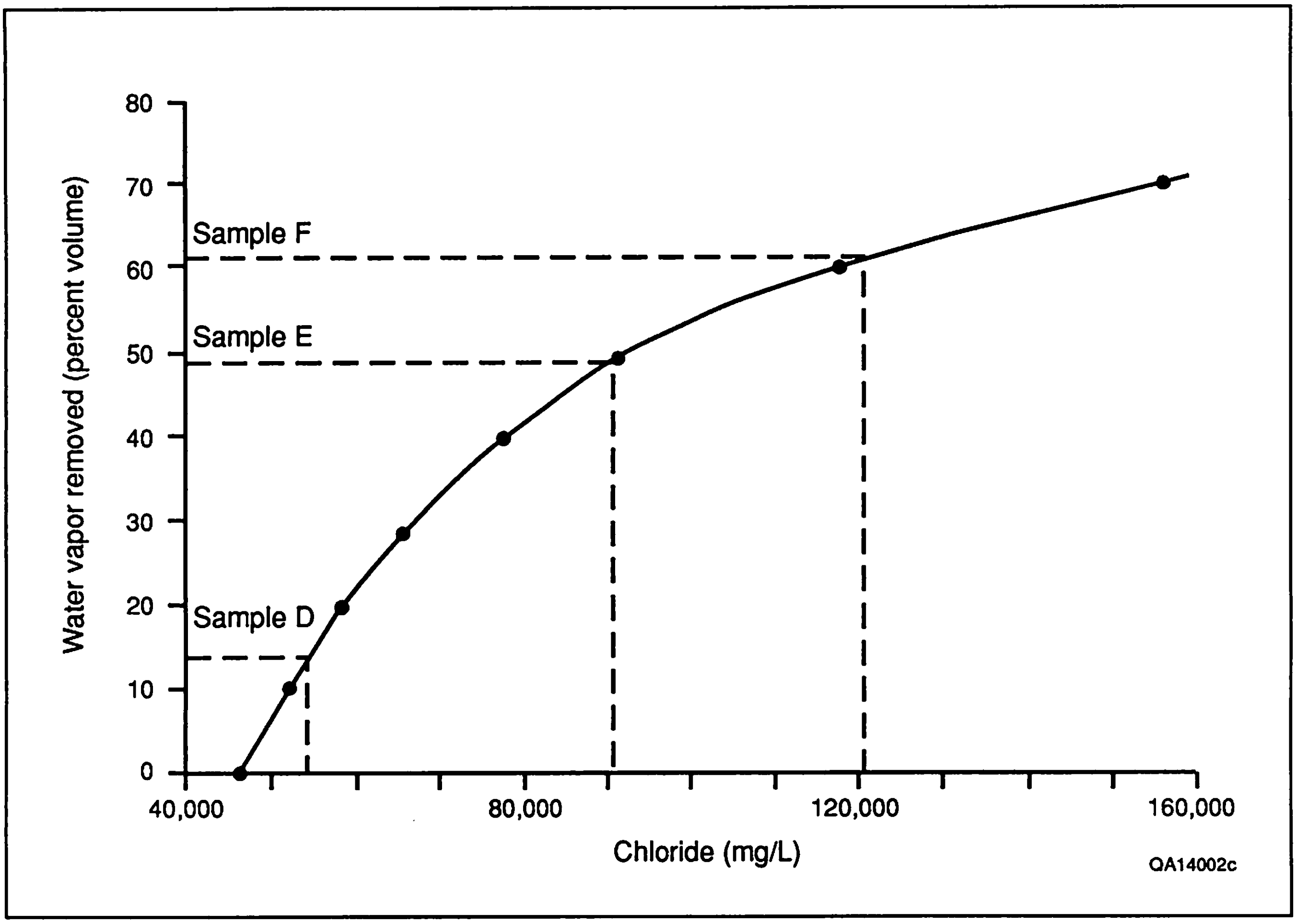


**Figure 3. Simulated evaporation path of chloride relative to water volume lost, computed by the program PHREEQE; measured chloride contents from salt-flat samples and inferred water lost are also plotted.**

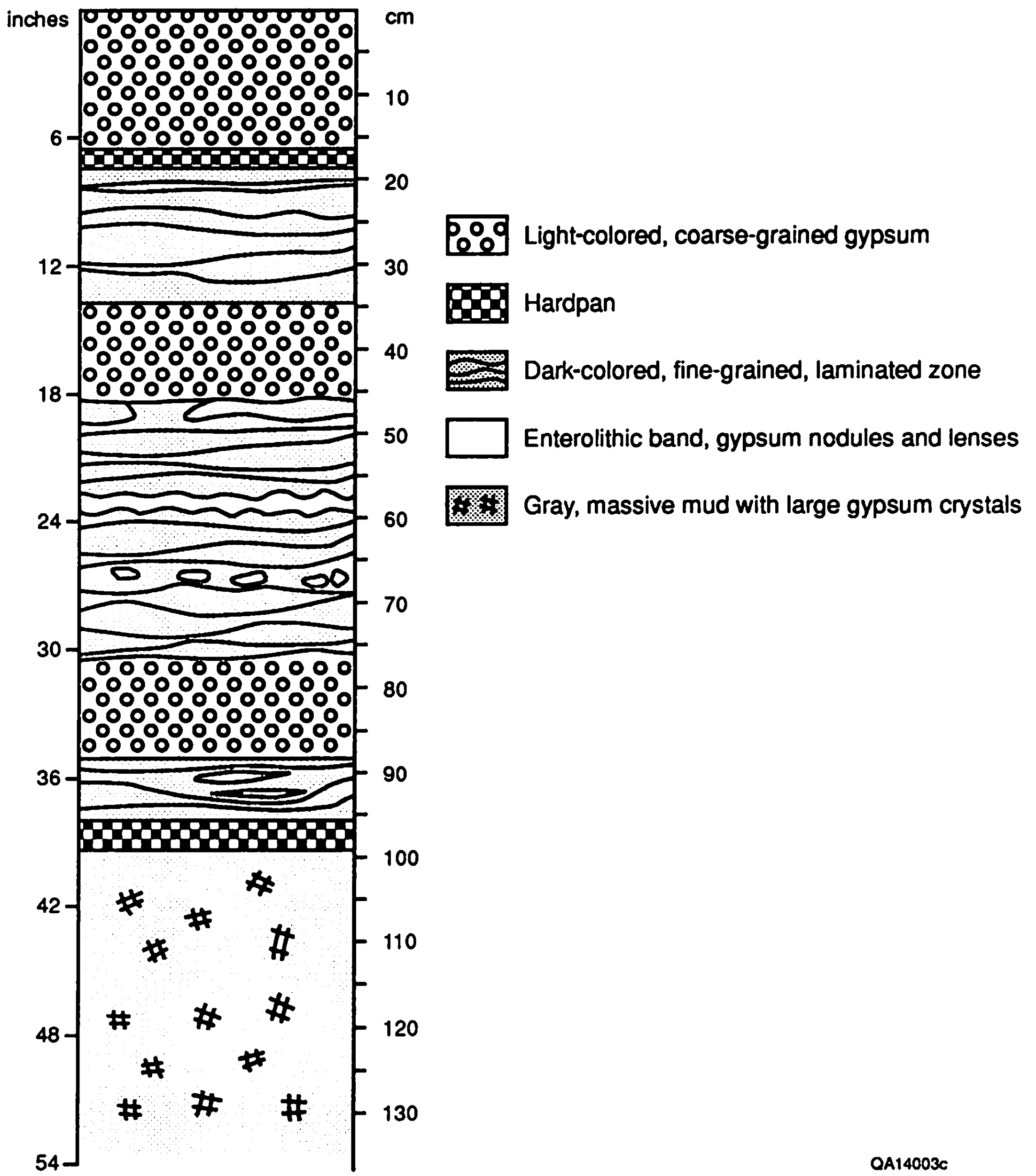


**Figure 4. Lithologic description of 54 inches (137 cm) of core taken from test site on salt flat.**

## Acknowledgments

We thank W. R. Muehlberger, B. C. Richter, and R. K. Senger for reviews of our manuscript. Illustrations were drafted by the Cartography staff of the Bureau of Economic Geology.

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# Selected Reprints on the Hydrogeology of the Trans-Pecos Area

EL PASO GEOLOGICAL SOCIETY GUIDEBOOK 18
p. 170-183

# HYDROGEOLOGY OF A GYPSUM PLAYA, NORTHERN SALT BASIN, TEXAS[1]

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## ABSTRACT

The northern Salt Basin in West Texas and New Mexico is a closed hydrologic system in which discharge of ground-water flow occurs in a series of playas, or salt flats. Ground water originating in peripheral consolidated rocks and alluvial fans flows toward the center of the basin and discharges by evaporation from the salt flats. Progressive increases in salinity are characteristic of the waters moving down gradient and are primarily attributed to evaporative concentration. Steady increases in sodium ($Na^+$) and chloride ($Cl^-$), and, to a lesser degree, magnesium ($Mg^{+2}$) and potassium ($K^+$) characterize increases in total dissolved solids (TDS). Intense evaporation above the water table in the salt flats concentrates the composition of brine to TDS values that range from 50,000 mg/L to greater than 300,000 mg/L. Precipitation of minerals from solution, primarily alkaline earth carbonates and gypsum, results in depletion of calcium ($Ca^{+2}$) and bicarbonate ($HCO_3^-$) and relative enrichment of sulfate ($SO_4^{-2}$) in the ground water. Inflowing ground water of a Ca-Mg-$SO_4$ composition undergoes evaporation and evolves to a final composition of a Na-Mg-$SO_4$-Cl brine in the salt flats. Gypsum currently precipitates in the salt flats at a depositional rate that ranges from 0.02 to 0.09 cm/yr. These deposits do not likely represent relict lakebed sediments.

Evaporative concentration and mineral precipitation are the most important processes in the basin and flats that affect the composition of the shallow subsurface fluids. However, the chemistry of the brines in the salt flats is occasionally modified by surface water that ponds on the playa surface and percolates downward.

## INTRODUCTION

The Salt Basin is one of the most prominent intermontane valleys of the Trans-Pecos region (Fig. 1). Located about 145 km east of El Paso, Texas, the Salt Basin trends northwest-southeast for more than 240 km, from the Sacramento River basin in New Mexico to Presidio County in Texas.[2]

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The purpose of this study was to determine whether the salt flats are currently forming by processes of ground-water

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[1]Excerpted from Report of Investigations No. 158 with permission of authors and Director, Bureau of Economic Geology.

[2]Major omissions from the original text are indicated by . . . . Figures have been renumbered to provide continuity.

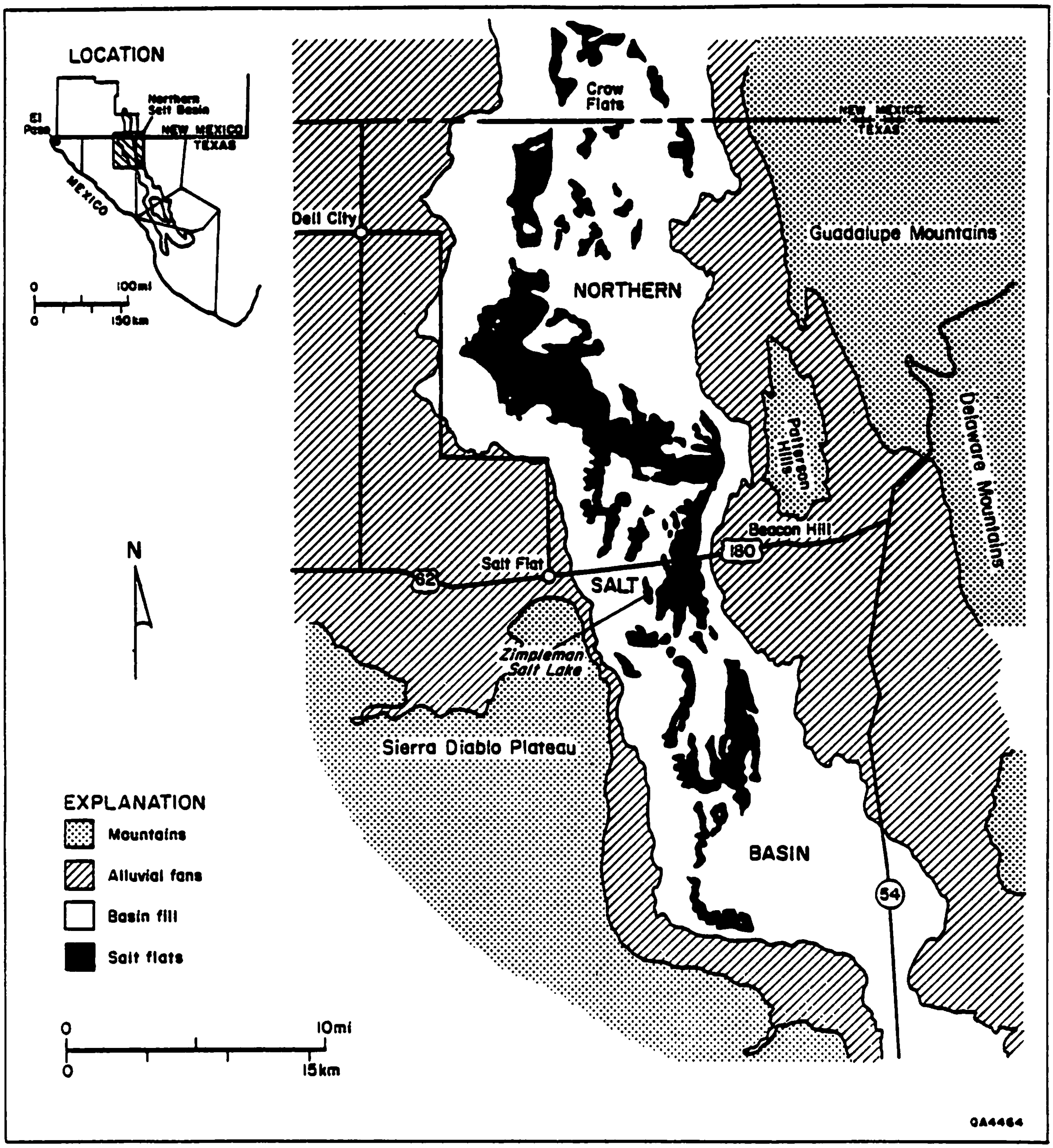


**Figure 1.** Location of the northern Salt Basin (modified from Barnes, 1975).

evaporation and mineral precipitation. It was not to argue the presence or absence of a relict lake. Gypsum precipitation may or may not be occurring on older lacustrine sediments. Our objective was accomplished by characterizing the (1) geology of the source rocks, (2) geomorphology of the playa surfaces, (3) shallow subsurface stratigraphy and mineralogy of the salt flats, (4) ground-water movement, and (5) geochemistry of the waters in the basin, and by relating these phenomena to the origin and evolution of the brines and to the processes and rates of gypsum precipitation.

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### Geomorphic Setting

The mountains surrounding the northern Salt Basin rise to more than 2,000 m high. Guadalupe Peak, the highest point in Texas at 2,667 m, overlooks the eastern boundary of the northern Salt Basin. A series of coalescing alluvial fans, interrupted only at wide intervals by rock ridges, extends as a belt along the west side of the Guadalupe and Delaware mountains, between the base of their west-facing escarpment and the floor of the Salt Basin. The gently sloping alluvial fans end abruptly at the nearly horizontal basin floor.

The Salt Basin itself is a large, flat-lying depression, best described in early observations by Richardson (1904):

> The rise is very gradual both to the north and to the south. Northward from the center of the basin to the State line (Texas-New Mexico), the slope of the surface is only about two feet in a mile. This inclination is unappreciable to the eye, and the basin appears to be flat in its longer extent and to rise only to the adjacent highlands to the northeast and southwest plain. The surface of the basin is not a level plain, but is characterized by numerous hillocks and local depressions. The former often are composed of wind-blown material, and the latter are locally called "lakes."

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## HYDROLOGY

The climate of the northern Salt Basin is typical of the semi-arid Southwest. The winters are generally mild, and the summers are hot and dry; temperatures during the year range from -12°C to 46°C. The average annual precipitation is about 21 cm in the basin and more than 50 cm in the Guadalupe Mountains east of the basin. Average relative humidity, as recorded for El Paso, is 35 percent at 11 a.m. and 27 percent at 5 p.m. (National Oceanic and Atmospheric Administration, 1984). Early spring is characterized by gusty winds, sometimes reaching 90 km/hr (Goetz, 1977).

The northern Salt Basin is a topographically and hydrologically closed basin. The basin receives drainage from a large area bounded by the Sacramento Mountains on the north, the Guadalupe and Delaware mountains on the east, and the Sierra Diablo Plateau on the south and west.

### Recharge

Ground water moving through the alluvial fans and underlying aquifers contributes the major volume of recharge to the system. Runoff from the mountains, where precipitation is greatest, is rapid and results in floodwaters that flow down the canyons and spread out over the alluvial fans flanking the salt flats. Water moves downward into the coarse-grained, permeable alluvium to serve as a primary source of recharge to the northern Salt Basin ground-water system. If precipitation is intense or prolonged, minor amounts of overland flow move rapidly through the draws directly onto the basin and flats.

Recharge to the salt flat area also consists of ground water moving through the underlying aquifers. The Sacramento River drains into the northwestern part of the closed Salt Basin. Infiltration of its water recharges the Bone Spring Limestone, the major aquifer underlying the Dell City irrigation area (Bjorklund, 1957). On the east side of the basin, recharge occurs in the outcrop areas of the Capitan and Goat Seep limestones and the sandstones of the Delaware Mountain Group in the Guadalupe Mountains, Patterson Hills, and the Delaware Mountains.

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Another means of recharge is infiltration of precipitation that falls directly onto the basin and flats. Data on annual precipitation in the northern Salt Basin (Salt Flat station) from 1945 to 1982 (Texas Water Development Board, 1973; National Oceanic and Atmospheric Administration, 1974-1982) indicate that the average rainfall in the basin is about 21 cm/yr. Estimates of discharge by evaporation (discussed in the "Discharge" section of this report) indicate that the volume of precipitation falling directly onto the basin is probably entirely evaporated and, therefore, an insignificant source of recharge.

**Movement**

Movement of ground water begins in the recharge areas, including alluvial fans and underlying aquifers around the margins of the Salt Basin and the ephemeral stream channels associated with the Guadalupe and Delaware mountains, and continues through the basin fill to areas of discharge in the salt flats (the lowest parts of the basin). Water returns to the surface mainly by percolating upward, drawn by evaporation through the capillary fringe in the flats. Circulation also occurs when rain falls directly onto the basin and infiltrates the shallow sediments. This water is in the system only a short time before evaporating. Because of the low permeability and stratification of the sediments and the strong evaporative force, very little precipitation is ever likely to reach the water table.

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**Discharge**

In the northern Salt Basin, ground water discharges mainly by evaporation, especially from the salt flats. The water table stands at a depth of 1 to 3 m below the ground surface in the flats. Water evaporates where the surface is bare or is evapotranspired where scattered growths of phreatophytes occur. Water collects in the lakes after unusually heavy rains from direct precipitation and from runoff, but the water table lies below the land surface most of the time. Davis and Leggat (1965) suggested that some discharge may also take place by subsurface flow through southeasterly dipping Bone Spring and Victorio Peak limestones, but the extent of loss by this means has not been determined.

An estimate of discharge can be made by assuming that evaporation is the primary mechanism. Bell and Sechrist (1970), in their studies of playa lakes in New Mexico and the High Plains of Texas, and Scalapino (1950), in his studies of the Dell City area of the West Texas Salt Basin, concluded that the evaporation rate from a free water surface in those areas is about 200 cm/yr.

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Bjorklund (1957) studied evaporation rates in the Crow Flats of the northern Salt Basin and concluded that the natural discharge from the salt flats by evaporation before the development of irrigation wells was probably equal to the recharge to the Bone Spring aquifer, which he calculated to be less than 100,000 acre-ft/yr ($1.23 \times 10^8$ $m^3$/yr). He estimated that the flats include an area of 37,000 acres (150 $km^2$) in New Mexico and Texas and that the evaporation rate is less than about 82 cm/yr.

The average rate of evaporation at the salt flats in the northern Salt Basin is an estimated 18 to 82 cm/yr (18 cm/yr represents Allmendinger and Titus's field measurement, and 82 cm/yr represents Bjorklund's estimates). Although direct precipitation onto the flats averages about 21 cm/yr, it is considered insignificant in the discharge calculations of the salt flat because of its rapid evaporation. These evaporation figures are estimates because no actual measurements were made, but they do represent order-of-magnitude approximations of discharge by evaporation.

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## HYDROCHEMISTRY

### Data Base and Analytical Methods

Table 3 [not shown here] includes data on the chemical composition of ground water from the alluvial fans and underlying aquifers collected from records on selected irrigation wells, windmills, domestic wells, and stock wells. These records were compiled by White and others (1980) for the eastern half of the northern Salt Basin (including the Beacon Hill irrigation district), by Scalapino (1950) and Dillard and Muse (1964) for the western half of the northern Salt Basin (including the Dell City irrigation district), and by Bjorklund (1957) for the New Mexico and Crow Flats region of the basin. For wells in the Dell City area, only the analyses of water sampled before 1950 were used in this report; subsequent analyses reflect changes in the natural water composition due to intense irrigation in that area. Six windmills and private water wells were sampled and analyzed by Boyd for this study.

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### Results

The analysis results show that the water composition in the shallow subsurface of the playas is that of a brine. The TDS values range in concentration from 50,000 mg/L to greater than 300,000 mg/L, with $Na^+$ and $Cl^-$ making up the bulk of the constituents. Relatively high $Mg^{+2}$ and $K^+$ concentrations are also characteristic of the playa brines. After $Cl^-$, $SO_4^{-2}$ is the most abundant anion, ranging in concentration from 6,000 mg/L to more than 89,000 mg/L. The $Ca^{+2}$ and $HCO_3^-$ concentrations in the brines are relatively low, being approximately the same as or lower than those of the water from the outer basin fill and from the highlands.

Composition trends for cations and anions in the ground water as it flows through the alluvial fans and basin fill are illustrated by a series of graphs representing suites of samples along hypothetical flow paths into the flats.

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Sections A-A' [not included], B-B' (Fig. 2), and C-C' [not included] illustrate the increase in total salts basinward and show a notable increase in $Na^+$, $Cl^-$, $Mg^{+2}$, and $K^+$. The $Ca^{+2}$, $HCO_3^-$, and $SO_4^{-2}$ ions remain relatively constant as the water moves toward the flats. Within the flats, the water is highly concentrated in dissolved solids, primarily $Na^+$ and $Cl^-$. As shown by section D-D' (Fig. 3) no apparent compositional trends relative to distance of travel exist within the discharge zone.

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## DISCUSSION

### Brine Chemistry Evolution

The chemical composition of the saline water and brines of the northern Salt

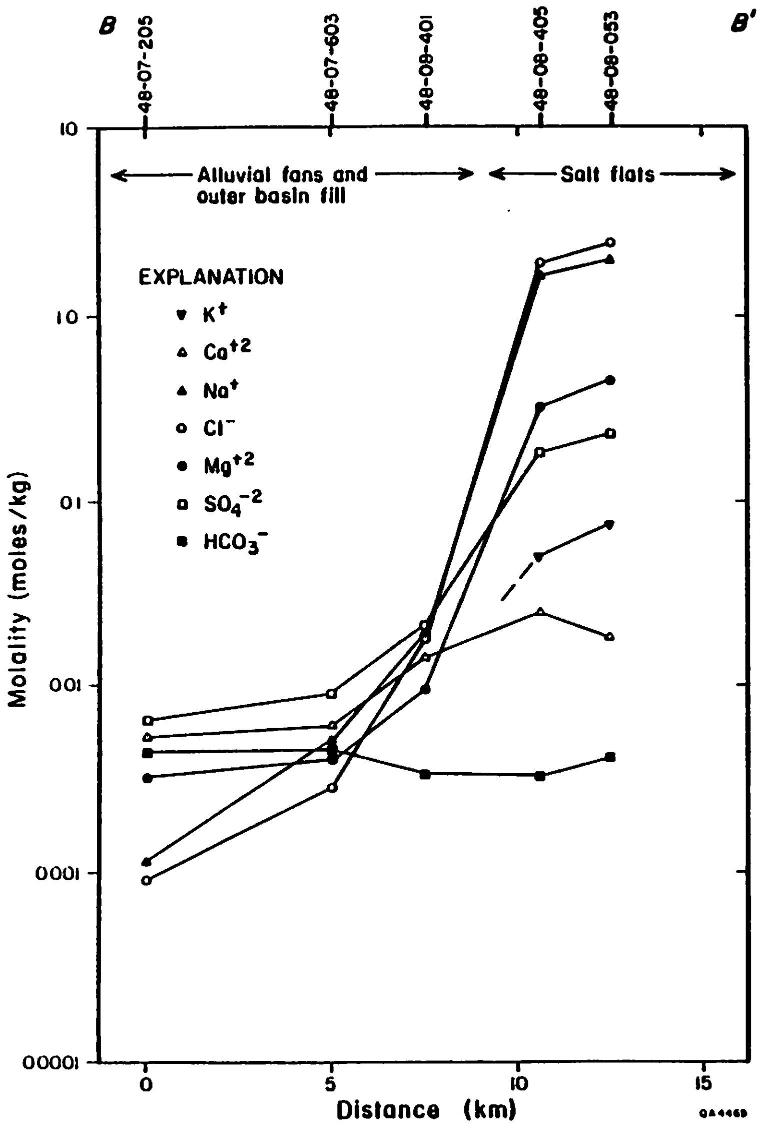


Figure 2. Section B-B'. Ground-water chemical composition from Dell City to the salt flats. See figure 5 for section location.

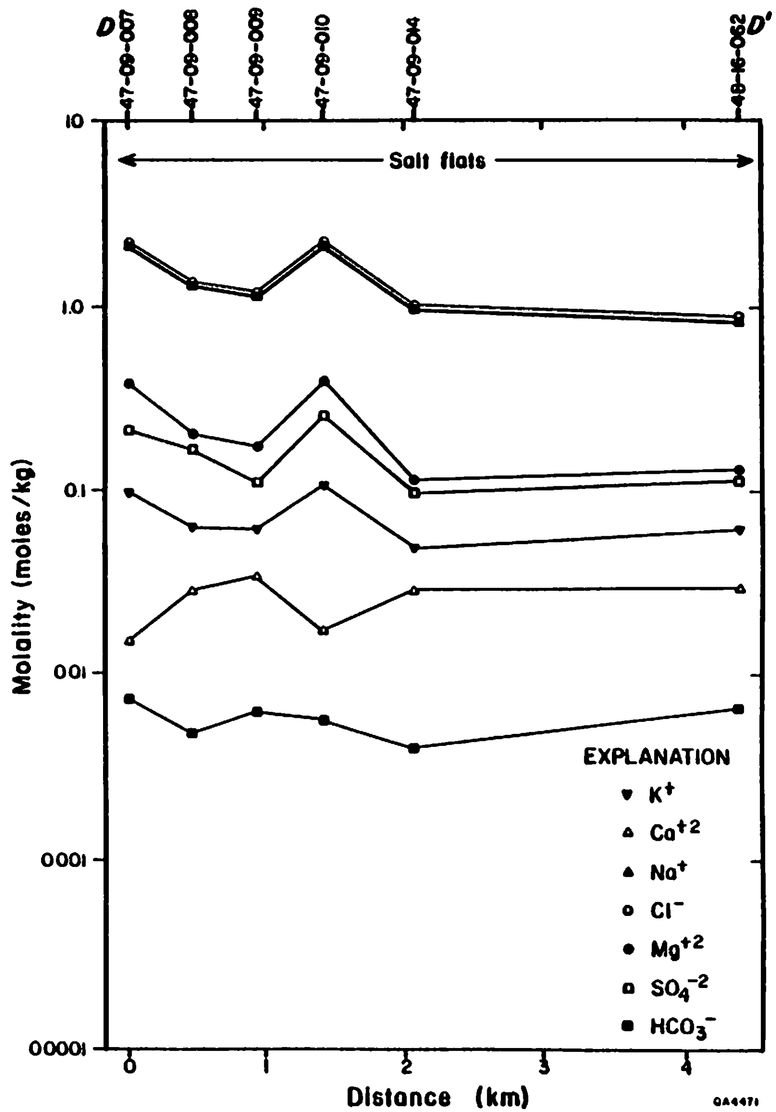


Figure 3. Section D-D'. Ground-water chemical composition within the salt flats. See figure 5 for section location.

Basin can be explained largely by evaporation of dilute inflow waters. The ground waters have an initial Ca-Mg-$SO_4$ composition in the gypsiferous Permian limestones of the recharge area. These waters flow through the alluvial fans and finally discharge by evaporation in the salt flats. Figure 4 exhibits a transition from Ca-Mg-$SO_4$ waters in the limestones to Ca-Na-Mg-$SO_4$-Cl waters in the alluvial fans and finally to a Na-Mg-$SO_4$-Cl brine in the salt flats.

A good indicator of evaporative concentration is $Cl^-$; because of its very high solubility, $Cl^-$ plays little or no role in processes affecting dissolved materials below NaCl saturation. In the salt flats of the northern Salt Basin, increases in $Cl^-$ are directly correlated with increases in TDS. Evaporation occurs within the fans and outer basin fill as well as on the flats, as indicated by the increase in chlorinity. The associated cations, $Na^+$ and $K^+$, are also reliable tracers during the evaporation process as long as there is little or no ion exchange or preferential sorption of these ions on playa sediments. Uniform ratios of $Na^+/K^+$ versus TDS and $Na^+/Cl^-$ versus TDS through the entire range from dilute inflow water to concentrated water in the flats and the parallel increases in concentration of $Cl^-$, $Na^+$, and $K^+$ in the cross-sectional graphs (Fig. 2) illustrate the conservative nature of these ions and the apparent absence of ion exchange processes involving these constituents.

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The theoretical brine evolution and mineral precipitation sequence (Hardie and Eugster, 1970) is shown in Figure 6.

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Continual loss of $Ca^{+2}$ by precipitation of carbonate minerals (in the alluvial fans) and sulfate minerals (in the salt flats) as water is evaporated and moves basinward results in a Mg-enriched and Ca-depleted solution. In hypersaline environments, ground water with $mMg^{+2}/mCa^{+2}$ ratios less than unity may be concentrated by evaporation to dense brines with very large $mMg^{+2}/mCa^{+2}$ ratios (Kinsman, 1966; Friedman and Sanders, 1967; Folk and Land, 1975). In the northern Salt Basin, the inflow waters have $mMg^{+2}/mCa^{+2}$ less than unity (circles in Fig. 5). The ratio increases in some of the playa brines, however, to as much as 51:1. The striking rise in the $mMg^{+2}/mCa^{+2}$ ratio corresponds to depletion of $Ca^{+2}$ from the brines by the precipitation of calcite, aragonite and gypsum, and the retention in solution of the $Mg^{+2}$. Dolomite, which is abundant in the salt flat sediments, may be forming because of this high Mg/Ca ratio.

Because of the initial relative concentrations of $Ca^{+2}$, $HCO_3^-$, and $SO_4^-$, the initial inflow waters in the Salt Basin should theoretically evolve to a brine containing $Cl^-$ and $SO_4^-$ as the major anions and $Na^+$ and $Mg^{+2}$ as the major cations, following the highlighted path indicated in Figure 6. The concentration-versus-distance graphs (Figs. 2 and 3) and the trilinear diagrams (Fig. 4), which show the evolution of the Salt Basin waters, verify that the composition of the final water in the Salt Basin flats is indeed a Na-Mg-$SO_4$-Cl brine. The trends in the trilinear diagrams identify the geochemical processes that occur with increases in TDS (Fig. 4). The trend in the anion diagrams is away from the $SO_4^-$ corner of the diagram, where the dots represent waters in the peripheral limestones and gypsums (note the low amounts of $HCO_3^-$ in the initial inflow waters), toward the $Cl^-$ corner, where the crosses representing the salt flat brines lie. Waters in the alluvial fans (squares) lie between the waters in the peripheral rocks (dots) and the brines in the salt flats (crosses). The trend away from the $SO_4^-$ corner toward the $Cl^-$ corner represents gypsum precipitation. In the cation diagrams, the composition of the water trends away from the $Ca^{+2}$-$Mg^{+2}$ part of the diagram, where the bedrock (dots) representing the inflow waters lies, through the alluvial-fan waters (squares), toward the $Na^+ + K^+$ corner (crosses), indicating calcite and aragonite precipitation, gyp-

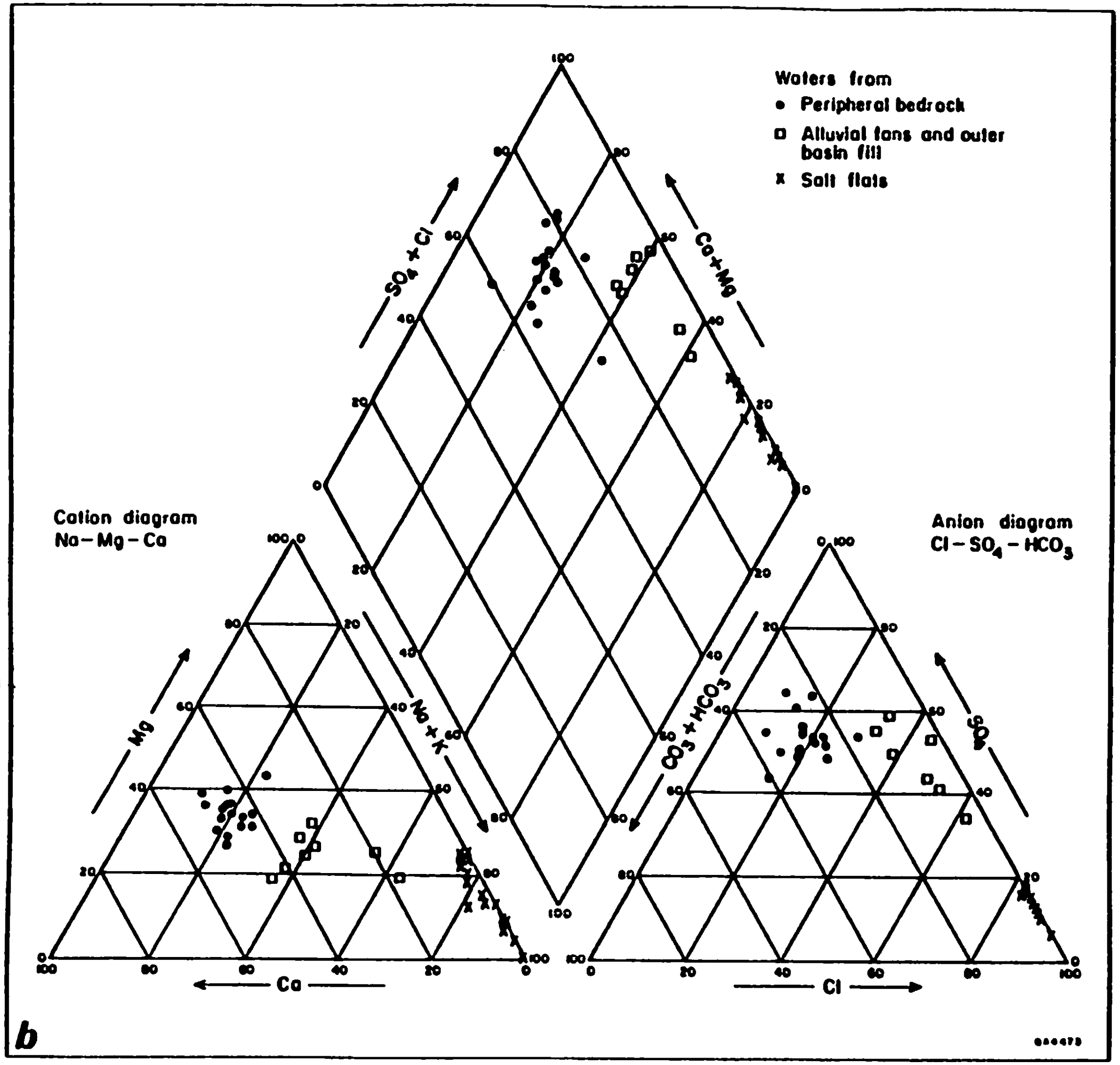


**Figure 4.** Piper diagrams of chemical composition of ground waters from the Salt Basin. (a) West side of basin. (b) East side of basin. In the cation diagram ($Na^{+}$-$Mg^{+2}$-$Ca^{+2}$) the water composition trends from the $Ca^{+2}$-$Mg^{+2}$ section of the diagram, where the inflow waters from the peripheral limestones (dots) lie, through the alluvial fans (squares) toward the $Na^{+}$ corner, where the salt flat waters (crosses) are located, indicating calcite and gypsum precipitation. Much of the $Mg^{+2}$ in the flats remains in solution, as shown by the scatter of crosses along the $Mg^{+2}$-$Na^{+}$ side. The trend in the anion diagram ($Cl^{-}$-$SO_4^{-2}$-$HCO_3^{-}$) is from the $SO_4^{-2}$ area of the diagram (low amounts of $HCO_3^{-}$ in the initial inflow waters [dots]) into the $Cl^{-}$ corner, where most of the water from the salt flats (crosses) occurs. In the cation diagram ($Na^{-}$-$Mg^{+2}$-$Ca^{+2}$) the water composition trends from the $Ca^{+2}$-$Mg^{+2}$ section of the diagram, where the inflow waters from the peripheral limestones (dots) lie, through the alluvial fans (squares) toward the $Na^{+}$ corner, where the salt flat waters (crosses) are located. The trend in the anion diagram ($Cl^{-}$-$SO_4^{-2}$-$HCO_3^{-}$) is from the $SO_4^{-2}$ area (low amounts of $HCO_3^{-}$ in the initial inflow waters [dots]) into the $Cl^{-}$ corner, where most of

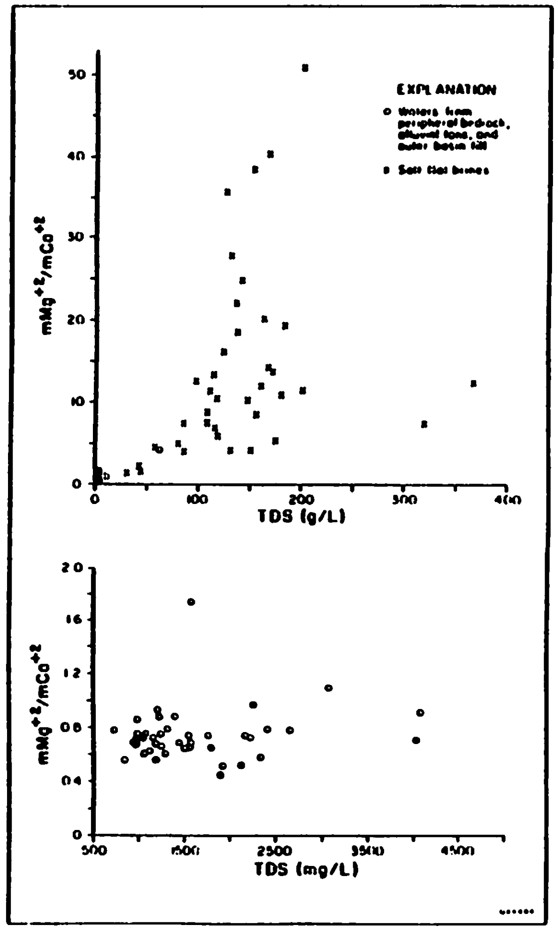


**Figure 5.** $Mg^{+2}/Ca^{+2}$ versus TDS. Continuous loss of $Ca^{+2}$ by precipitation of carbonate and sulfate minerals occurs as water is evaporated and results in a $Mg^{+2}$-rich solution. In the northern Salt Basin, the inflow waters have $Mg^{+2}/Ca^{+2}$ ratios less than unity. Within the flats, all $Mg^{+2}/Ca^{+2}$ ratios are greater than unity, increasing to as much as 51:1, but generally ranging from 5:1 to 15:1. Inset shows the $Mg^{+2}/Ca^{+2}$ ratio of inflow waters with low TDS.

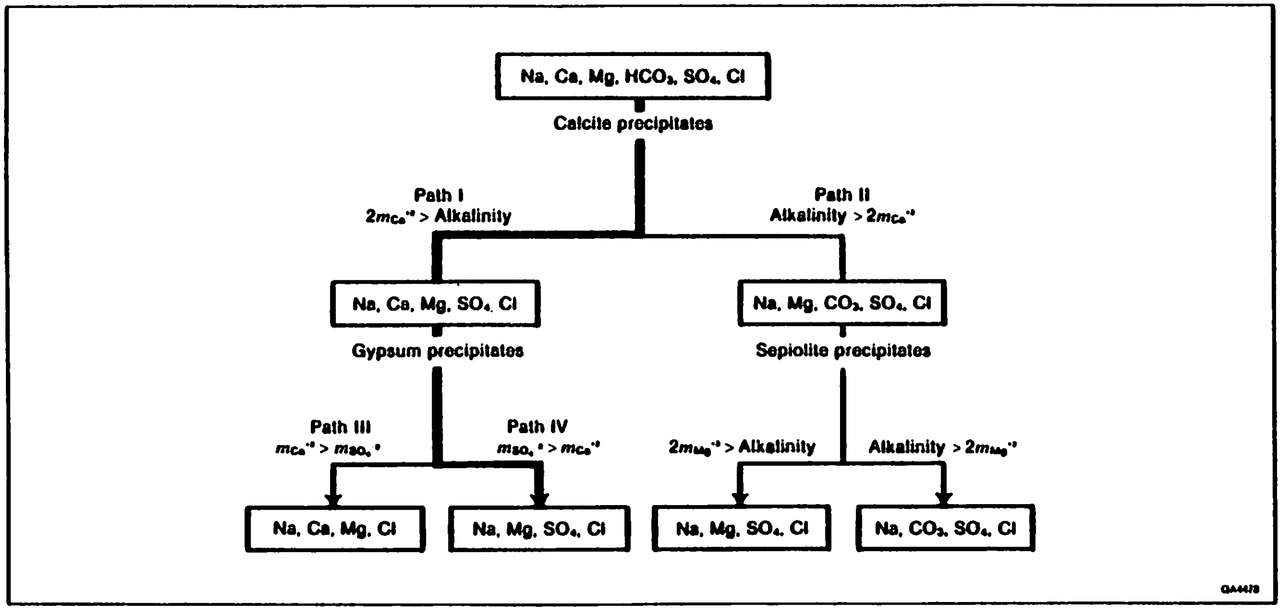


**Figure 6.** Brine evolution and mineral precipitation sequence (modified from Hardie and Eugster, 1970, and Drever, 1982). According to the initial relative concentrations of $Ca^{+2}$, $HCO_3^-$, and $SO_4^{-2}$, the initial inflow waters in the northern Salt Basin should evolve to a brine with $Cl^-$ and $SO_4^{-2}$ as the major anions and $Na^+$ and $Mg^{+2}$ as the major cations, following the highlighted path. In the northern Salt Basin, the $Ca^{+2}$ concentrations are initially higher than the $HCO_3^-$; with evaporative concentration and increasing TDS, calcite will begin to precipitate, and the waters will evolve to a $SO_4^{-2}$-$Cl^-$ brine. Because the inflow waters contain appreciable amounts of $SO_4^{-2}$, gypsum will be the next mineral to precipitate. Because of calcite precipitation, $SO_4^{-2}$ is greater than $Ca^{+2}$; with gypsum precipitation, $SO_4^{-2}$ will build up in solution, $Ca^{+2}$ will be depleted, and the water will evolve to a Na-Mg-$SO_4$-Cl brine.

ite precipitation, gypsum precipitation, and dolomite precipitation. There is a relative loss of Mg (relative to Na or Cl), as indicated on anion diagrams and on chemistry-versus-distance plots (Figs. 2 and 3). Presumably, the Mg is being precipitated in dolomite. More Mg is probably not lost because of the slow rate of dolomite precipitation and the limited amount of carbonate to precipitate with it. All of the salt flat brines, represented by crosses, lie along the $Mg^{+2}$-$Na^{+}$ line and in the $Na^{+} + K^{+}$ corner. The inflow Ca-Mg-$SO_4$ waters in the northern Salt Basin move down hydraulic gradient, from the peripheral limestone aquifers, through the alluvial fans and outer basin fill (Ca-Na-Mg-$SO_4$-Cl composition), and evolve to a final composition of a Na-Mg-$SO_4$-Cl brine in the salt flats.

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## Gypsum and Halite Precipitation

Gypsum precipitation by ground-water evaporation is an important geologic process in forming the salt flats. The puffy playa surfaces indicate ground-water discharge. Pore-fluid brine chemistry results from ground-water evaporation, calcite precipitation in the alluvial fans, and then gypsum precipitation in the salt flats. Trenches and cores indicate that the salt flats are composed predominantly of gypsum.

The amount of gypsum precipitation can be estimated if the input concentration of $Ca^{+2}$ and $SO_4^{-2}$ in the recharge water (ground water in the Permian limestones) and the evaporation rate from the salt flats are known. If the system is considered to be steady state, then the volume of water discharging by evaporation is equal to the volume of water flowing from the surrounding rocks and sediments to the salt flats. The amount of dissolved $CaSO_4$ being transported to the salt flats and precipitated by evaporation is approximated by the average $CaSO_4$ concentration calculated for these recharge waters. Additional sulfate is not added to the salt flats by dissolving salt flat gypsum because these waters are already saturated with respect to gypsum by the time they reach the evaporite sediments.

Average $CaSO_4$ concentration of the input water is 760 mg/L, and the estimated evaporation rate ranges from 18 cm/yr (Allmendinger and Titus, 1973) to 82 cm/yr (Bjorklund, 1957). Using a gypsum density of 2,320 mg/cm$^3$ and a porosity of 0.3, we estimate that evaporation results in precipitation of 0.02 to 0.09 cm$^3$/yr of gypsum, or a deposition rate of 20 to 90 cm/1,000 yr. This accumulation rate represents an average maximum because wind deflation is not considered. Gypsum will accumulate until the level exceeds the limit of the sediments being kept moist by the upward flow of water from the water table. These deposition rates, however, are comparable to other chemical sedimentation rates for sabkha environments; Illing and others (1965) and Kinsman (1966) estimated accumulation rates of 1 m/1,000 yr for Sabhka Faishak (Persian Gulf) and 0.5 m/1,000 yr for sabkhas on the Trucial Coast, respectively.

Halite should also be accumulating, considering that salt has been mined from the flats in the past (see appendix). Estimates of halite accumulation are made using the approach previously described for calculating gypsum accumulation rates. A range of estimated evaporation rates of 18 to 82 cm/yr is again assumed. The average NaCl concentration of the ground water in the Permian limestones is 207 mg/L. Using a halite density of 2,170 mg/cm$^3$ and a porosity of 30 percent, we estimate that evaporation results in the precipitation of 0.006 to 0.02 cm$^3$/yr of halite, or a depositional rate ranging from 6 to 20 cm/1,000 yr. The salt flats are gypsum-dominated rather than halite-dominated for two reasons: (1) the recharge water has lower initial NaCl concentrations than gypsum concentrations; therefore, more gypsum will ultimately precipitate than NaCl, and (2) because halite is more soluble, it will precipitate mainly in the unsaturated section close to land surface and be more subject to wind deflation.

## CONCLUSIONS

The northern Salt Basin is a hydrologically closed, arid basin in which the end product of ground-water flow is a series of playas, or salt flats. The largest percentage of ground-water inflow to the salt flats is contributed by rain-water that originates in the surrounding highlands and alluvial fans and moves as ground water into the basin. A much smaller volume of recharge is from precipitation directly onto the basin. Discharge occurs primarily by evaporation from the flats. The northern Salt Basin playas are influenced by ground water and, to a lesser extent, by surface water, as evidenced by trends in shallow ground-water chemistry, shallow subsurface stratigraphy, and playa surface features.

Evaporation occurring in the basin and, in particular, in the flats is thought to be the most influential process in the development of the chemical composition of the shallow ground water. Evaporative concentration produces a compositional trend with increases in salinity as ground water moves down the hydraulic gradient from the mountains, plateaus, and alluvial fans surrounding the basin, through the basin fill, and discharges from the salt flats. Evaporation above the water table in the flats concentrates the water to the composition of a $Cl^-$ brine in which total dissolved solids values range from 50,000 to greater than 300,000 mg/L.

The concept of brine evolution can be applied to the saline water undergoing evaporation in the northern Salt Basin. As fluids are removed by evaporation and the concentrations of dissolved components increase, certain minerals are removed from solution by means of precipitation, and the remaining solution becomes enriched or depleted in one ion or another. Calcite, the least soluble of the minerals, is the first to precipitate, resulting in the reduction of the carbonate species ($HCO_3$-$CO_3$). Gypsum then precipitates; initial molar ratios in favor of $SO_4^{-2}$ ($Ca:SO_4<1$) dictate that $SO_4^{-2}$ concentrations will increase and $Ca^{+2}$ will be depleted. Dolomite appears to be precipitating in the high-Mg brines. In the northern Salt Basin, the dilute inflow water evolves from a Ca-Mg-$SO_4$ water to a Na-Mg-$SO_4$-Cl brine. These processes are modified by surface-water infiltration and dilution.

Gypsum and other evaporitic minerals are currently accumulating in the salt flats by evaporitic discharge of ground water. On the basis of estimated accumulation rates, we conclude that precipitation of gypsum by ground-water evaporation is a significant depositional process. The salt flat sediments are not relict Pleistocene lacustrine deposits. This does not, however, preclude the presence of a lake within the northern Salt Graben during Pleistocene time.

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## APPENDIX

### History of Salt Mining at the Salt Lakes

Although all of the lakes within the Salt Basin are saline, only the so-called "Zimpleman Salt Lake" (Figs. 1, 2a and 2b), which is about 1 km long and .5 km wide, contains salt in appreciable quantities. It is located 1 km south of Highway 62-180 (location of well 48-24-020). Prior to the 1950s, this lake was minable and was the source of all salt used by farmers and ranchers within a radius of 150 km or more. Older workings, in what was called the "Maverick Salt Basin," about 3 km south of Zimpleman Salt Lake, were closed around 1900.

According to Richardson (1904), the presence of salt had been known to the Mexicans who, "in the early days of the occupation of the country . . . traveled for it from distant parts of Chihuahua." The first wagon road to the salt deposits was built in 1863. In 1877, the salt became the subject and cause of the "Great Salt Wars" of Texas. Mr. Charles Howard, a Missouri lawyer, attempted to stake a claim on the lakes and to collect a charge for the salt, which previously had been free to all. Local political strife and ill feelings toward Howard in the Mexican population resulted in riot and bloodshed in nearby San Elizario, and peace was restored only after intervention by the U.S. Army.

Zimpleman Salt Lake was a chief source of salt for the area until the 1950s. In 1904, Richardson observed a few scattered occurrences of "beautiful hopper-shaped crystals" forming after rain in localities where the surface salt had been recently removed. The abundant "surface salt" was the commercially valuable deposit, commonly attaining thicknesses of greater than 2 cm. According to Richardson (1904), "when the surface layer of salt is removed, its place is taken by . . . brine, which evaporates and deposits salt, so that, within a few weeks after stripping an area, salt completely replaces that which was removed." This strongly suggests that evaporation of shallow ground water and surface water is an important mechanism of evaporite deposition in the salt flats. The supply was generally believed to be inexhaustible. At that time, the cost of salt was $1 per horse load.

In 1913, Mr. Arthur Grable established a homestead on the salt flats, leased an area including the salt lake, and charged 50 cents a head for whatever a man could haul out. According to his nephew, Mr. Clyde Grable (personal communication, 1980), wagons would be loaded up with so much salt that they could not move, and "Uncle Arthur" would have to pull them out of the playa with a donkey. After a rain, a "good salt" would form, and seed forks would be used to split the grainy top layers. Later, Arthur Grable built dikes, thereby forming brine vats to facilitate salt percipitation. King (1948) described the lakes as follows:

> On May 26, 1946, the entire surface of the lake, inside the outer dike, was covered with a crust of salt that averaged about half an inch thick. This crust was nearly free of wind-blown sand and clay and so must have formed since the heaviest sand storms in March. To judge by the taste and appearance, the crust is mainly sodium chloride, although the somewhat bitter taste of sulfates can be detected in it. Brine is present immediately below the surface crust and this, in turn, is underlain by the next solid material, which is salt mixed with clay and fine sand. This layer of clayey salt is about six inches thick, according to Mr. Grable, and forms a "hardpan" that will support a loaded truck. Beneath the "hardpan" the salt, clay, and sand is [sic] soft, porous, and permeable.

According to Dunlap (in King, 1948), records indicate almost continuous production from the Zimpleman Salt Lake from 1911 to 1946; Dunlap estimates that the total production from the lake was between 5,000 and 15,000 tons during that period. King (1948) reports that by 1948 the demand for salt had begun to decline but that there would "probably continue to be a small local market for the product." Today, however, no salt is worked or produced commercially from the salt flats of the northern Salt Basin.

New Mexico Geological Society Guidebook, 31st Field Conference, Trans-Pecos Region, 1980

# GIANT DESICCATION POLYGONS IN WILDHORSE FLAT, WEST TEXAS

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## INTRODUCTION

Giant polygonal contraction cracks have developed in a restricted portion of the Salt Basin playa. These fractures are located in "Wildhorse Flat" on the terminus of an alluvial fan originating in the eastern Baylor Mountains at 31°15' N latitude, 104°43' W longitude (figs. 1 and 2).

They were first noticed on a 1946 Edgar Tobin aerial photograph and again on a 1953 aerial photograph. Pratt (1958) photographed and discussed the growth of these features. Underwood and DeFord (1975) described the Baylor Mountain fractures and compared them to those on Eagle Flat, West Texas. In February, 1977, I flew over and photographed these polygons. Figure 3 shows the development of the fractures sketched from aerial photographs taken in 1946, 1953, 1957, 1961, and 1977.

The long history of drought in the region is well documented. Groundwater level is a critical factor in the development of giant desiccation cracks and the combination of drought and pumping for irrigation strongly affects its position. Figure 4 shows a plot of the water-level fluctuations in the wells of Hudspeth and Culberson Counties. Only selected wells were measured by the Texas Water Development Board; they were not measured every year, nor were they checked at regular intervals. It is important to note that the number of wells has increased greatly and that they draw water from several different aquifers.

It can be observed that for all but six years (1949, 1955, 1956, 1961, 1966, 1972) since records began to be kept in 1948, the number of wells in which the water level fell three meters exceeded or equaled the number of wells in which the water table rose three meters.

Water well 47-43-701 pumps from the Salt Basin aquifer; it is less than a kilometer north of the giant contraction cracks. Water-level fluctuations indicate the conditions under which the fractures grew: The water level fell twelve meters between 1953 and 1964. It rose nine meters between 1964 and 1965. Between 1965 and 1970 the water table remained within a half meter of the 1965 level. The level fell four and one-half meters in the year 1970 to 1971 and rose the same amount in the ten months between February and December, 1971, (Texas Water Development Board, 1977).

## ORIGIN OF BAYLOR MOUNTAIN POLYGONS

The most important process contributing to the formation and growth of the giant contraction cracks at the base of the Baylor Mountains is desiccation. Prolonged drought in Salt Basin and the increasing rate of water withdrawal in the Wildhorse subbasin combine to create a severe drying effect both on the playa surface and in the subsurface.

Resistivity work by White and others (1977) indicates a large percentage of clay in the soils fractured by desiccation. The soils of the playas and fans in the Wildhorse subbasin are rich in carbonates, particularly calcite, as they are weathered from the limestones of the Beach, Baylor, and Apache Mountains, as well as the Diablo Platform. Sheet silicates may have been derived from the sandstone units and talc deposits farther upstream from the playas. Thus, the soils supporting these fractures exhibit many of the characteristics indicated by Langer and Kerr (1966) as conducive to desiccation fracturing.

Three geologic processes are suggested as possible causes of the stresses necessary to produce the observed orthogonal pattern of the desiccation cracks. The simplest explanation is that the fractures are relict, inherited from the drying of Pleistocene lakes. The original fractures would have paralleled the receding lakeshore, while secondary fractures would have developed orthogonal to the lakeshore. However, the present fracture patterns are not confined within the boundaries of the older lakes; instead, the fractures appear only on the west side of the subbasin in three or

*Figure 1. Looking east across the northern half of the giant desiccation cracks in the Baylor Mountain fan. Light-colored line crossing north-south is a ranch road.*

*Figure 2. Looking east across the southern half of the desiccation cracks in the Baylor Mountain fan. The Apache Mountains are on the horizon.*

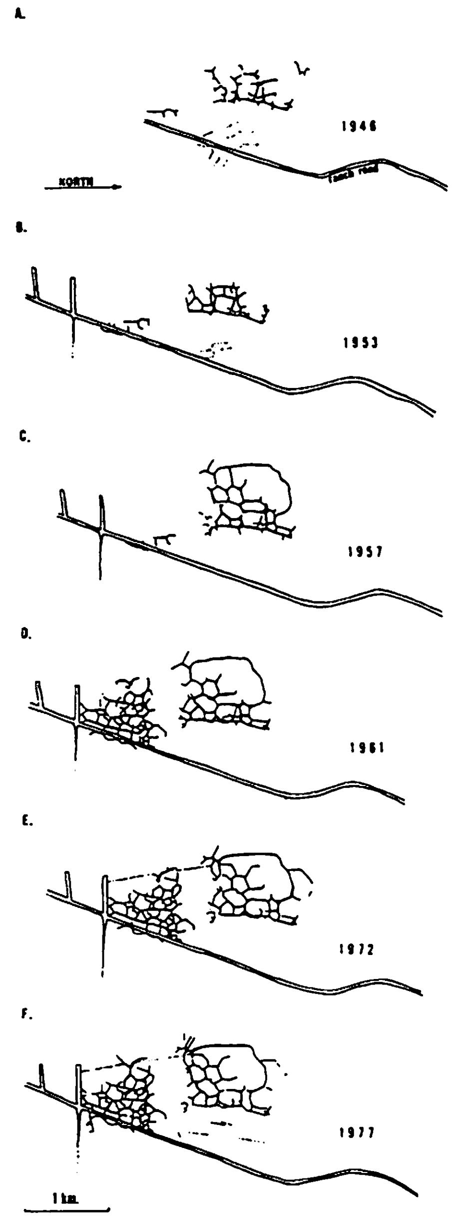


*Figure 3. Orthogonal views of the desiccation cracks, sketched from aerial photographs taken in years indicated.*

possibly four locations. Primary fractures are aligned subparallel to north-trending fault scarps left by Quaternary tectonism (fig. 5).

A second possibility is that continuing subsidence of the western side of the basin floor, coupled with the lowering of the piezometric surface by drought and pumping, could cause the groundwater surface to withdraw down a slope. This would also create an orthogonal pattern similar to that described for the drying Pleistocene lakes, but the cracks would be more likely to form along the west side than around the entire basin perimeter.

The singular parallelism of the primary fractures leads one to suspect a third possible mechanism, interaction with a fault cutting older Holocene alluvium. If such a fault exists, the related gouge would act as a semipermeable boundary along the western edge of the subsiding basement block; the fault would disturb the radial tension field produced by simple basin subsidence coupled with the drying of a playa lake bed. Uneven subsidence in the alluvium across the faulted basement blocks would increase the tension already present in the capillary zones due to desiccation. If the fault zone acted as a semipermeable barrier, as Kreitler (1976) has described, then horizontal tensions in the alluvial capillary zones would be increased dramatically as aquifers on the upslope side of the fault were recharged by ephemeral streams crossing into the graben, while aquifers on the other side of the fault were depleted by water withdrawal for irrigation.

The network of desiccation cracks at the base of the Baylor Mountains has more than doubled in area since Pratt (1958) reported primary fracture lengths of 600 and 1050 meters. The importance of soil type and the proximity to moisture at depth in controlling the development of these fractures is suggested by the slower propagation rate of the fractures that are exposed to runoff or playa flooding. Fractures on higher ground or farther away from stream courses have grown more quickly.

Soil differences may account for the break in the primary fractures in the polygonal pattern at the base of the Baylor Mountains. A tongue of coarse fanglomerate may cover this area, obscuring the effects of desiccation while transmitting the induced stresses to the clay-rich soils about it.

## CONCLUSION

The Baylor Mountain contraction cracks are desiccation features produced in hard calcite-rich clay soils. The polygonal pattern of fractures with some orthogonal trends was caused by an additional east-west tensional stress. This stress probably resulted from recent tectonic subsidence of the basin, coupled with effects of water withdrawal for irrigation. Continued pumping in the basin will probably further increase the area covered by these giant polygonal fractures.

## ACKNOWLEDGMENTS

The bulk of the data for this paper comes from master's thesis research; the guidance and encouragement of master's committee chairman, W. R. Muehlberger is asknowledged gratefully. I wish to express special gratitude to the Minerals Department, Conoco, Inc., Albuquerque office and particularly D. W. Wentworth, District Geologist. Field support was from National Aeronautics and Space Administration Grant NSG 7250.

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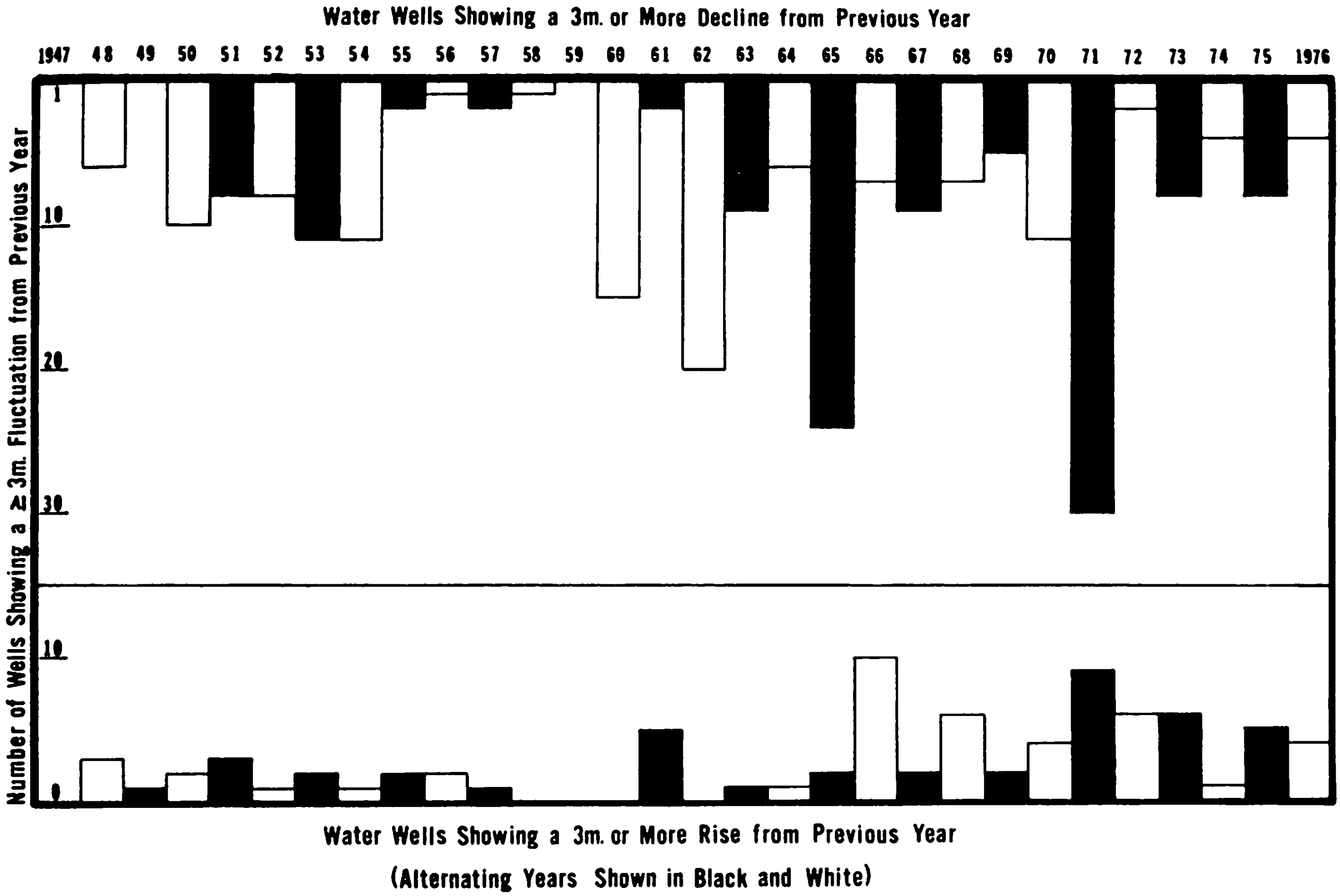


*Figure 4. Yearly fluctuations of levels in water wells of Hudspeth and Culberson Counties, Texas (data from Texas Water Development Board).*

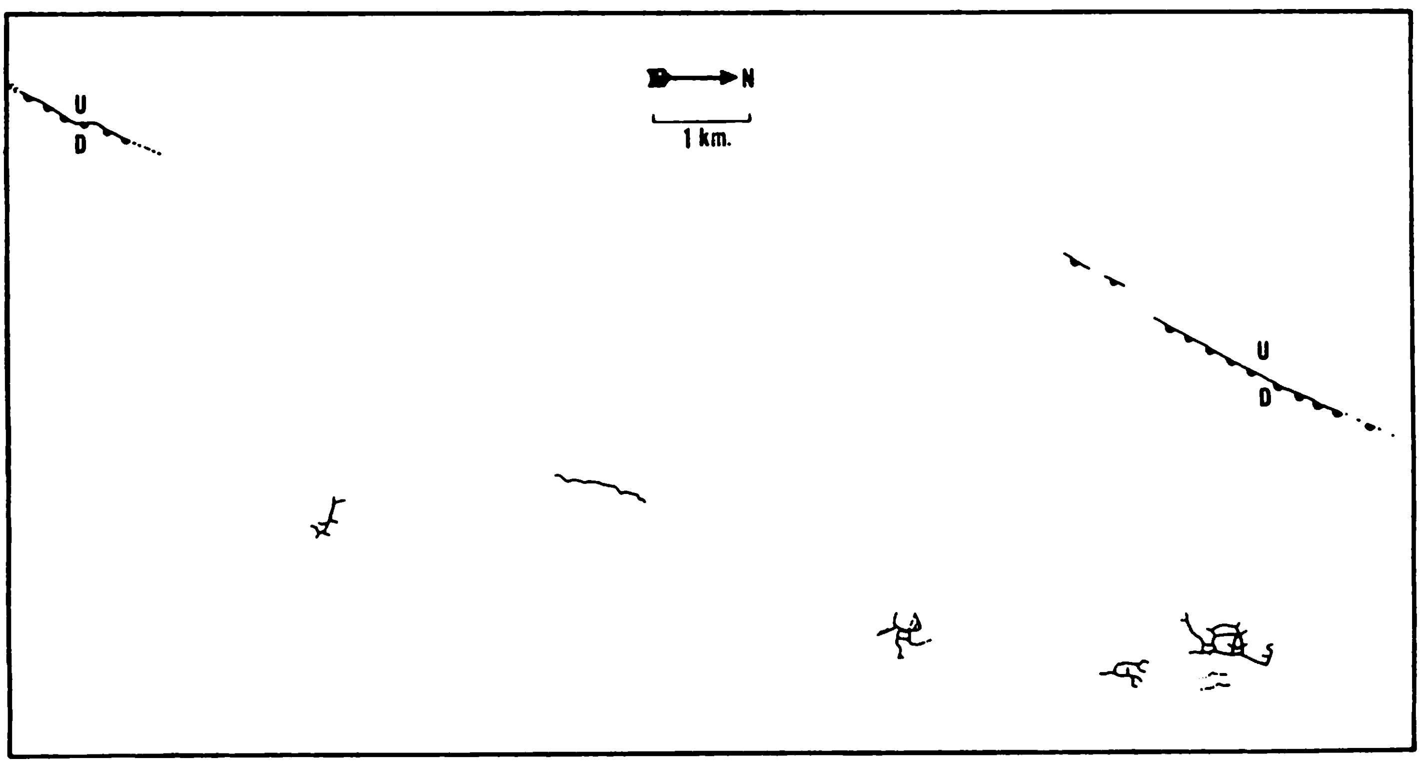


*Figure 5. Subparallel orientation of alluvial fault scarps and giant desiccation cracks, as seen on 1953 aerial photographs.*

# MOVEMENT OF GROUND WATER IN PERMIAN GUADALUPIAN AQUIFER SYSTEMS, SOUTHEASTERN NEW MEXICO AND WESTERN TEXAS

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## AQUIFER SYSTEMS

Permian Guadalupian-age strata can be divided into three aquifer systems. Hiss (1975a, p. 132) described and named them the Capitan, shelf, and basin aquifers (fig. 1). In most areas, they are readily distinguished by differences in lithology, geographic position, stratigraphic relationships, hydraulic characteristics, and quality of the contained water (Hiss, 1975b and c; 1976a).

### Capitan Aquifer

The Capitan aquifer is a lithosome that includes the Capitan and Goat Seep Limestones and most or all of the Carlsbad facies of Meissner (1972). Shelf-margin carbonate banks or stratigraphic reefs in the upper part of the San Andres Limestone are included within the Capitan aquifer where they cannot be readily distinguished from the Goat Seep Limestone and Carlsbad facies (Silver and Todd, 1969, figs. 12 and 13).

### Shelf Aquifers

Saturated strata yielding significant quantities of water from the San Andres Limestone and the Bernal and Chalk Bluff facies of Meissner (1972) constitute the shelf aquifers. The lithologic contact between the Capitan and shelf aquifers is gradational and is difficult to discern with accuracy in some areas. Observations of the geometry and lithologic relationships of the shelf-margin rocks in the field suggest that the width of the Capitan Limestone (reef) is considerably less than is shown in many geologic reports (Dunham, 1972, fig. I-1).

The present-day ground water regimen is strongly influenced by the Pecos River in New Mexico. As a result, the hydraulic conductivity of the shelf aquifers west of the Pecos River has been greatly enhanced by the leaching of soluble beds from the Chalk Bluff facies (Meissner, 1972; Motts, 1968). Locally and west of the Pecos River valley between Carlsbad and Roswell, the hydraulic conductivities of the shelf aquifers are quite large and may be similar to that of the Capitan aquifer. The hydraulic conductivity of the shelf aquifers in the Carlsbad and Roswell underground water basins is several orders of magnitude higher than that generally encountered in the shelf aquifers east of the Pecos River at Carlsbad. The water contained in the shelf aquifers is also much better in the shallow zones exploited in these basins than elsewhere in the same aquifers within the area studied. East of the Pecos River near Carlsbad the hydraulic conductivity of the shelf aquifers is generally one to two orders of magnitude less than that of the Capitan aquifer.

### Basin Aquifers

Saturated strata yielding significant quantities of water from the Brushy Canyon, Cherry Canyon and Bell Canyon Formations of the Delaware Mountain Group are referred to as the basin aquifers. Although the Capitan aquifer abuts and overlies the Delaware Mountain Group along the margin of the Delaware Basin, the lithologic and hydrologic characteristics of the basin and Capitan aquifers are quite different. The average hydraulic conductivity of the basin aquifer ranges from one to two orders of magnitude less than that of the Capitan. Therefore, only a relatively small amount of water can be expected to move from the basin aquifers to the Capitan aquifer, or vice versa. The difference in quality of water contained in the two aquifers—relatively good in the Capitan, bad in the basin—is also a distinguishing characteristic (Hiss, 1975b).

## CONSTRUCTION OF POTENTIOMETRIC SURFACES

Reliable pressure-head and water-level data were adjusted to freshwater heads to construct generalized potentiometric surfaces representative of two conditions in the three aquifer systems. Figure 2 is a map representing conditions in the aquifer systems prior to both development of water supplies for irrigation and discovery and production of oil and gas and associated waste water. Figure 3 is a similar map representing the shelf and basin aquifer for the period 1960 to 1969 and of the Capitan aquifer for the latter part of 1972.

A potentiometric surface represents hydraulic head in an aquifer; the general direction of ground-water movement is inferred to be normal to the illustrated head contours. Hiss (1975, p. 220-255) discusses the computation of ground-water head and the procedures followed in determining the heads used in these maps. The potentiometric maps support the inferred movement of water shown in figure 4.

## MOVEMENT OF GROUND WATER

During the latter part of the Cenozoic Era, the movement of ground water through the rocks of Permian Guadalupian age in southeastern New Mexico and western Texas has been controlled or influenced by the following: (1) the regional and local tectonics; (2) the evolution of the landscape; (3) the relative transmissivities of the various aquifers; (4) the amount of recharge; and (5) the exploitation of the petroleum and ground-water resources in the last five decades (fig. 4).

### Control by Regional Tectonics

The flow of ground water through the shelf, basin and Capitan aquifers after the uplift of the Guadalupe and Glass Mountains but prior to the excavation of the Pecos River valley at Carlsbad is shown diagrammatically in figure 4A. The three aquifer systems were recharged by water originating as rain or snowfall on the outcrops along the western margin of the Delaware Basin. Evidence of major surface drainage within the Trans-Pecos area of southeastern New Mexico and western Texas has not been reported.

Ground water moved generally eastward and southeastward through the shelf and basin aquifers under a gradient of probably only a few feet per mile toward natural discharge areas along

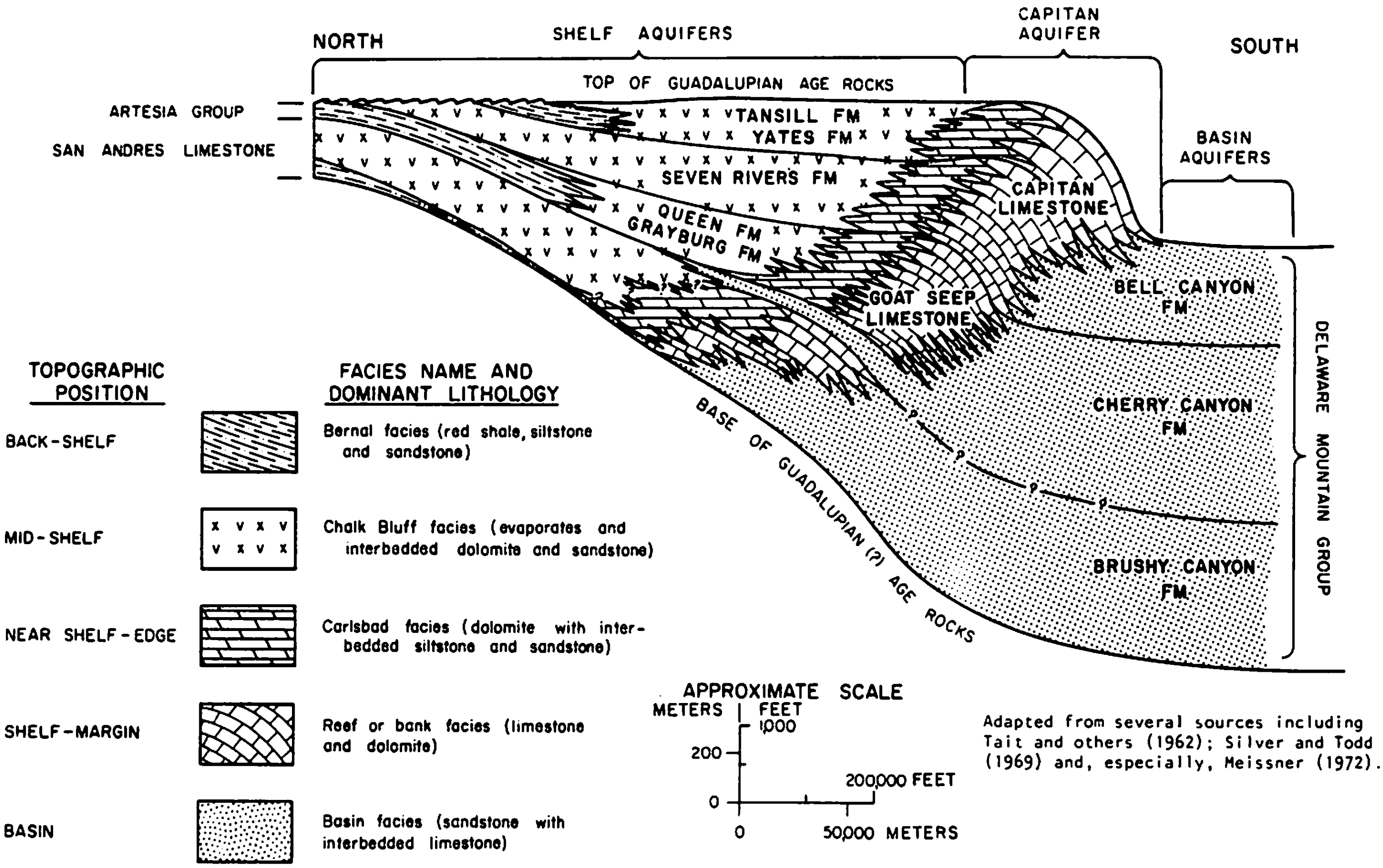



*Figure 1. Highly diagrammatic north-south stratigraphic section showing the positions and relationships of the major lithofacies in the rocks of Guadalupian age, eastern New Mexico.*

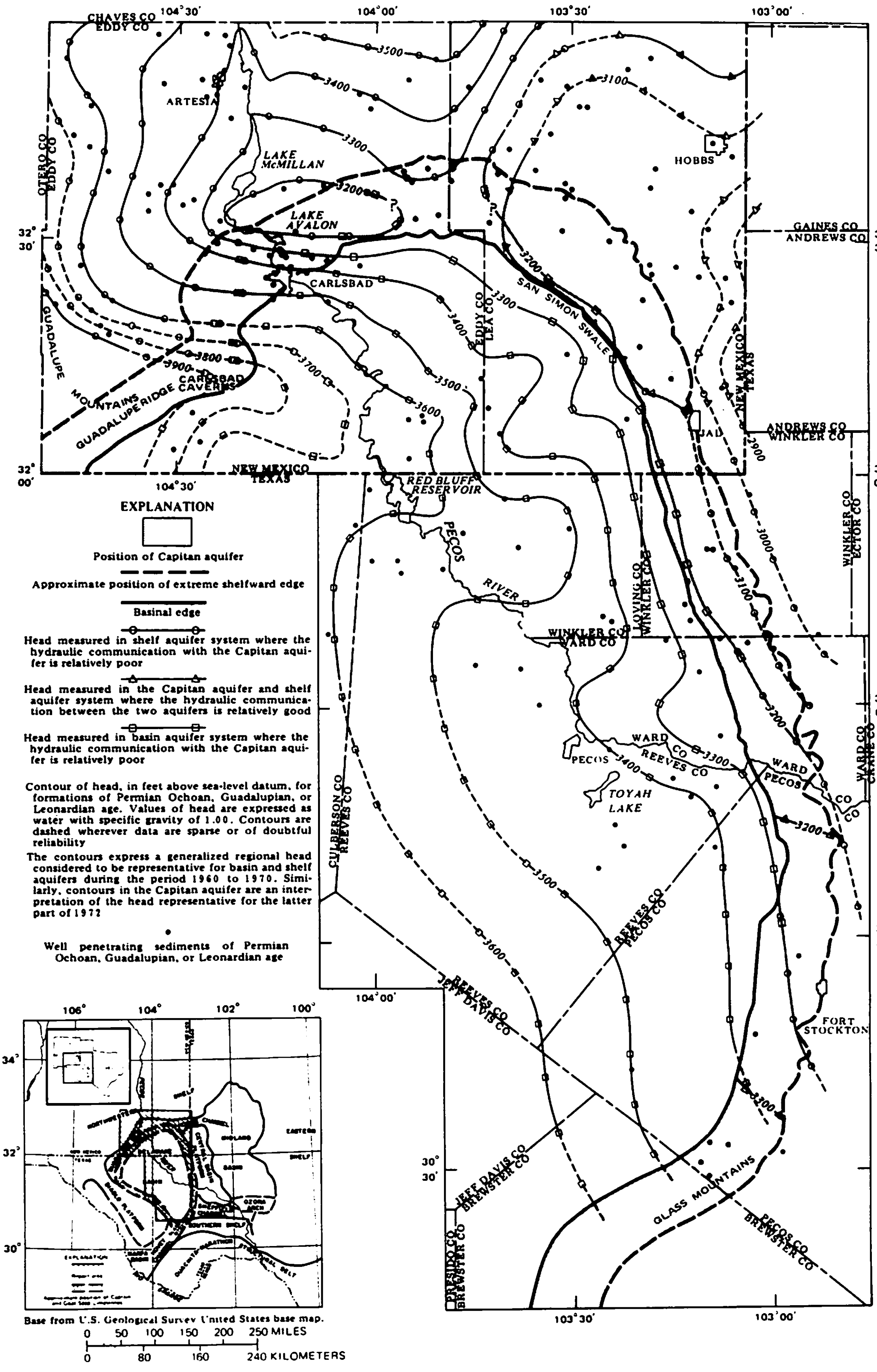



*Figure 2. Pre-development potentiometric surface.*

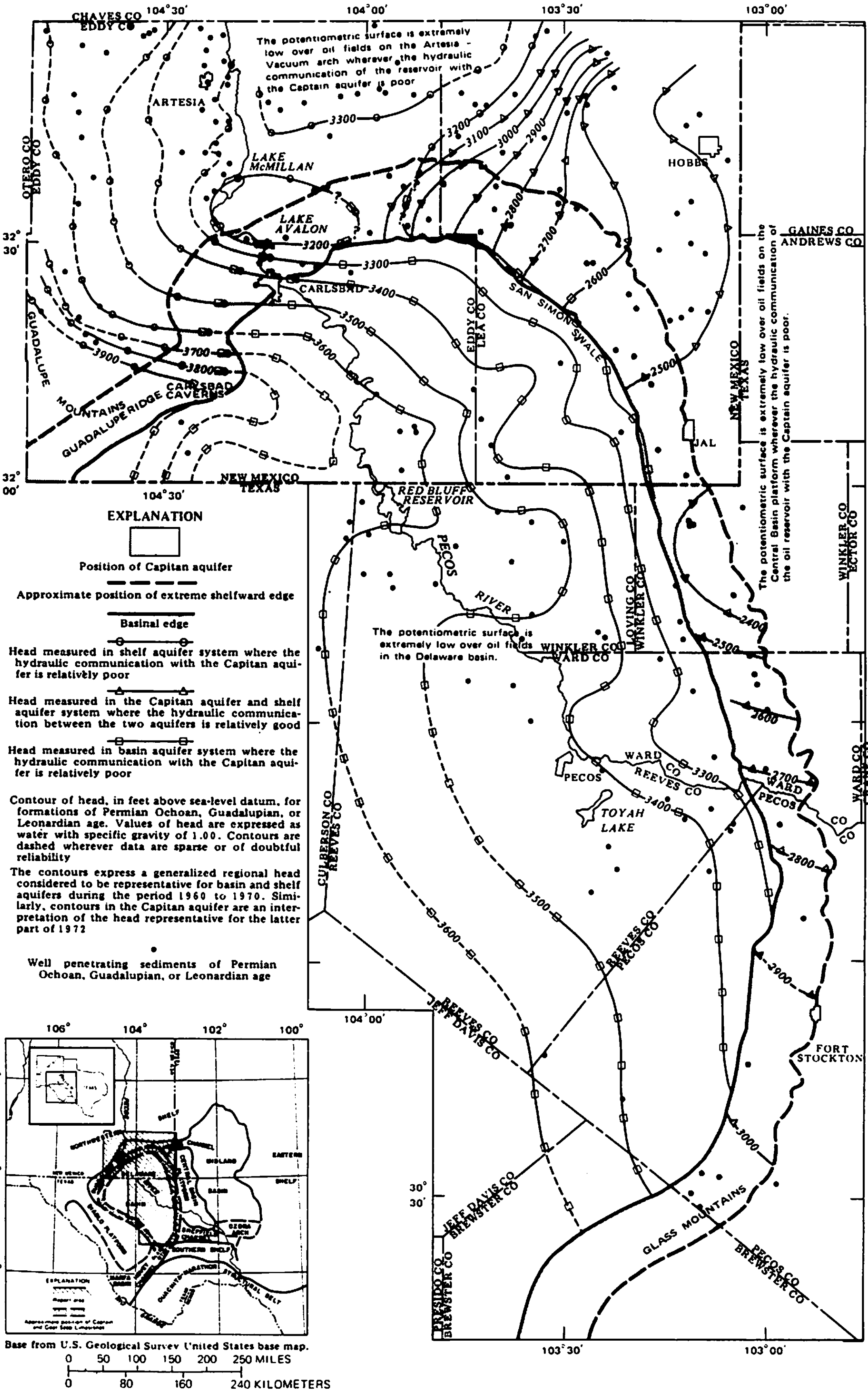
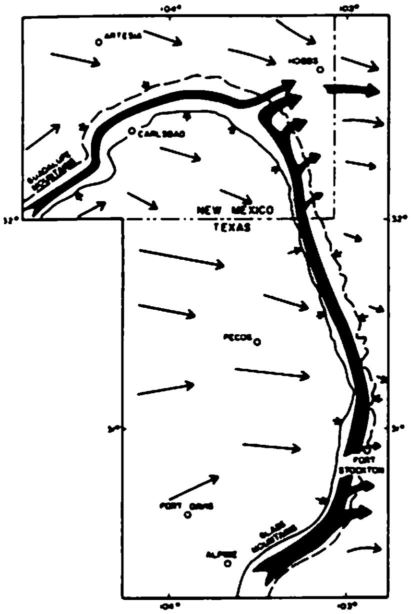
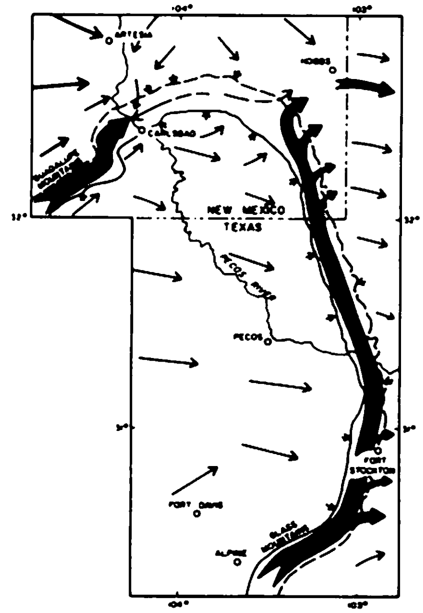


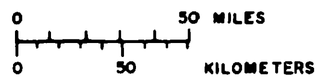

*Figure 3. Post-development potentiometric surface.*




A. Regimen principally controlled by regional tectonics prior to development of the Pecos River.




B. Regimen influenced by erosion of Pecos River at Carlsbad downward into hydraulic communication with the Capitan aquifer.

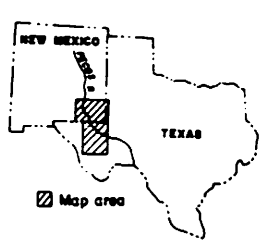



EXPLANATION

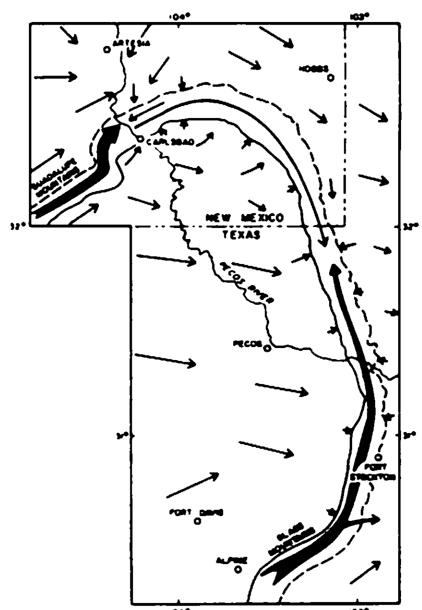
Capitan aquifer

Highly diagrammatic ground-water flow vectors:

1. Vector size indicates relative volume of ground-water flow.
2. Orientation indicates direction of ground-water movement.




INDEX MAP




C. Regimen influenced by both communication with the Pecos River at Carlsbad and the exploitation of ground-water and petroleum resources.

*Figure 4. Diagrammatic maps depicting the evolution of ground water regimens in strata of Permian Guadalupian age in southeastern New Mexico and western Texas.*

streams draining to the ancestral Gulf of Mexico. Water entering the Capitan aquifer in the Guadalupe Mountains moved slowly northeastward and then eastward along the northern margin of the Delaware Basin to a point southwest of present-day Hobbs. Here it joined and comingled with a relatively larger volume of ground water moving northward from the Glass Mountains along the eastern margin of the Delaware Basin. From this confluence, the ground water was discharged from the Capitan aquifer into the San Andres Limestone, where it then moved eastward across the Central Basin Platform and Midland Basin, eventually to discharge into streams draining to the Gulf of Mexico.

## Influence of Erosion of Pecos River at Carlsbad

Some time after deposition of the Ogallala Formation, perhaps early in Pleistocene time, the headward-cutting Pecos River extended westward across the Delaware Basin to the exposed soluble Ochoan beds. It then turned northward following this natural weakness in the sedimentary rocks to pirate the streams draining to the east from the Sacramento and Guadalupe Mountains (Plummer, 1932; Bretz and Horberg, 1949b; Thornbury, 1965). As the excavation of the Pecos River valley progressed, the hydraulic communication with formations of Guadalupian age gradually increased until the Pecos River functioned as an upgradient drain. Eventually, the hydraulic gradients in the shelf, basin and Capitan aquifer were reversed along the eastern side of the Pecos River valley, and ground water that formerly flowed eastward was diverted westward as spring flow into the Pecos River (fig. 4B). Water recharged to the same aquifers in the Guadalupe Mountains began to follow the shorter path to springs in the Pecos River. Many of the solution features observed in the Guadalupian sedimentary rocks west of the Pecos River near Carlsbad probably were initiated during this period.

Movement of water eastward toward Hobbs from the Guadalupe Mountains into the Capitan aquifer was decreased by the lowering of the hydraulic head along the Pecos River. At the same time, a trough in the potentiometric surface of the shelf and basin aquifers began to develop east of Carlsbad, and water began to drain into the Capitan aquifer from the surrounding sedimentary rocks. Meanwhile, ground water continued to move northward from the Glass Mountains in the Capitan aquifer toward a point of discharge into the San Andres Limestone southwest of Hobbs. This part of the aquifer was unaffected by the cutting of the Pecos River valley across the Delaware Basin and the Central Basin Platform.

## Influence of Exploitation of Ground Water and Petroleum Resources

Regionally, the movement of ground water in the shelf and basin aquifers east of the Pecos River at Carlsbad has changed very little as a result of the exploitation of ground water and petroleum during a period of approximately 50 years (fig. 4C). Locally, however, the movement of ground water within these same aquifers is controlled by the effects of the numerous producing oil fields.

The shape of the regional potentiometric surface representative of the hydraulic head in the Capitan aquifer east of the Pecos River at Carlsbad has been changed significantly in response to withdrawal of both ground water and petroleum during the past 50 years. The westward movement of saline water from the Capitan aquifer in Eddy County east of Carlsbad into the Pecos River has been greatly diminished or eliminated by a reduction in hydraulic head.

Similarly, the movement of water in the San Andres Limestone and Artesia Group eastward across the northern part of the Central Basin Platform from New Mexico into Texas has been decreased. Eventually, the movement of water probably will be reversed. Water may be diverted from the San Andres Limestone and Artesia Group westward from Texas back toward Hobbs and then into the Capitan aquifer along the western margin of the Central Basin Platform. The effects of exploitation of the ground water and petroleum resources will continue to be the dominant factor influencing the movement of ground water in the Capitan aquifer for many years into the future.

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# Origins of Ground Water Discharging at the Springs of Balmorhea

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## INTRODUCTION

Six large springs arise in the Balmorhea* area of southwestern Reeves and northern Jeff Davis counties (Figure 1). The largest, San Solomon Spring, is located in Balmorhea State Park near Toyahvale. It flows into the bottom of a large swimming pool at an elevation of 3306 ft (1008 m). Phantom Lake Spring, the next largest, is at the northwestern edge of a solution collapse feature approximately seven km west-southwest of San Solomon Spring. It issues from an A-shaped opening in a bluff of lower Cretaceous limestone (Figure 2) at an elevation of 3475 ft (1059 m). Giffin Spring is on the other side of Highway 17 from Balmorhea State Park and issues at approximately the same elevation as San Solomon Spring. Saragosa Spring is located 0.5 mile (1.2 km) west of Balmorhea at an elevation of 3250 ft (991 m). West and East Sandia (Brogado) Springs are located just east of Balmorhea at elevations of 3205 ft and 3187 ft, respectively.

San Solomon, Giffin, Saragosa, West Sandia, and East Sandia Springs flow from gravels directly overlying the Lower Cretaceous limestones. The combined average discharge of San Solomon and Phantom Lake Springs is nearly 33,000 acre-feet/year. The combined discharge of the remaining springs is, by comparison, negligible.

The springs have long been an important source of water in this semiarid region. Brune (1975, 1981) reports that the springs were a popular campground for prehistoric people. Traces of irrigation canals used by the Jumanos, and later the Mescaleros have been found near San Solomon Springs. The Spanish explorer Espejo is believed to have visited all six springs on his journey through the Southwest in 1583. The siting of Fort Davis (Jeff Davis County) in 1854 was partly predicated on controlling the southern pass leading to the springs (then called Head Springs) in order to stop Comanche raiding parties. In the late nineteenth century, water from Phantom Lake and San Solomon Springs provided power for several mills and a cotton gin. Today these springs feed a network of canals which transport water to irrigate approximately 6000 acres (2428 ha). Winter spring flow is stored in the Balmorhea Reservoir. The Reeves County Water Control and Improvement District No. 1 oversees the delivery of water from the springs to participating irrigators.

With continued development of the area's ground-water resources, diminished spring flow is a definite possibility. Comanche and Leon Springs in the Fort Stockton area dried up in response to pumping of ground water. The canals leading from San Solomon Spring are now the sole habitat of the transplanted Comanche Springs pupfish. Heavy pumping of the Toyah Basin aquifer (LaFave and Sharp, 1986) led to drying up of Irving, Buck and Alamo Springs near Pecos, Texas (Brune, 1981). Consequently, it is a matter of practical importance, as well as scientific interest, to determine the source of recharge to the springs of Balmorhea.

Most previous investigations (White and others, 1941; Ogilbee, 1962; Couch, 1978; Pearson, 1985) have suggested that the recharge to the springs originates from precipitation in the nearby Davis Mountains. This model requires that the water enter the Lower Cretaceous limestone along the southwest limb of the Rounsaville syncline via seepage from the overlying volcanic aquifer and river gravels, and from creeks which lose their flow as they pass over the syncline. The water moves down dip, confined by low-permeability Upper Cretaceous strata, and discharges at the springs where the Upper Cretaceous is displaced by faulting (Figure 3). This is also the explanation shown at the display in Balmorhea State Park.

In addition, Couch (1978) and Harden (1972) have suggested that ground waters from the Capitan reef in the Apache Mountains and areas beyond contribute up to about 30,000 acre-feet per year to the spring flow.

Our analyses (LaFave and Sharp, 1986) indicate that the Capitan reef rocks in the Apache Mountains may contribute a significant amount of water to the springs. This conclusion is supported by comparison of water analyses from the Capitan reef, the Davis Mountains, and the springs; geologic structure between the Apache Mountains and the springs; and the regional flow system of the southern part of the Salt Basin. We also note that water in a significant portion of the Toyah Basin alluvial aquifer, between Balmorhea and Pecos, Texas, possesses a chemical signature very similar to waters from the Capitan reef.

## ANALYSIS OF SPRING WATERS

During periods of low rainfall Phantom Lake and Solomon Springs exhibit nearly constant discharge,* temperature, and water chemistry. The temperature of the water is about 78°F (25.6°C) for all the springs. The sodium-chloride-sulfate type water discharging from all the springs is mineralized and remarkably similar in chemical signature.

Piper and Schoeller plots of available analyses bear this out (Figures 4 and 5). However, after heavy rainfall spring discharge increases rapidly, temperature decreases, water salinity decreases, and turbidity increases. During peak discharge periods the mineralized flow is suppressed, and the springs preferentially flow fresher water. These characteristics imply two separate recharge sources for the springs - a large, distant one for the steady flow and a small, nearby one for the flashy or "storm" flow.

The chemical quality of the water in the Tertiary volcanic aquifer, underlying Cretaceous limestones, and streams

*Balmorhea is named for three of the early developers of the area - Balcom, Morrow, and Rhea.

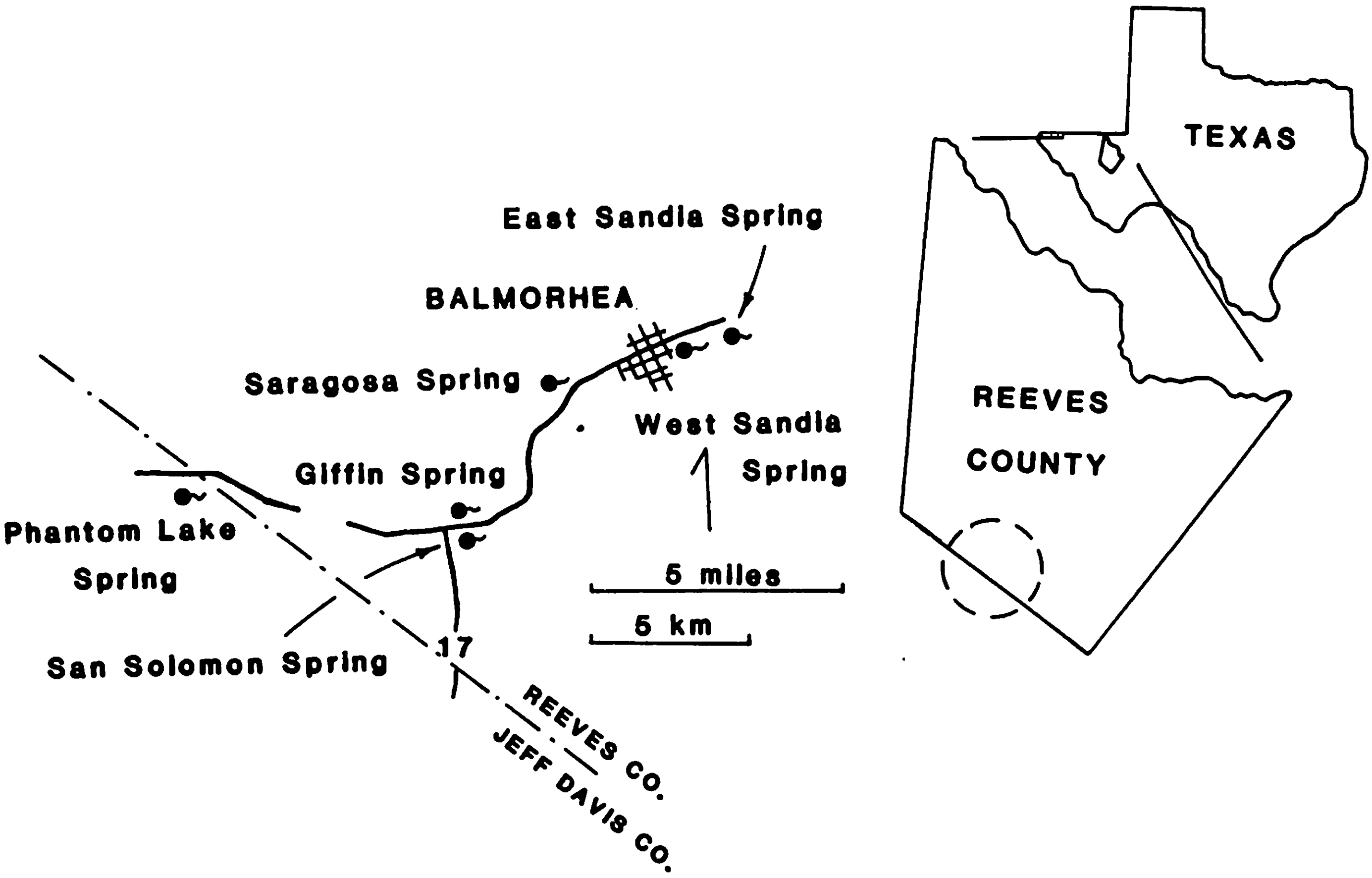


Figure 1. Location of springs in the vicinity of Balmorhea, Texas.

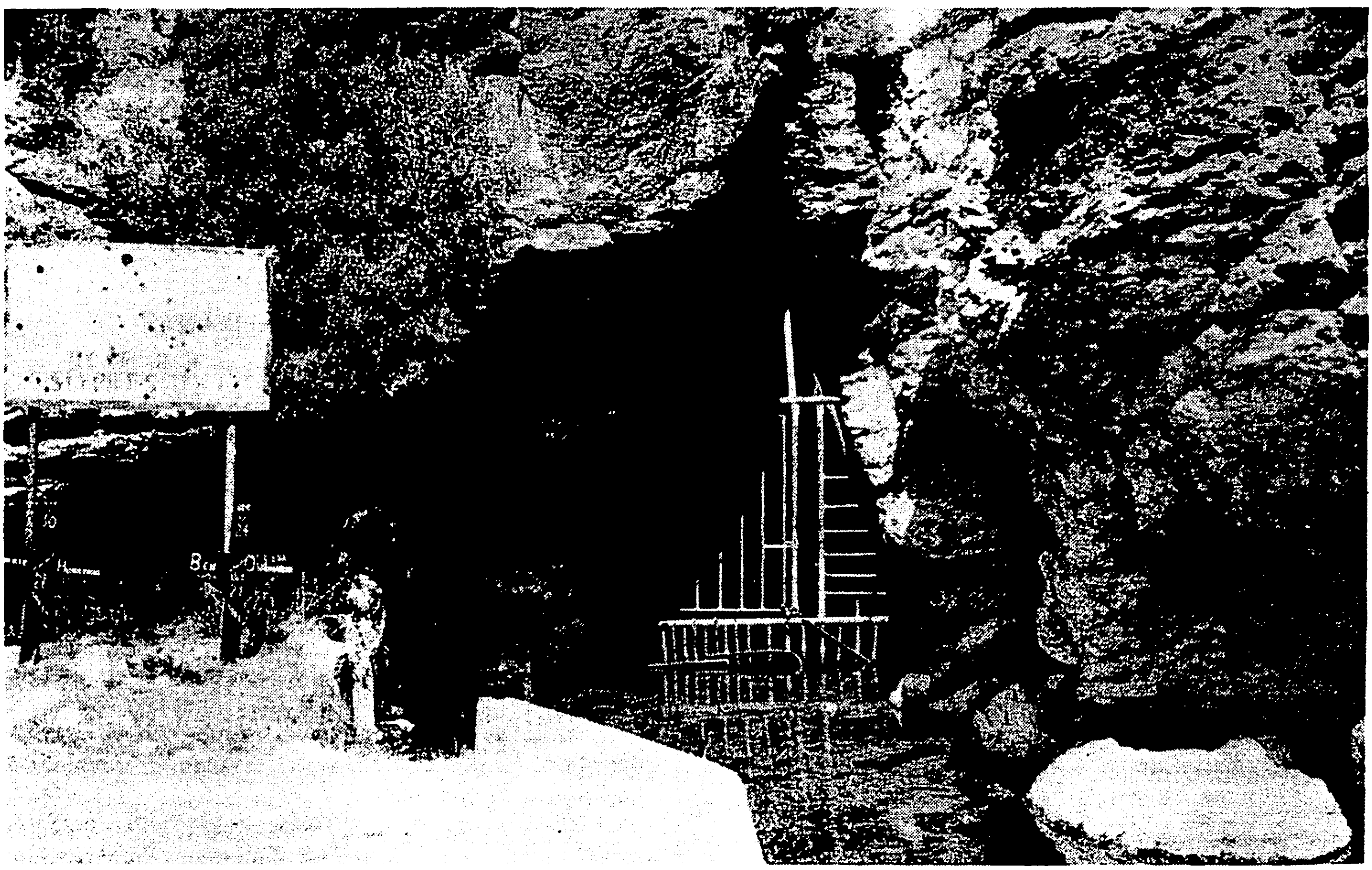

Figure 2. Phantom Lake Springs.

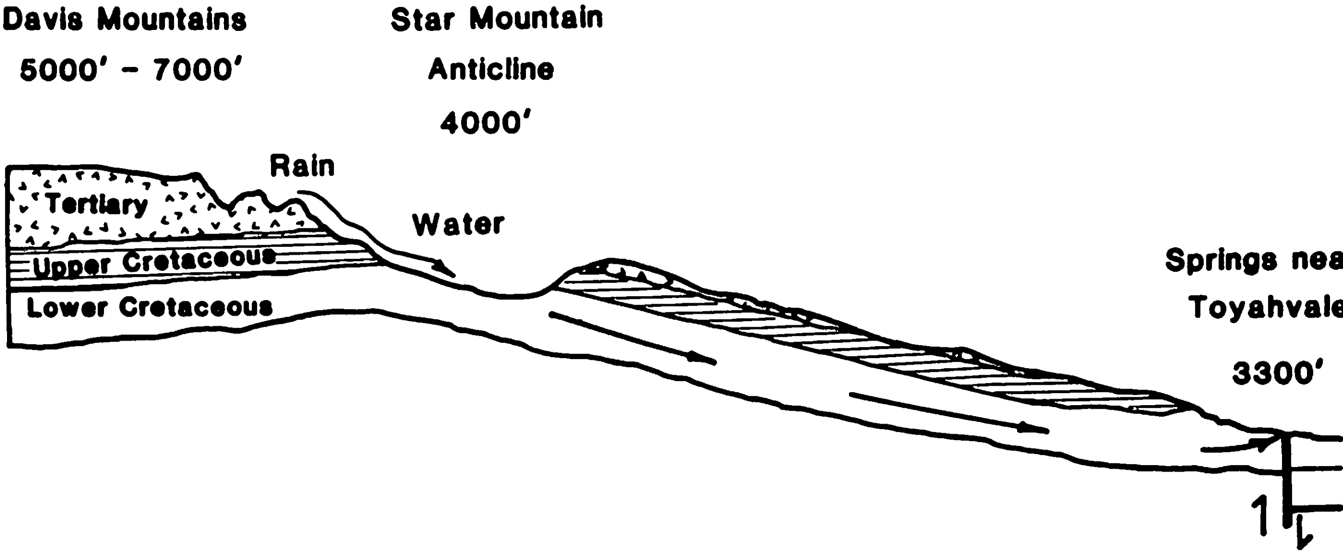


**Figure 3. Explanation of spring flow originating from the infiltration of surface waters from the Davis Mountains (after Pearson, 1985).**

which drain the Davis Mountains is markedly dissimilar from the chemical quality of the water discharging from the springs. Ground water from the Davis Mountains is quite fresh, typically containing less than 500 mg/1 total dissolved solids (TDS), while the water discharging from the springs ranges from 2000 to 2400 mg/1 TDS (Table 1). Figure 6 depicts water quality data from wells in both the Cretaceous limestones and the Tertiary volcanic rocks in Jeff Davis County. The differences in water chemistry cannot be accounted for by dissolution of the host rock as the water moves from the Davis Mountains to the springs. The lithologies encountered along this flow path include rhyolite in the Davis Mountains, gravel derived from rhyolite in the creek beds, limestone of the Lower Cretaceous unit, and possibly marl and clay in the Upper Cretaceous unit. The relative abundance of sodium, chloride, and sulfate would suggest the interaction of the water with evaporites (Hem, 1985). There are no known evaporite sequences between the Davis Mountains and the springs, although it must be noted that the carbonates beneath the volcanic pile are poorly known.

The springs also have a unique isotopic signature, which differs from Davis Mountain waters. The mean $\delta^{18}O$ value of the springs during steady flow is $-8.52$ ‰. A single $\delta^{18}O$ value obtained from a spring on Limpia Creek in the Davis Mountains was $-5.71$ ‰ (Table 2). This discrepancy also implies that the Balmorhea Springs receive some of their water from a source other than the Davis Mountains. Normal water/rock interactions, which could occur in the subsequent passage of ground water through Cretaceous limestones would tend to enrich, not deplete, the ground water in $\delta^{18}O$. The $\delta^{18}O$ values of the springs are closer to those of ground water in the Toyah Basin aquifer (LaFave, 1987). This interpretation is considerably strengthened by the $^{14}C$ analysis of Phantom Lake spring flow. A single sample, collected in the summer of 1986, gave an age (corrected) of 8954 ± 235 years before present. This argues strongly against a local origin for the spring flow and suggests that the light $\delta^{18}O$ values may owe their origin to recharge at higher elevations or during a cooler Pleistocene climate.

Only one set of trace-element analyses is now available to compare Davis Mountain and Balmorhea Spring waters. These data are shown in Table 3. It must be noted that the San Solomon Spring sample was taken from the pool (not drained at the time) and is probably diluted with flashy flow or rainwater. The bicarbonate content, for instance, appears to have reequilibrated to near atmospheric conditions. Nevertheless, these data indicate different trace metal contents. This again implies a non-Davis Mountain source of spring flow. The Limpia Creek sample, although much less saline, has greater vanadium, iron, manganese, and barium concentrations.

## THE CAPITAN REEF AS A POTENTIAL SOURCE OF SPRING FLOW

Water analyses from the Permian Capitan reef in the Apache Mountains and water analyses from the springs are similar; both the relative and absolute quantities of dissolved constituents are nearly identical (compare Figure 7 with Figures 4 and 5). Sodium, chloride and sulfate are the predominant ions, and TDS for Capitan reef water and the springs range from 2000 to 2400 mg/l. This striking similarity in water chemistries suggests that the Capitan reef and the Lower Cretaceous limestone in the vicinity of the springs are hydrologically connected.

Head data from wells in the Capitan reef indicate that the springs are downgradient from the Apache Mountains. Water level measurements (TNRIS data) in the Apache Mountains show the water level in the reef is above 3500 ft (1067 m). Although this gradient is not large, these Permian limestones are famous for their cavern systems. Carlsbad Caverns is, of course, the prime example and Phantom Lake Springs issue directly from a cave in the Cretaceous limestones. Interbasin ground water flow through karstic limestones is not uncommon in the semiarid southwestern United States (Mifflin, 1968).

The Apache Mountains are bounded to the west by the

*Accurate gaging records are not generally available for the other four springs. We suggest that continuous gaging should be reinstalled for all six springs.

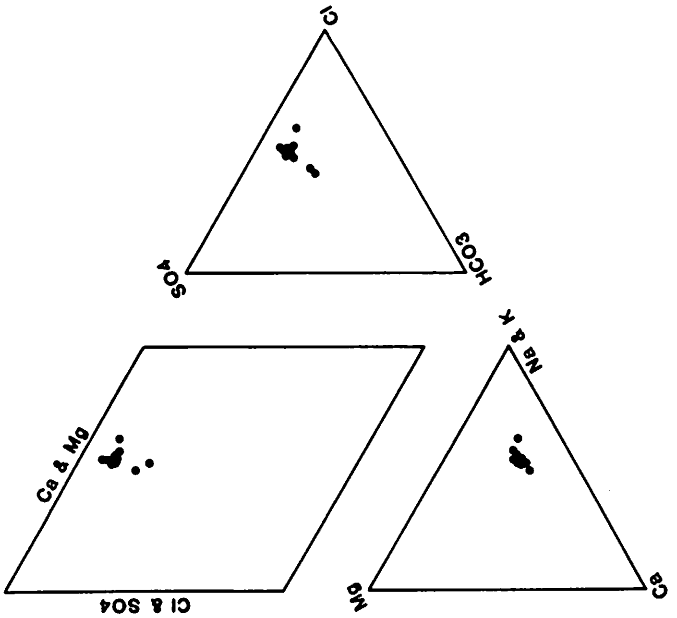


**Figure 4. Piper diagram of Phantom Lake and San Solomon Springs water chemistry.**

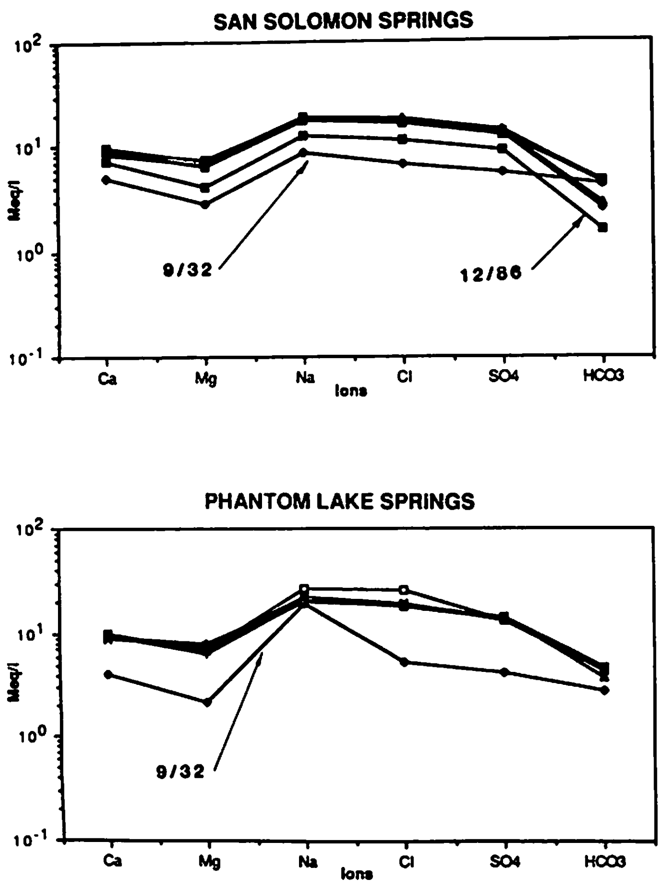


**Figure 5. Schoeller plots of water analyses from San Solomon and Phantom Lake Springs. Note the anomalously low values for the 12/86 and 9/32 analyses which were taken during a period of high discharge.**

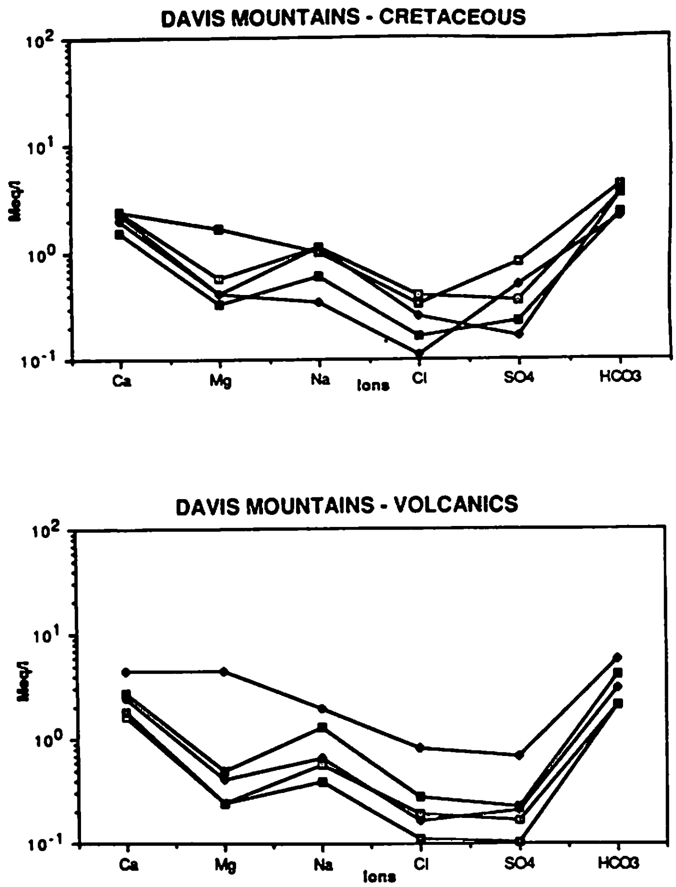


**Figure 6. Schoeller plot for water analyses from wells in Cretaceous limestone and Tertiary volcanic units of the Davis Mountains, south and west of Balmorhea.**

Salt Basin which extends north into New Mexico (Figure 8). The Salt Basin can be divided into three separate ground-water systems: northern, central and southern (Nielson and Sharp, 1985). The northern and central sections are closed hydraulically; there is no through-flowing surface drainage; each is bounded by a ground-water divide; and in the center of each are extensive salt flat deposits, (Boyd, 1982; Nielson and Sharp, 1985). The ground-water system is recharged by precipitation in the adjacent highlands of the Guadalupe, Delaware, and Sierra Diablo Mountains. Water infiltrates Permian limestones and alluvial fans which flank the mountains and then flows toward the centers of these basin sections. Discharge is by evaporation at the salt flats (Boyd, 1982). In contrast, the southernmost section lacks salt flat deposits and regional flow, as indicated by the potentiometric surface, is eastward suggesting the possibility of interbasin flow through the Apache Mountains (Nielson and Sharp, 1985). A regional, eastward flow system in these Permian limestones has also been suggested by Hiss (1980) and Mazzullo (1986).

Nielson and Sharp (1985) noted that highly permeable Capitan reef and associated limestones provide a conduit for flow through the Apache Mountains, but they did not identify a discharge site. Based on water chemistry, hydraulic heads, and the structural controls, the springs near Balmorhea are a possible discharge site for water moving eastward from the southern section of the Salt Basin through the Capitan reef

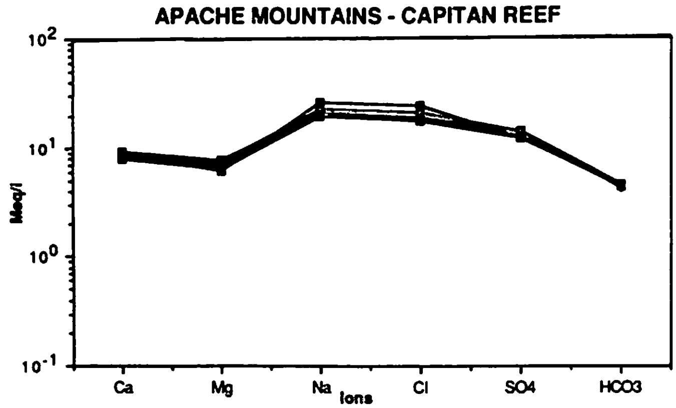


**Figure 7. Schoeller plot for water analyses from Capitan reef rocks in the Apache Mountains.**

in the Apache Mountains to the Lower Cretaceous limestones. This could account for the warm, steady mineralized flow of the springs.

This direction of regional flow is also hinted at by the alignment of faults in the Apache Mountains. Regional flow in many limestone aquifers (for instance, the Edwards) is nearly parallel to the fault trends which also parallels the direction of greatest permeability. Faults in this area appear to play two major roles: 1) they place water-bearing formations of different geologic ages in contact, and 2) they create a zone of relatively higher permeability in the limestones. The Stocks Fault which forms the northern escarpment of the Apache Mountains also juxtaposes the Capitan reef and the Lower Cretaceous limestone (Wood, 1965). The Stocks Fault extends from the Apache Mountains southeastward into Jeff Davis and Reeves Counties parallel to the front of the Davis Mountains. Throw along the fault ranges from 500 to over 1100 ft (Brand and DeFord, 1962). Wood (1962), Brand and DeFord (1962), and the Bureau of Economic Geology (B.E.G.) (1976, 1979, 1982, 1983) mapped many steeply dipping, southeast-trending, normal faults between the Apache Mountains and the northwestern flank of the Davis Mountains (Figure 9). These faults may provide a zone of high secondary permeability in the limestones which facilitates movement of the water from the Capitan reef through the Lower Cretaceous towards the springs. Because the springs are situated in the structurally low part of the Rounsaville Syncline, they present a likely discharge site for ground water moving through this structure.

## WATER BUDGET CALCULATIONS

In order to account for all the water issuing from the springs, 7% of the total volume of precipitation falling on the Davis Mountains watersheds which flow into the Toyah Basin must reach the springs. This is a very high recharge rate for a semiarid area. A general estimate is between 1 and 2% (Maxey, 1968; Maxey and Eakin, 1949). The high value (7%) also seems unlikely because much of the ground water infiltrating in the Davis Mountains will not infiltrate the Cretaceous limestones, but instead flows basinward where much will evaporate or infiltrate into the alluvium. Therefore, because of the discrepancies in water chemistry, isotopic composition, and the water balance, the springs apparently receive a significant portion of their flow from another source. The computer simulation results presented by Nielson and Sharp (1985) calculated conservatively an annual flow out of the southern portion of the Salt Graben of about 2850 acre-feet/year. While this is much less than the average annual spring flow of 33,000 acre-feet/year, LaFave (1987) suggests that the Apache Mountains also contribute ground water to the Toyah Basin. These data suggest that additional groundwater recharge is added to the regional flow system in the Apache Mountains and that a significant portion of the spring flow is produced from a regional carbonate aquifer.

The Davis Mountains are the probable source of the flashy, "storm" component of spring flow. We propose the following recharge model based upon our data and the ideas of White and others (1941) and Pearson (1985). After heavy rainfalls, water infiltrates into the volcanic rock and moves downward through fractures until it reaches low-impermeability layers of tuff or underlying Upper Cretaceous strata. It then flows laterally and appears as springs or seeps in the mountain canyons. From here the water flows down the creeks or infiltrates the creek gravels and moves downstream. This water, as well as precipitation on the outcrop, infiltrates the Lower Cretaceous limestone along the southwest limb of the Rounsaville syncline, where it moves rapidly down dip through the fractures and solution-enlarged cavities of the limestone and discharges at the springs. Therefore, after heavy rains, water from the Davis Mountains is flushed through the springs and mixes with the steady flow components from the Capitan reef of the Apache Mountains. This could account for 1) the increase in discharge, 2) the decrease in salinity, 3) the decrease in temperature, and 4) the increase in turbidity observed in spring flow.

To test this hypothesis, we employed the geochemical reaction model PHREEQE (Parkhurst and others, 1980) to simulate the mixing of Davis Mountains water with the steady spring flow water. Based on an ion-pairing aqueous model, PHREEQE calculates solution speciation and saturation states of an aqueous solution, as well as simulating several types of reactions including 1) addition of reactants to a solution, 2) mixing of two waters, and 3) titrating one solution with another. In order to 'mix' two waters the user must specify the composition of the two initial solutions and the relative proportion in which they are to be mixed. Fortunately, San Solomon and Phantom Lake Springs have been continuously gaged for several periods. White and others (1941) documented a heavy rainfall near the springs and recorded the changes in discharge and water quality; data from that report were used to simulate the mixing. We'd like to note that this event is an exception to the general steady flow of the springs. The steadiness of the flow also implies discharge from a large (regional) aquifer.

From the end of August through September 1932, a record rainfall of 14.36 in (36.5 cm) was recorded in the Balmorhea district. The first heavy rain totaling 2.72 in (7.0 cm) fell on August 30th. White and others (1941) reported that the discharge of Phantom Lake Spring, which had been at a constant 12 to 13 ft³/sec (0.4 m³/s) for months, began to increase almost immediately and reached 46 ft³/sec (1.3 m³/s) by September 11. San Solomon Spring, which had been flowing at

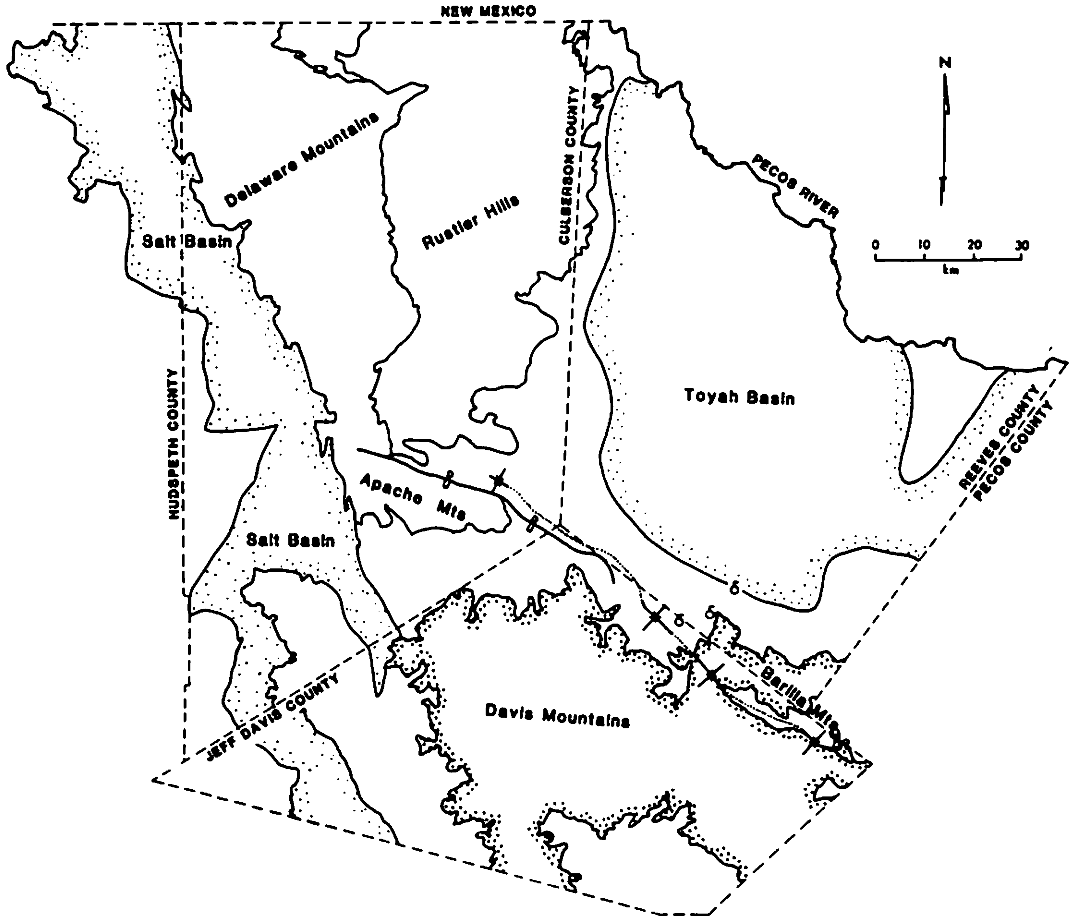


Figure 8. The Salt Basin, the Apache Mountains, and the Toyah Basin. Regional flow analyses indicate regional ground-water flow from west to east through Permian limestones. Note the location of the Stocks Fault and the Rounsaville Syncline.

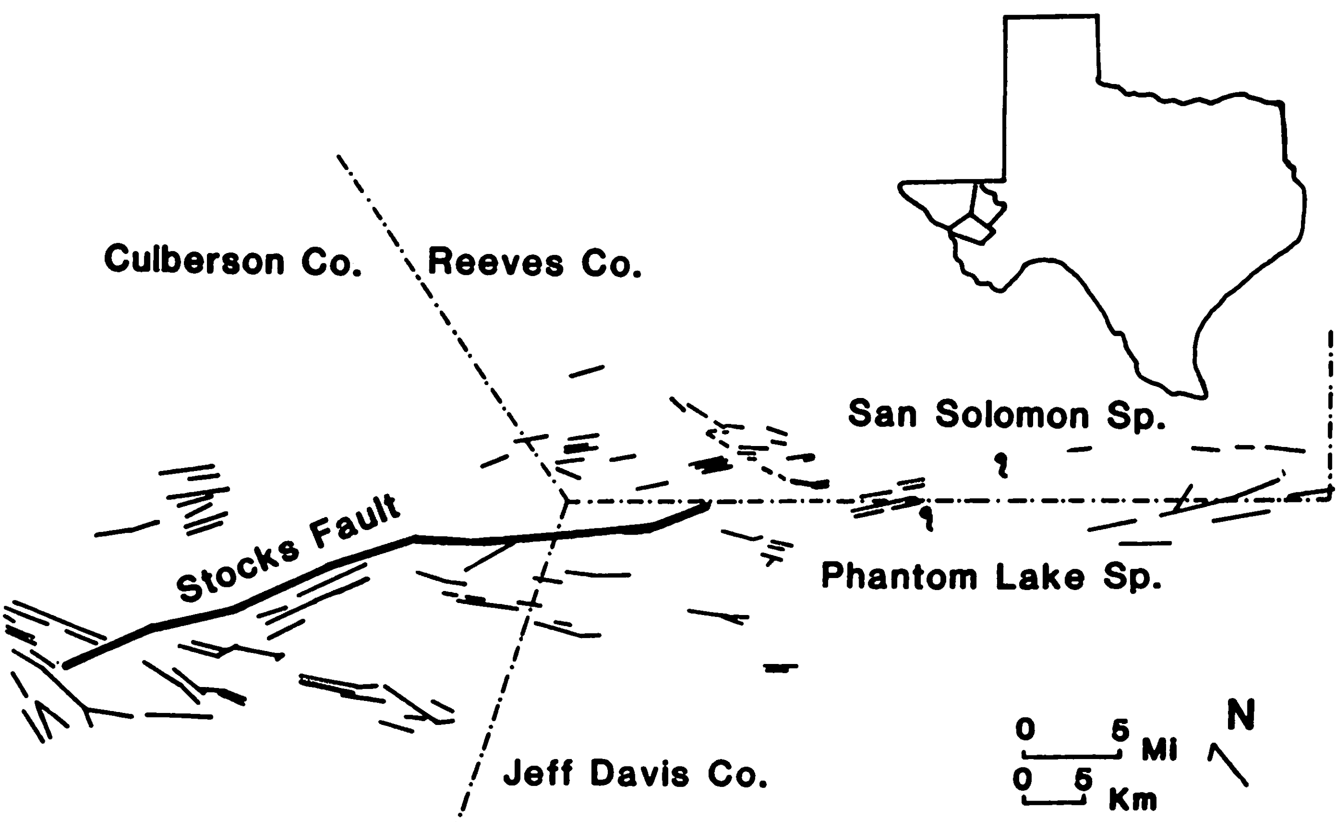


Figure 9. Trend of faults between the Apache Mountains and the northwestern flank of Davis Mountains. Note the location of the Stocks Fault.

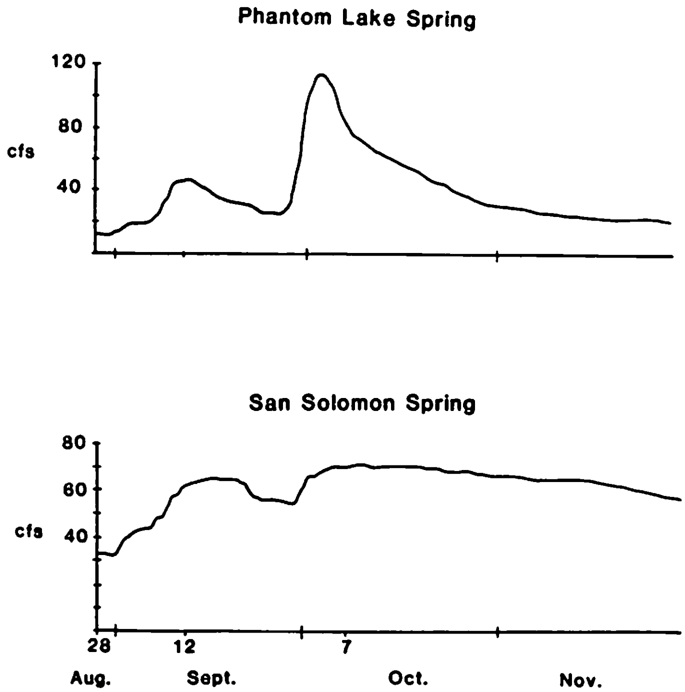


**Figure 10. Hydrographs for San Solomon and Phantom Lake Springs for September and October, 1932. Data are from White and others (1941).**

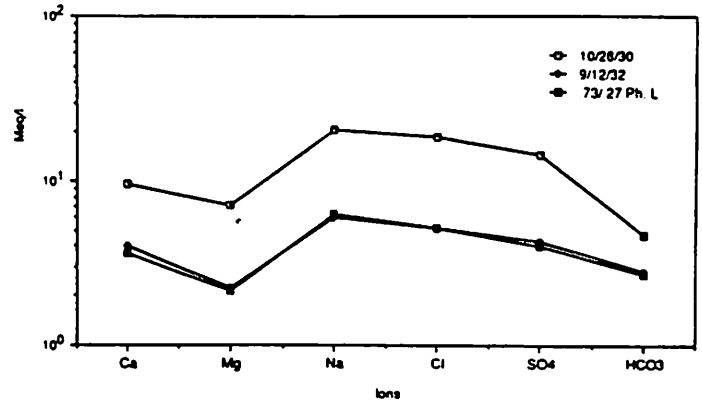


**Figure 11. PHREEQE simulation results for Phantom Lake Spring. Semi-log Schoeller plot of 1) a "normal" Phantom Lake Spring analysis (10/28/30), 2) a storm-flow analysis (9/12/32), and 3) a PHREEQE simulated mix of 73% Davis Mountains water with 27% normal spring water.**

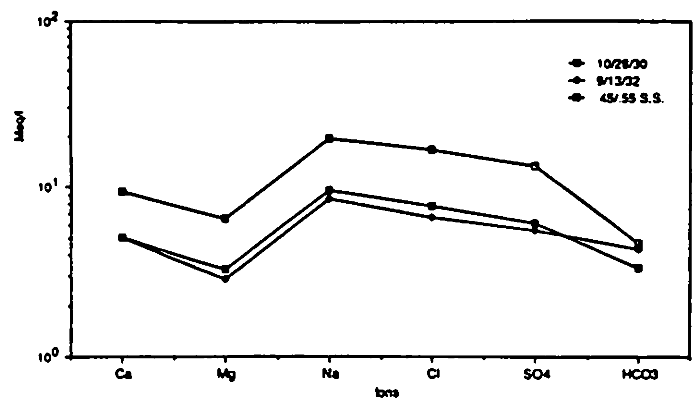


**Figure 12. PHREEQE simulation results for San Solomon Spring. Semi-log Schoeller plot of 1) a "normal" San Solomon Spring analysis (10/28/30), 2) a storm-flow analysis (9/13/32), and 3) a PHREEQE simulated mix of 55% Davis Mountains water with 45% normal spring water.**

a constant 32 to 33 $ft^3$/sec (0.9 $m^3$/s) for several months, increased at a slightly slower rate but reached a discharge of 65 $ft^3$/sec (1.8 $m^3$/s) by September 14 (Figure 10). Water analyses were collected from Phantom Lake and San Solomon Springs during these periods of high discharge. Using typical analyses for Davis Mountains water (Limpia Creek analysis, Table 1) and the steady flow component of the springs, the result of mixing these two solutions was compared to the actual composition of the spring water at high discharge. In order to determine the relative proportion in which the two solutions were mixed it was assumed that 12 $ft^3$/sec (0.4 $m^3$/s) for Phantom Lake Spring and 33 $ft^3$/sec (0.9 $m^3$/s) for San Solomon Spring were derived from the steady source. Any increase in discharge above these levels is attributed to the Davis Mountains storm flow. For example, on September 12, 1932, when the analysis was taken for Phantom Lake Spring, the discharge was 45 $ft^3$/sec (1.3 $m^3$/s), hence 27% was attributed to the steady source and 73% to the Davis Mountains. Mixing was simulated for Phantom Lake and San Solomon Springs. Note that we are not implying that the "storm" component is all rainwater. Standard analyses (e.g., Sklash and Farvolden, 1979) usually show that the infiltration of surface waters increases the hydraulic gradient and thus increases the discharge of ground water already in the system.

PHREEQE results show a high correlation between the predicted and the actual composition of the springs during storm flow for Phantom Lake Spring (Figure 11) and fair correlation for San Solomon Spring (Figure 12). The results of this mixing model imply that the flashy component is derived from the Davis Mountains and mixes with the steady flow component of the springs. Thus, this may account for the observed changes in flow characteristics during periods of high discharge.

## DISCUSSION

Analysis of water quality data demonstrates that the steady flow component of the springs at Balmorhea is chemically very similar to waters from the Permian (Capitan reef) limestones in the Apache Mountains to the west. Water quality data, including preliminary isotopic and trace-element data, indicate that spring waters are not similar to waters of the Davis Mountains, whether in the Cretaceous limestones or Tertiary volcanics.

These findings, coupled with chemical mass balance modeling, indicate that there are two recharge mechanisms for the springs, a steady flow component and a flashy "storm" component. A significant portion of the steady flow is derived from water in the southern Salt Basin and the Apache Mountains which flows through the highly permeable Capitan reef into the Lower Cretaceous limestone and discharges at the springs. This water is warm and mineralized. The "storm" flow is derived from local precipitation on the volcanic rocks of the Davis Mountains and on outcrop of the Lower Cretaceous limestone. The water moves rapidly through fractures and solution cavities and discharges at the springs. This water is cooler and less saline.

Uncertainties still exist in the determination of the ultimate source of the spring flows at Balmorhea. For instance, how much water from the Davis Mountains really infiltrates and recharges the Cretaceous limestone? Why is the $^{18}O$ ratio of the spring waters so light? How much ground water can be pumped from local aquifers without diminishing spring flow? Our model of the springs discharging a regional carbonate aquifer, if correct, indicates that increased pumping of ground water in the Davis Mountains will have little effect on spring flow. Some of these questions can be addressed by further geochemical analyses, including isotopes and trace elements, which are now being conducted by The University of Texas. Others need detailed water budget analyses, which will require establishment of stream gaging stations and delineation of aquifer heads and hydraulic properties. Unless we wish on the springs of Balmorhea the fate of many other important Texas springs, the sooner these studies commence, the better.

## ACKNOWLEDGMENTS

Karl Hoops conducted the trace-element analysis. $\delta^{18}O$ analyses were performed by Lynton Land and Amy Wilkerson. $^{14}C$ analysis was conducted by The University of Texas Radiocarbon Laboratory under a grant from Sigma Xi. Cindy Fong did the drafting. Rosemary Brant assisted in the editing. Manuscript preparation was funded by the Owen-Coates Fund of the University of Texas Geology Foundation.

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**Table 1.** Water Chemistry Data
(values in mg/l)

| Well# | Depth | Date | Ca | Mg | Na | K | $HCO_3$ | $SO_4$ | Cl | TDS | pH |
|---|---|---|---|---|---|---|---|---|---|---|---|
| CAPITAN REEF ANALYSES: | | | | | | | | | | | |
| 47-52-301 | 1722 | 9/11/70 | 181 | 94 | 478 | - | 281 | 690 | 670 | 2394 | 7.6 |
| 47-52-602 | 1560 | 11/11/70 | 176 | 88 | 478 | - | 272 | 690 | 650 | 2354 | 7.7 |
| 47-60-603 | 500 | 2/25/71 | 169 | 82 | 478 | - | 285 | 620 | 670 | 2304 | 7.7 |
| 47-61-401 | 577 | 12/18/70 | 167 | 91 | 530 | 26 | 262 | 670 | 740 | 2486 | 8.1 |
| 47-61-401 | 577 | 3/17/71 | 161 | 79 | 447 | - | 275 | 600 | 630 | 2192 | 7.5 |
| 51-04-301 | - | 2/24/71 | 177 | 76 | 600 | - | 278 | 600 | 870 | 2601 | 8.3 |
| 52-02-202 | 500 | 2/18/71 | 177 | 76 | 600 | - | 278 | 600 | 870 | 201 | 7.8 |
| PHANTOM LAKE SPRING: | | | | | | | | | | | |
| | | 10/28/30 | 191 | 86 | 453 | 20 | 285 | 691 | 655 | 2381 | - |
| | | 9/22/32 | 81 | 27 | 139 | - | 170 | 205 | 186 | 808 | - |
| | | 10/27/40 | 182 | 87 | 468 | - | 283 | 672 | 636 | 2328 | - |
| | | 1/14/41 | 193 | 86 | 470 | - | 282 | 687 | 646 | 2364 | - |
| | | 1/30/41 | 193 | 87 | 458 | 30 | 282 | 689 | 646 | 2385 | - |
| | | 3/14/41 | 190 | 83 | 614 | - | 269 | 633 | 900 | 2689 | - |
| | | 10/15/47 | 191 | 90 | 451 | - | 230 | 692 | 650 | 2304 | - |
| | | 1/28/50 | 187 | 95 | 473 | - | 282 | 695 | 660 | 2392 | 7.4 |
| | | 2/15/67 | 190 | 83 | 470 | 21 | 280 | 684 | 655 | 2383 | 7.5 |
| | | 3/22/67 | 188 | 78 | 470 | 20 | 284 | 696 | 650 | 2386 | 7.2 |
| | | 9/14/67 | 198 | 82 | 465 | 21 | 284 | 688 | 660 | 2398 | 7.6 |
| | | 7/22/85 | 177 | 94 | 508 | 20 | 286 | 625 | 674 | 2384 | 7.0 |
| SAN SOLOMON SPRING: | | | | | | | | | | | |
| | | 10/28/30 | 190 | 80 | 432 | 16 | 286 | 651 | 610 | 2265 | - |
| | | 9/13/32 | 102 | 35 | 200 | - | 264 | 270 | 238 | 1109 | - |
| | | 1/30/41 | 184 | 78 | 405 | 18 | 284 | 612 | 570 | 2151 | - |
| | | 10/15/47 | 168 | 81 | 402 | - | 156 | 638 | 590 | 2035 | - |
| | | 1/28/50 | 179 | 90 | 421 | - | 273 | 635 | 600 | 2198 | - |
| | | 6/10/69 | 192 | 87 | 429 | - | 282 | 650 | 600 | 2240 | - |
| | | 6/15/70 | 197 | 87 | 434 | - | 279 | 680 | 640 | 2317 | - |
| | | 3/20/86 | 181 | 80 | 417 | 18 | 177 | 650 | 625 | 2157 | 7.1 |
| | | 12/9/86[1] | 148 | 50 | 286 | 15 | 99[2] | 430 | 400 | 1427 | - |
| DAVIS MOUNTAINS ANALYSES: | | | | | | | | | | | |
| Spring, Limpia Creek | | 2/17/85 | 47 | 5 | 26 | - | 205[2] | 8 | 9 | 300 | 7.9 |
| Limpia Creek | | 12/9/86 | 31 | 4 | 14 | 3 | 135[2] | 11 | 6 | 205 | - |
| Volcanics: | | | | | | | | | | | |
| 52-09-501 | 428 | 5/27/69 | 33 | 3 | 13 | - | 129 | 8 | 7 | 193 | 7.3 |
| 52-10-202 | - | 5/21/69 | 48 | 5 | 15 | - | 183 | 10 | 6 | 267 | 7.4 |
| 52-11-501 | - | 7/30/70 | 36 | 3 | 9 | - | 128 | 5 | 4 | 185 | 7.5 |
| 52-25-302 | 152 | 6/28/79 | 90 | 17 | 45 | - | 352 | 33 | 29 | 593[3] | 7.9 |
| 52-25-305 | 303 | 6/25/85 | 54 | 6 | 30 | 0.3 | 246 | 11 | 10 | 357 | 8.0 |
| Cretaceous limestones: | | | | | | | | | | | |
| 51-23-103 | 420 | 6/25/85 | 49 | 7 | 26 | 0.2 | 201 | 17 | 14 | 314 | 8.5 |
| 52-01-401 | 350 | 5/4/73 | 40 | 5 | 8 | - | 126 | 24 | 4 | 207 | 7.9 |
| 52-01-902 | 496 | 10/18/69 | 49 | 20 | 23 | - | 245 | 39 | 12 | 388 | 7.7 |

[1] Sample probably diluted with precipitation; some equilibration with atmospheric concentrations of $CO_2$ has occurred.
[2] Bicarbonates determined by mass balance, not field-determined.
[3] Included 27.3 mg/l nitrate.

**Table 2.** Isotopic Analyses

| Location | Date (all 1985) | $\delta^{18}O$ |
|---|---|---|
| **Balmorhea Springs:** | | |
| San Solomon | February 17 | $-8.20 \pm 0.09$ |
| Phantom Lake | March 20 | $-8.53 \pm 0.2$ |
| Phantom Lake | July 22 | $-8.50 \pm 0.06$ |
| Giffin Spring | March 20 | $-8.56 \pm 0.08$ |
| Giffin Spring | June 26 | $-8.62 \pm 0.14$ |
| West Sandia Spring | March 20 | $-7.96 \pm 0.09$ |
| **Davis Mountains:** | | |
| Spring on Limpia Creek | February 17 | $-5.91 \pm 0.09$ |
| **Toyah Basin wells (LaFave, 1987):** | | |
| #85-3 | July 21 | $-7.00 \pm 0.07$ |
| #85-4 | July 21 | $-7.54 \pm 0.1$ |
| #85-6 | July 21 | $-7.34 \pm 0.06$ |
| #85-8 | July 22 | $-6.72 \pm 0.16$ |
| #85-13 | July 22 | $-7.71 \pm 0.08$ |
| #85-16 | July 23 | $-7.26 \pm 0.10$ |
| #85-17 | July 23 | $-6.51 \pm 0.10$ |

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**Table 3.** Trace-Element Analysis.
All values in ppb, unless otherwise indicated.

| | Limpia Creek (12/9/86) | San Solomon Spring Pool (12/9/86) |
|---|---|---|
| Zn | 1.7 | 8 |
| Ni | 35 | 71 |
| Al | 43 | 85 |
| P | 1.1 ppm | 14 ppm |
| V | 5 | --- |
| Li | 21 | 328 |
| Fe | 50 | 19 |
| Sr | 0.2 ppm | 2.7 ppm |
| Mn | 7 | trace |
| Ba | 41 | 31 |

# TECTONIC CONTROLS ON THE HYDROGEOLOGY OF THE SALT BASIN, TRANS-PECOS TEXAS

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## INTRODUCTION

The portion of the Salt Basin that extends from Van Horn, Texas, northward to the Texas–New Mexico border is a series of north-trending half grabens (deeper to the west) which form a continuous intermontane valley 110 km long. Geological controls on the hydrogeology of the study area are the result of primarily two events. The first was the deposition of Permian shelf carbonates around the Delaware Basin. The second was late Cenozoic Basin and Range block faulting which resulted in closed basins that filled with bolson deposits and developed the present topography. Major aquifers in the valley include bolson (basin) alluvium and down-dropped blocks of permeable Permian limestone.

## AQUIFER SYSTEMS

### Permian Strata

In the northern arc of the Delaware Basin, extending into New Mexico, Permian age strata are divided into three aquifer systems: Basin, Capitan, and Shelf aquifers (Hiss, 1980). Within the study area, a similar classification applies with the Permian system divided into a low-permeability basin facies, a high-permeability shelf margin facies, and a shelf facies with permeability dependent upon fracture porosity (fig. 1).

*Basin Aquifer.* Basinal sediments of the Delaware Mountain Group lie at the eastern border of the study area (fig. 2). Because of poor water quality and low well yields, very few wells pump from these predominantly sandstone deposits. The few stock wells that have been drilled have well yields ranging from 5 to 20 gpm (0.3 to 1.2 l/sec.).

Collectively, the basin sediments of the Delaware Mountain Group, as well as the overlying evaporites of the Castile and Rustler Formations, have low to very low permeability within and directly east of the study area.

*Shelf Margin Aquifers.* Included in this group are the reef facies of the Capitan and underlying Goat Seep limestones of Guadalupian (Late Permian) age. Wells are drilled into the Capitan Reef south of Patterson Hills, where the reef has been downfaulted into the basin. Transmissivities as high as 1,500 $m^2$/day (16,000 $ft^2$/day) have been reported by Reed (1965), Gates and others (1980) estimated a mean transmissivity of 500 $m^2$/day (5,400 $ft^2$/day) based on specific capacity measurements. Based on cores, electric logs and test drilling, it appears that porosity and permeability are well developed throughout the subsurface reef (Reed, 1965). A detailed study of the hydrology of the Permian reef in the Carlsbad area revealed similar findings with the exceptionally high permeability of the Capitan attributed to the susceptibility of the microcrystalline limestone and coarsely textured carbonates of the reef zone to groundwater solution (Motts, 1968), a prime example of which are the Carlsbad Caverns.

*Shelf aquifers.* This term is applied to the earlier Leonardian limestones that do not exhibit the well-developed reef facies of the Capitan Formation. Within the study area, pumpage from these limestones occurs in the Dell City area and in the southern portion of the Salt Basin, at Wildhorse Flat.

In the Dell City area, pumpage is from the undifferentiated Bone Spring–Victorio Peak limestones. Well yields are highly dependent upon fracture porosity. Approximately 50 percent of the early wells drilled failed to yield sufficient quantities of water for irrigation (Scalapino, 1950). More recently, attempts are being made to map fracture zones to aid in well placement (Logan, 1984, personal communication).

In successful wells, drillers have encountered such cavernous zones that they could not get well cuttings from the hole. Therefore, while there are relatively large openings along joints and fractures, porosity is erratic. Based on specific capacities, average transmissivities of these wells have been reported at 930 $m^2$/day (10,000 $ft^2$/day) (Davis and Leggat, 1965) and at 3,100 $m^2$/day (33,000 $ft^2$/day) (Scalapino, 1950). At Wildhorse Flat in the southern part of the Salt Basin, wells have recently been drilled through more than 1,000 ft of bolson de-

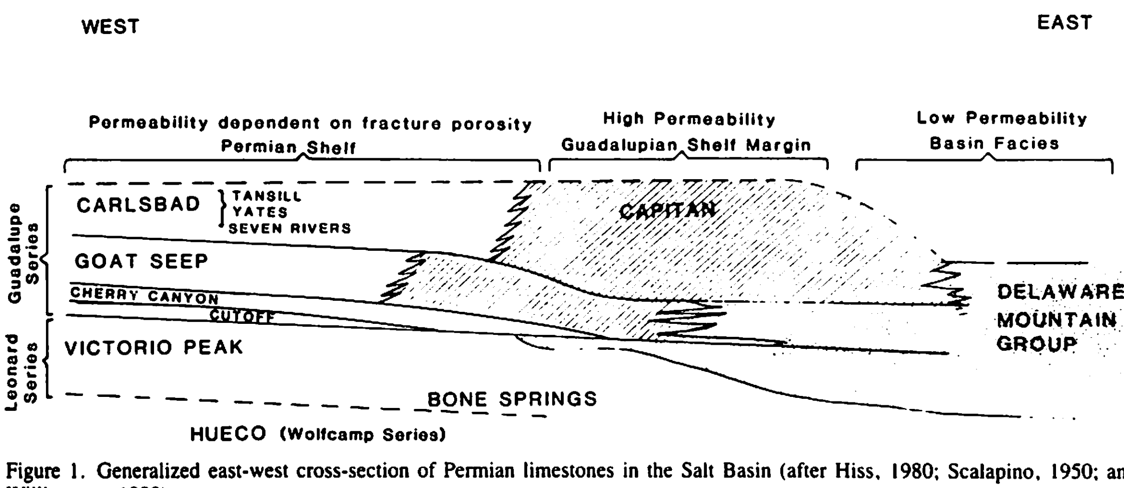


Figure 1. Generalized east-west cross-section of Permian limestones in the Salt Basin (after Hiss, 1980; Scalapino, 1950; and Williamson, 1980).

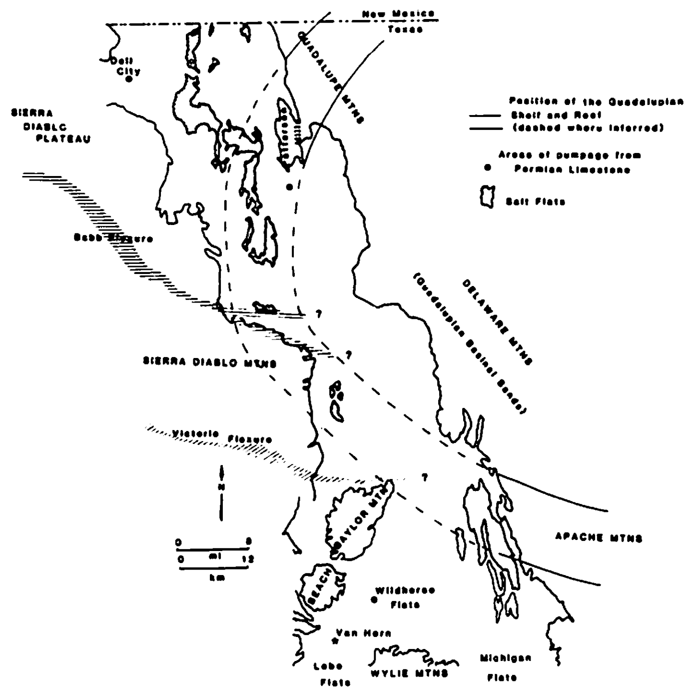


Figure 2. Salt Basin showing inferred position of high-permeability reef trend and positions of major flexures (after Goetz, 1980).

posits into limestones underlying the bolson fill. These deep, subsurface limestones, which lie midway between the Bone Spring limestones of the Baylor Mountain uplift and the Hueco limestones of the Wylie Mountain uplift, are probably either Leonardian or Wolfcampian shelf limestones.

Short-term development tests indicate a bimodal distribution of specific capacities with four tests falling in the 3 to 5 l/sec (50 to 80 gpm/ft) range and 6 tests falling within the .09 to 1.3 l/sec (15 to 20 gpm/ft) range (TDWR files). This bimodal distribution suggests that, similar to the Dell City limestones, the hydrology of the Wildhorse Flat limestones is controlled by fracture porosity.

### Bolson Aquifer

During the late Cenozoic, normal faulting produced the Salt Basin graben which extends from south of Van Horn into New Mexico. At Van Horn, the graben bifurcates to the south to form two smaller basins—Lobo Flat on the west, and Michigan Flat on the east. From Van Horn to the Texas–New Mexico border, the graben is composed of three segments, with each segment offset to the west of its adjacent southern segment along Late Paleozoic fault trends. As noted earlier, the fracturing associated with these tectonic events resulted in the development of high-porosity fracture zones in the consolidated Permian limestones.

Sediment deposition in the graben was typical of closed basins with internal drainage; coarse gravels were deposited at the base of the mountains and finer sediments were carried to the center of the basin. However, rather than sharp facies boundaries, the transition from gravels to muds is typically gradational with broad zones of interlayered gravels, sands and muds (Groat, 1972). This interlayering of gravels, sands, and muds probably imparts a strong anisotropy to the bolson deposits with vertical movement of groundwater impeded by the clay interlayers.

An additional characteristic of groundwater movement in the bolson deposits was recognized by Gates and White (1980). They determined, based on resistivity measurements, that clay content in the alluvial deposits increased northward in the Salt Basin, with corresponding decreasing permeability.

In the northern section of the Salt Basin, the only wells drilled into the alluvial fill are low-yield stock wells. In contrast, irrigation wells with yields of 25 to 75 l/sec (400 to 1,200 gal/min) have been drilled in the basin fill of Wildhorse Flat (Gates and others, 1980). Based on aquifer test results and well specific capacities, the transmissivity of the fill in Wildhorse Flat was estimated to range from 250 to 1,100 $m^2$/day (2,700 to 12,000 $ft^2$/day) with an average transmissivity of approximately 430 $m^2$/day (4,600 $ft^2$/day) (Gates and others, 1980).

## MOVEMENT OF GROUND WATER

The traditional pattern (Maxey, 1968) of groundwater flow in the Basin and Range province consists of recharge in the highlands and on the alluvial fans. Groundwater movement from these recharge zones is to the basin floors, where discharge is by evapotranspiration or baseflow to streams. In addition to localized intrabasin flow, regional (or interbasin) groundwater flow has been documented (Mifflin, 1968).

The presense of salt flats in the northern half of the basin indicates a classical interbasin flow pattern for this area. However, the lack of salt flats in the southern section of the basin (Wildhorse Flat) suggests the possibility of intrabasin flow in the south. To test this hypothesis, a potentiometric map was constructed, using water levels measured in wells by the Texas Department of Water Resources during the 1950s and 1960s. These early water levels were used to infer the pre-pumpage flow patterns.

Based on the resultant head-level map (fig. 3), the basin can be divided into three sections separated by two groundwater divides. In the northern and central sections, the expected pattern of movement from the highlands to the basin predominates. However, in the south-

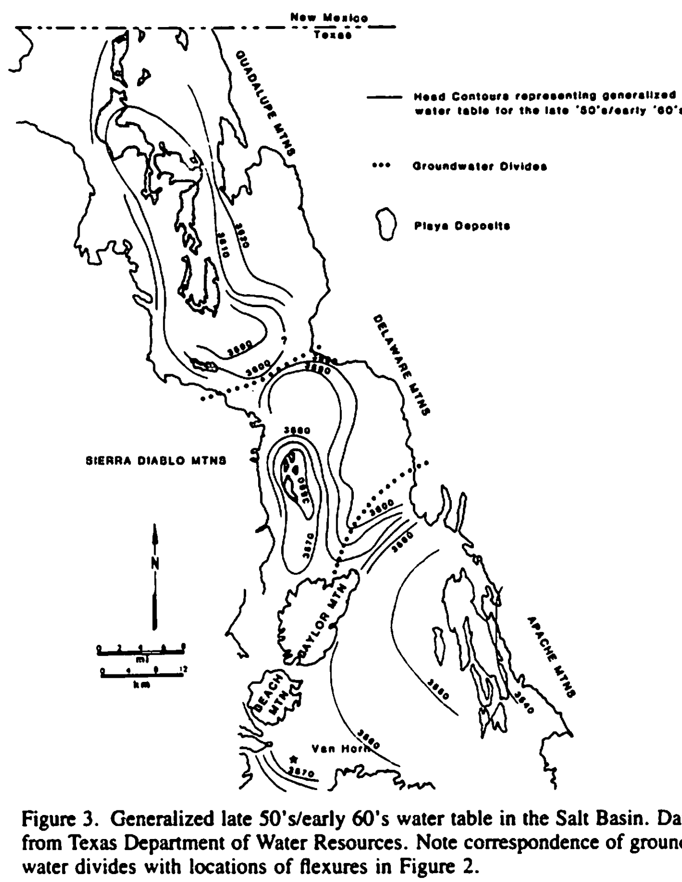


Figure 3. Generalized late 50's/early 60's water table in the Salt Basin. Data from Texas Department of Water Resources. Note correspondence of groundwater divides with locations of flexures in Figure 2.

ern section, regional flow is indicated, with the pre-pumpage flow moving eastward from the basin alluvium into the Apache Mountains.

This pattern is strongly influenced by a regional geologic setting created by Paleozoic and Cenozoic tectonic events. The present-day topography and permeability distributions within Salt Basin are the two results of tectonism that have the most influence on groundwater flow.

Late Cenozoic block faulting produced the characteristic Basin and Range topography. This topography promotes the development of local flow patterns with recharge along the flanks of the uplifted mountains and discharge on the basin floors. The influence of topography is most pronounced in the northern half of the basin where groundwater movement is from the highlands to the lowlands and where discharge is by evaporation in salt lakes.

The development of regional flow patterns in the southern portion of the basin indicates the importance of permeability distributions in defining flow patterns in the basin. Cenozoic block faulting produced a region characterized by abrupt changes in permeability between the consolidated rocks surrounding the basin and the basin fill. The most significant variations in the consolidated rocks are functions of an earlier feature—the Permian Delaware Basin.

In the northern and central sections, any eastward flow is impeded by the low-permeability sediments of the Delaware Mountain Group. To the west, flow is restricted by the low fracture porosity and corresponding low permeability of the stable Diablo Platform. As a consequence, in these sections of the basin, flow is toward the flats or playas which serve as areas of groundwater discharge.

In contrast, in the southern section of the basin there was no permeability barrier to the east. Instead, the highly permeable Permian limestones of the Apache Mountains served as a conduit for flow out of the basin as indicated on Figure 2. As a result, a regional flow pattern developed. This is further emphasized by the fact that the Baylor and Beach Mountains are topographically higher than the Apaches. The discharge area for this interbasin flow is not identified.

A final interesting point concerns the presence of the three groundwater divides. The lowest elevation in the basin is at the salt flats in the central section of the basin. In usual groundwater flow systems, flow within the basin sediments mirrors the topographic slope with the resultant flow moving towards this central section. As noted previously, the "pirating" of groundwater by the Permian limestones in the south serves to disrupt the expected flow in this section of the basin. This does not explain the northern groundwater divide, however.

Two factors could contribute to producing the northern divide. First, the low permeability of the clayey basin sediments would retard the development of long flow paths and enhance the development of short flow paths. Second, the northern divide could indicate the presence of a permeability barrier that correlates with an east-striking fault zone (Bitterwell Break) identified by Goetz (1977). A similar east-striking flexure or fault zone (Victorio Flexure) correlates with the groundwater divide at the northern end of Wildhorse Flat.

## COMPUTER MODELING OF WILDHORSE FLAT

Computer modeling of the Wildhorse Flat area was used to quantify movement in the bolson deposits and in the limestones underlying the bolson fill. Because a lack of information on the geometry and regional transmissivities made it impossible to directly calculate the relative amounts of water moving through the consolidated limestones, an indirect calculation method was used. First, the annual recharge to the flats from precipitation and underflow from adjoining areas was calculated. Secondly, the computer model was used to simulate the volume of this average annual recharge moved just through the bolson fill. The remaining surplus of average annual recharge is presumed to move through the limestones underlying the bolson fill.

Recharge to the basin occurs by infiltration of precipitation and by underflow from adjacent basins (fig. 4). Recharge by precipitation was estimated at 1 percent of the average annual precipitation of 254 cm (10 in) falling over the drainage areas of the Baylor, Beach and Wylie Mountains (Nielson, 1984). Recharge by underflow was derived from hydraulic gradients and transmissivity measurements. Total recharge from these two sources is estimated at between $2.5 \times 10^6$ – $3.5 \times 10^6$ $m^3/yr$ (1,950 and 2,850 acre-ft/yr).

To determine how much of the average annual recharge was moving through the bolson deposits, a two-dimensional finite-difference model for groundwater flow simulation was used (Trescott, Pinder, and Larson, 1976). Values for recharge, hydraulic conductivity and area are input data into the model and the model generates a simulated head distribution. This simulated head distribution can then be compared to the actual head distribution as determined by water well measurements.

The geometry of the basin was delineated by resistivity studies by White and others (1977). This information was used to determine the cross-sectional area of the bolson fill. Hydraulic conductivity (K) values, calculated by dividing well transmissivities by the portion of the total aquifer being pumped, were estimated to be between 3 and 6 m/day (10 and 20 ft/day).

By successive model runs, values of recharge were determined which resulted in a successful simulation of the observed water table. Using the higher K value of 6 m/day (20 ft/day), the simulations indicated that approximately $2 \times 10^6$ $m^3/yr$ (1,600 acre-ft/yr) of water moved through the bolson fill. However, if the lower K value of 3 m/day (10 ft/day) was used, only $10^6$ $m^3/yr$ (850 acre-ft/yr) of water moves through the bolson fill.

Assuming that average annual recharge to the basin is $2.5 \times 10^6$ $m^3/yr$ (2,200 acre-ft/yr) [Estimated range, using the standard procedures of Maxey and Eakin (1949), was $2.5 \times 10^6 - 3.5 \times 10^6$ $m^3/yr$ (1,950 to 2,850 acre-ft/yr)], the modeling indicates that only 40 to 70 percent of the average annual recharge moves through the bolson fill. The remaining 30 to 60 percent is assumed to move through the limestones underlying the bolson fill. While this is a preliminary estimate, these results indicate that the limestones underlying the basin play a significant role in the hydrology of Wildhorse Flat.

## SUMMARY

The hydrogeology of the Salt Basin is a function of large-scale permeability distributions created by Cenozoic block faulting in the graben area. In the northern sections of the basin, permeability barriers to the east and west have resulted in local flow patterns. In the south, the permeable limestones of the Apache Mountains served as a conduit for flow and resulted in the development of a regional flow system.

The importance of permeable Permian limestone to the aquifer system is not limited to the limestones adjacent to the bolson fill. In Wildhorse Flat, computer modeling suggests that as much as 60 percent of the groundwater moving through this portion of the basin could be moving through limestones underlying more than a thousand feet of bolson fill.

## ACKNOWLEDGMENTS

This research was funded by grants from Sigma Xi and the University of Texas Geology Foundation. Manuscript preparation was funded by the Owen-Coates Fund of the University of Texas Geology Foundation. Figures were drafted by Jeff Black. The counsel of Homer Logan (U.S. Soil Conservation Service), Don White (U.S. Geological Survey–El Paso), and Bill Muehlberger (University of Texas) is greatly appreciated.

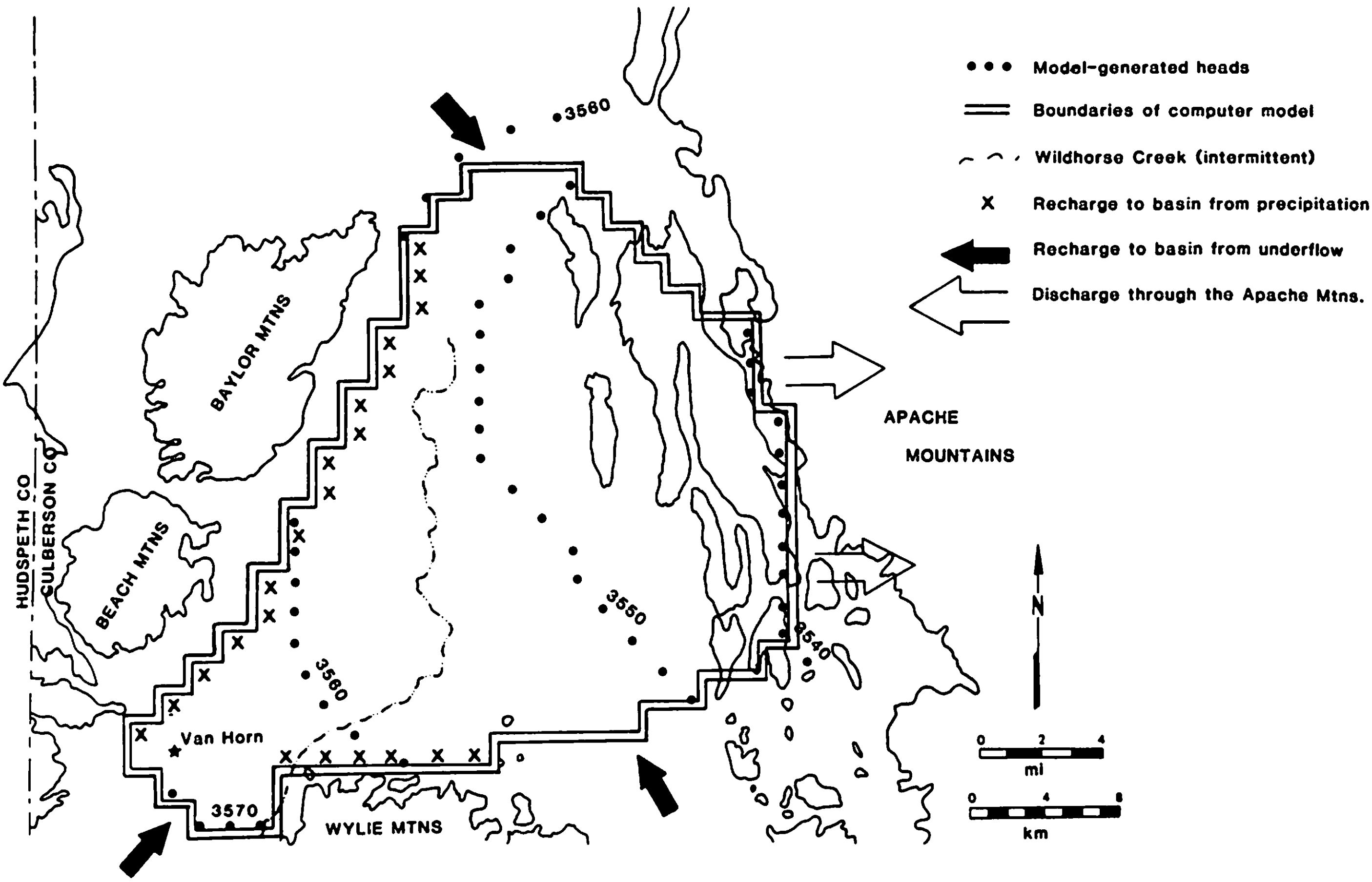


Figure 4. Computer-generated potentiometric surface of Wildhorse Flat showing zones of underflow, recharge from precipitation, and interbasin discharge.

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# ANALYSIS OF UNSATURATED FLOW RELATED TO LOW-LEVEL RADIOACTIVE WASTE DISPOSAL, CHIHUAHUAN DESERT, TEXAS

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## ABSTRACT

Site characterization studies are being conducted for low-level radioactive waste disposal in an area 65 km southeast of El Paso, Texas. To evaluate moisture movement in the thick (150 m) unsaturated section we monitored moisture content with a neutron probe and water potential with psychrometers and measured moisture content and water potential in the laboratory. Variations in water content correlate with the distribution of different lithologies. The upper sandy and gravelly loam (12 to 15 m thick) is ~ 10 to 20% saturated, whereas the deeper (135-m-thick) clays are near saturation. The lack of temporal variations in moisture content monitored in deep (41 m) access tubes indicates that recharge pulses are not moving through the system. Water potentials are low (as negative as -15.6 MPa), and gradients are positive upward, except in the shallow subsurface after rainfall events. These low water potentials indicate that water is primarily in the vapor phase, and the gradients indicate upward flow, probably controlled by evapotranspiration. Numerical simulations of steady-state upward flow match observed water potentials. Thermal gradients generally oppose water potential gradients and may result in downward movement of water vapor.

## INTRODUCTION

The unsaturated zone in West Texas is being considered as a potential repository for low-level radioactive waste (Fig. 1). Regulatory guidelines for disposal of such waste were established by the U.S. Nuclear Regulatory Commission and are outlined in 10 CFR 61.[1] According to the regulations, the waste must be isolated from the accessible environment for 500 years. Shallow burial in the unsaturated zone (< 20 m below land surface) is preferred for disposal of low-level radioactive waste.[2] Potential pathways for waste migration include (1) infiltration of water into the facility, leaching of radioactive material, and transport to the water table or (2) surface erosion and transport of radioactive material in runoff (Fig. 2). The disposal facility will be installed to a depth of approximately 12 m and will occupy an area of approximately $2 \times 10^{5}$ $m^2$. Radioactive material in solid form will be disposed of in cannisters. Although an engineered barrier will be designed to prohibit radionuclide migration from the facility, reliance will be placed on the natural system to contain the waste for 500 years.

Because water and vapor movement are considered as the primary pathways for radionuclide transport, our objective is to determine the potential for radionuclide transport by evaluating the predominant direction and rate of water and/or vapor movement in the unsaturated zone. A physical approach was used to evaluate water movement in the unsaturated zone by monitoring the moisture content and water potential. The methodology and preliminary results of the unsaturated zone studies are outlined.

## SETTING

The site, approximately 65 km southeast of El Paso, is located in the Chihuahuan Desert within the Basin and Range physiographic province (Fig. 1). Site topography is relatively flat (slopes of less than 1%), and the elevation is approximately 1300 m. Arroyos drain southwestward to the Rio Grande and are generally dry except after rainfall events. The unsaturated zone consists of 12 to 15 m of sandy and gravely loam underlain by ~ 140 m of clays with interbedded silts and sands. A discontinuous layer of caliche occurs at a depth of ~ 2 m. Surficial sediments consist of wind-blown sandy

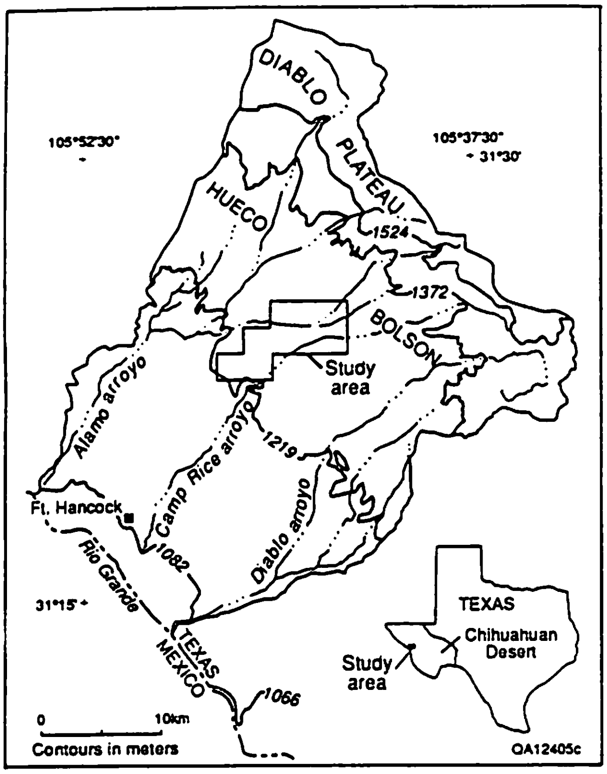


Figure 1. Location of study area.

loam in dunes and interfluves, loam in fluvial areas, and sandy gravel in upland areas (R. W. Baumgardner, pers. comm., 1989) (Fig. 3a). Extreme local variations in sediments from clay to gravel were observed in trenches dug to a depth of 7 m. Shrubs such as creosote (*Larrea tridentata*) and mesquite (*Prosopis glandulosa*) are common, with rooting depths of 1 to 5 m.

The regional climate, monitored at Fort Hancock (Fig. 1) and El Paso, is subtropical arid.[3] Average annual precipitation is approximately 280 mm, approximately 60% of which falls as local, intense, short-duration convective storms from June through September, when temperature and potential evaporation are highest. Minor winter storms are of longer duration. Mean annual class-A pan evaporation is approximately seven times mean annual precipitation. High-intensity precipitation results in surface runoff. High rates of surface runoff and evaporation may limit water movement through the unsaturated zone.

## METHODS

### Field Techniques

Soil-moisture content and water potential were used to evaluate the direction and rate of

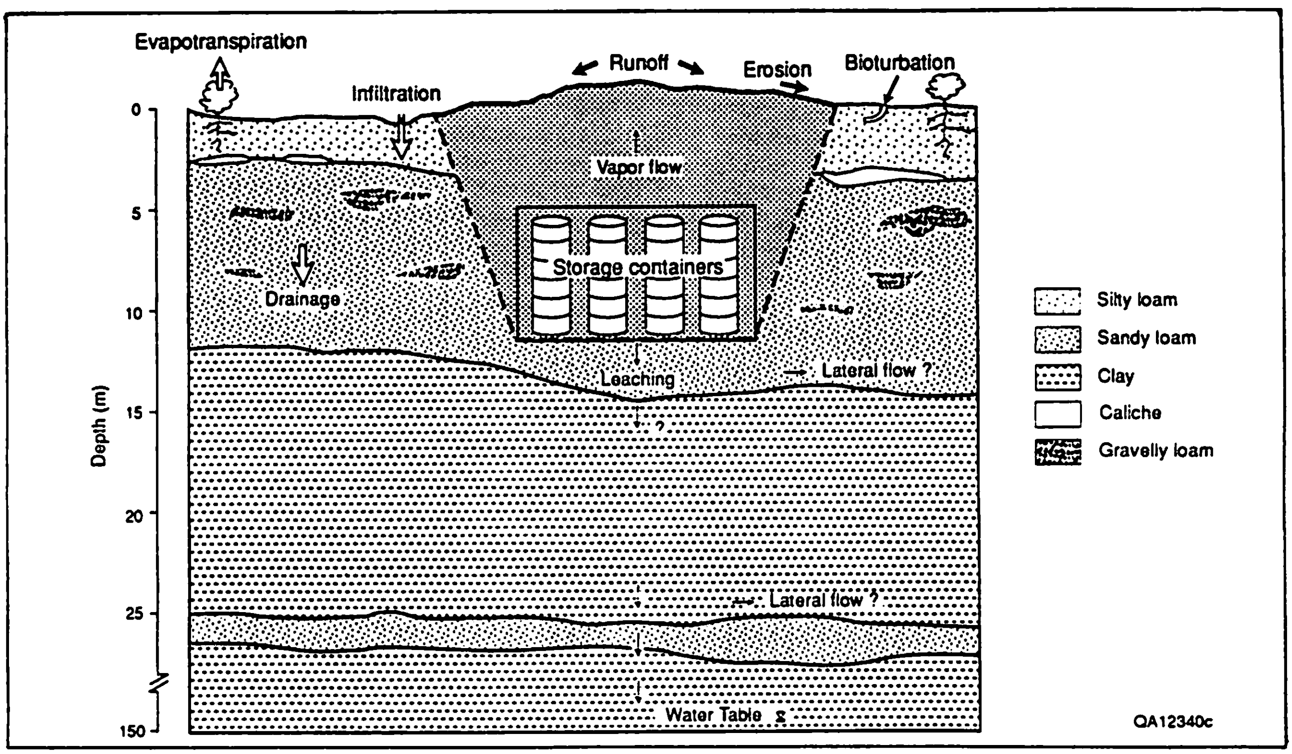


Figure 2. Schematic diagram of unsaturated section including conceptual model of potential pathways for radionuclide migration from the proposed facility.

moisture movement through the unsaturated zone. Moisture content was monitored in 5 neutron probe access tubes, and water potential was measured with 20 psychrometers. The monitoring equipment was installed in an ephemeral stream setting where recharge rates were expected to be greater than those of any other geomorphic region. In addition to the monitoring, soil samples were collected from various geomorphic settings for laboratory measurement of moisture content and water potential.

### Moisture Content

Soil samples were collected for gravimetric moisture content and calibration of the neutron probe. Volumetric-moisture and bulk-density samples were collected in the top 1.5 m after ponding a site. Neutron-probe access tubes were installed in two boreholes drilled to 21 m and 41 m depth approximately 6 m apart with an air rotary rig (Figs. 3b and 4).[4] Because of drilling difficulties, steel drill pipe (70 mm O.D., 60 mm I.D.) was used instead of conventional aluminum access tubes. The access tubing was advanced every 3 m during drilling to avoid drying the overlying material. The top 21 m of the deeper borehole was cemented for stability. Several other boreholes (numbers 61 to 63) were drilled to depths of 1.8 m with a solid-stem auger (50 mm diameter) for installation of neutron-probe access tubes. Moisture content was monitored monthly using a CPN model 503 DR neutron probe. When possible after rainfall events, moisture content was measured daily at shallow levels.

Thermocouple psychrometers were installed in the field in March 1989 (Figs. 3b and 4). To install psychrometers at shallow depths, a pit was dug to 1.4 m and psychrometers were placed in horizontal holes (13 mm diameter, 0.5 m long) drilled laterally into the pit wall. This procedure ensured that material overlying the psychrometers was undisturbed and that a good contact existed between the psychrometers and the surrounding sediments. Because the psychrometers were not retrievable, they were installed in duplicate for data verification at 0.3 m below land surface and at 0.3-m intervals between depths of 0.5 and 1.4 m. The pit was backfilled with the original sediments.

At greater depths, psychrometers were installed in duplicate in a borehole that was drilled to a depth of 14.5 m with a solid-stem auger (50 mm diameter). For protection during installation, the psychrometers were emplaced in a PVC screen filled with commercial (Ottawa) sand (0.1 to 0.4 mm grain size). Borehole cuttings below 6 m depth were used to backfill immediately around the psychrometers; however, the material from depths shallower than 6 m was considered too coarse, and commercial Ottawa sand was used. Epoxy (DER324/DEH24, Dow Chemical Company) was used to prevent preferential gas flow between psychrometer stations within the borehole and as a seal at the surface to minimize preferential flow down the borehole. The small-diameter borehole and backfill with natural materials were designed to minimize psychrometer equilibration time.

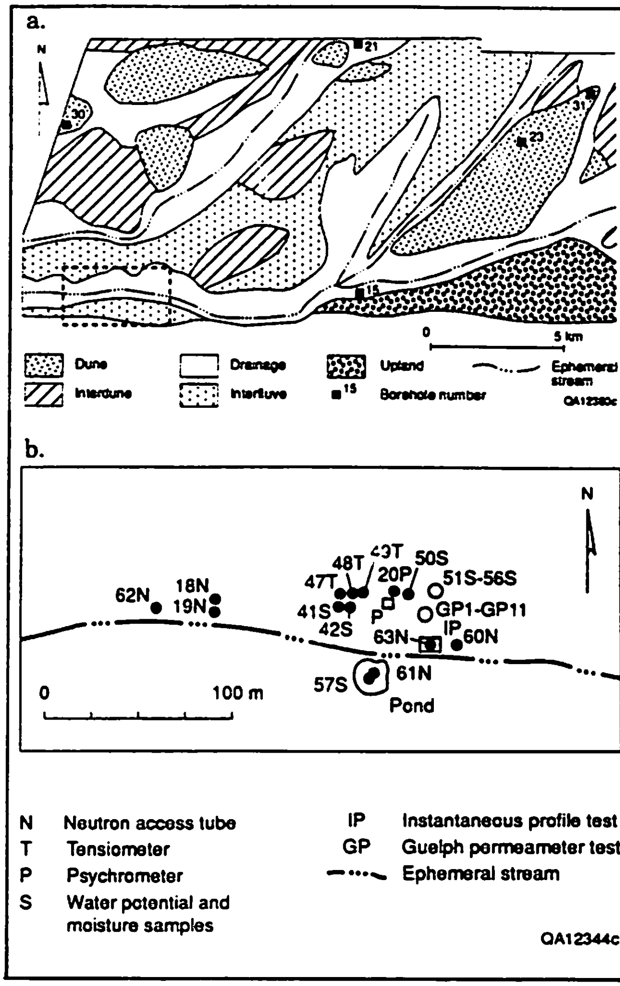


Figure 3. a. Location of boreholes sampled for moisture content and water potential in relation to surface geomorphic features (R. W. Baumgardner, pers. comm., 1989). b. Location of monitoring equipment, sampled boreholes, and hydraulic conductivity measurements.

Soil temperature was measured with thermocouple psychrometers and with thermistors, both connected to a Campbell CR7 data logger. Water potentials and temperatures were logged daily at

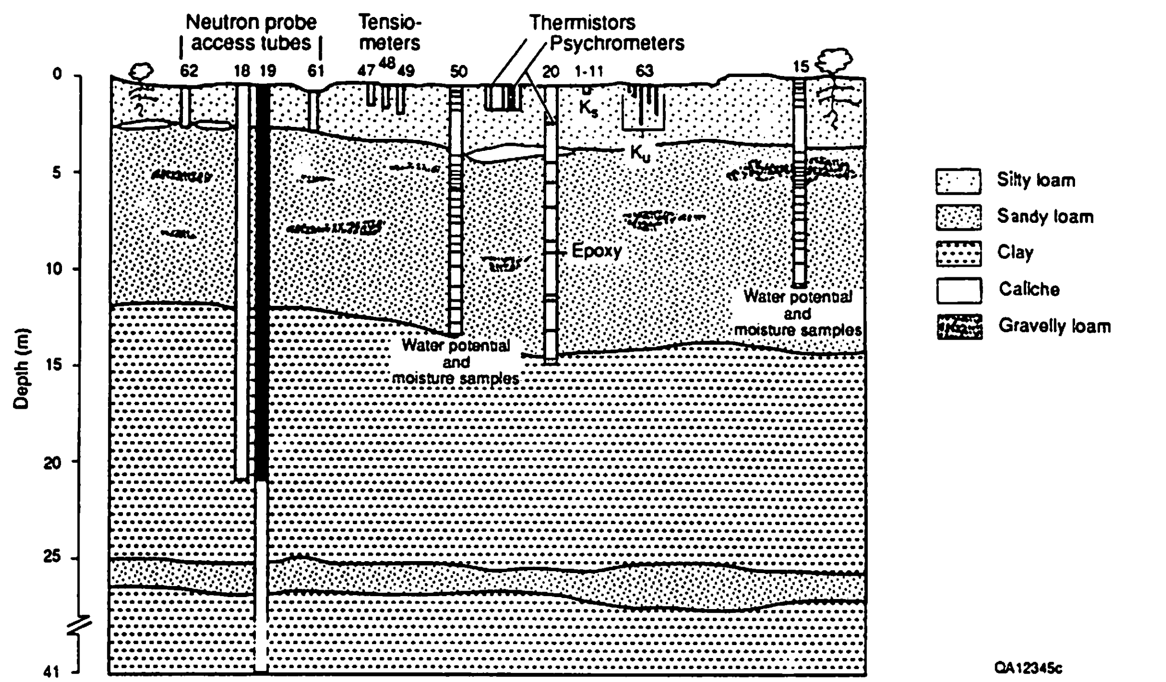


Figure 4. Schematic cross section detailing vertical distribution of monitoring equipment, sampled boreholes, and hydraulic conductivity measurements. This cross section generally represents studies depicted in figure 3b.

0900 hours. Hourly data were recorded for 1 to 4 days each month.

Soil samples were collected for water potential measurements in the laboratory. The boreholes were drilled with a hollow stem auger (0.2 m diameter), and samples were collected in shelby tubes (76 mm diameter).

Laboratory Methods

Water potential was measured in the laboratory with a Decagon psychrometer SC-10 sample changer. The psychrometer was calibrated with NaCl solutions of known osmotic potential. Water potentials of -0.01 to -10 MPa correspond to relative humidities of 93 to 100%; therefore, all measurements were conducted in a glove box that was lined with wet paper towels to minimize moisture loss from the samples. One-second readings were recorded for a period of 120 seconds on a Terra8 data logger.

Field psychrometers were calibrated in the laboratory prior to installation at five different water potentials and at three different temperatures. Calibration procedures similar to those outlined in 5 were followed. Cooling current (5 ma), condensation time (30 s), and voltage endpoint determination were kept constant during calibration and field measurement of water potential.

## RESULTS

### Water Content

Volumetric moisture content monitored with the neutron probe ranges from 1 to 32% (Fig. 5). Differences in soil moisture with depth correspond to variations in sediment type (sandy and gravelly loam, and clay). Although moisture content within the clays is high, sand beds within the clay have low moisture content. Porosities of surficial sediments calculated from bulk densities range from 40 to 56%, which indicates that these materials are ~ 10 to 20% saturated. Porosities of the clays (borehole 23, 11 to 19 m; borehole 27, 14 to 24 m) range from 26 to 36% and average 31%.[6] Based on these porosity data, moisture content of the deep clays logged at access tubes 18 and 19 is close to saturation. Gravimetric moisture contents determined from soil samples range from 2 to 18%. Highest moisture contents are from the upper 1 m after rainfall (Fig. 6; borehole 15).

Monthly monitoring of the two deep access tubes (18 and 19) shows than moisture content remains constant with time throughout the section

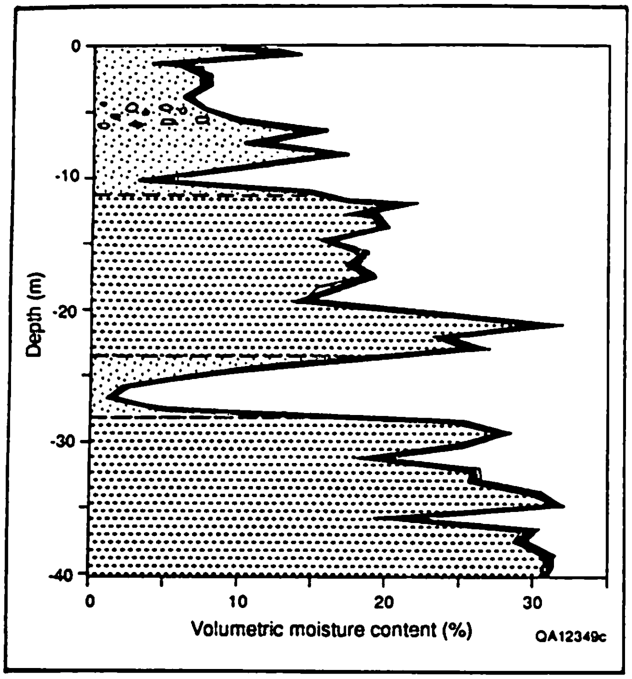


Figure 5. Variation in moisture content with depth and time in access tubes 18 and 19. Moisture content was monitored approximately monthly between July 1988, and July 1989, and 12 curves are represented. The relationship between moisture content and lithology is also shown. For location of access tubes, see figures 3b and 4.

(Fig. 5). Although neutron probe data are invalid at depths of less than 0.3 m, comparison of laboratory gravimetric moisture content data from borehole 50 (sampled after rainfall) and nearby borehole 51 (sampled after a long dry period) shows temporal variations in moisture content down to a depth of 0.3 m. Wetting fronts were observed by shallow coring down to a depth of 0.15 m after rainfall events.

Water Potential

Results from laboratory psychrometric measurements of samples collected in June and July 1988 show that water potential generally increases with depth, except in the shallow subsurface after rainfall events (Fig. 6). Water potentials range from -0.1 to -15.6 MPa. The highest water potentials were measured in the top 1 m of soil after rainfall. The lowest water potentials were also measured at shallow depths after long dry periods.

Samples collected after a long dry period in the summer exhibited water potentials that range from -15.6 MPa near the surface to -1.5 MPa at 9 m depth (Fig. 6). The hydraulic gradient is steepest near the

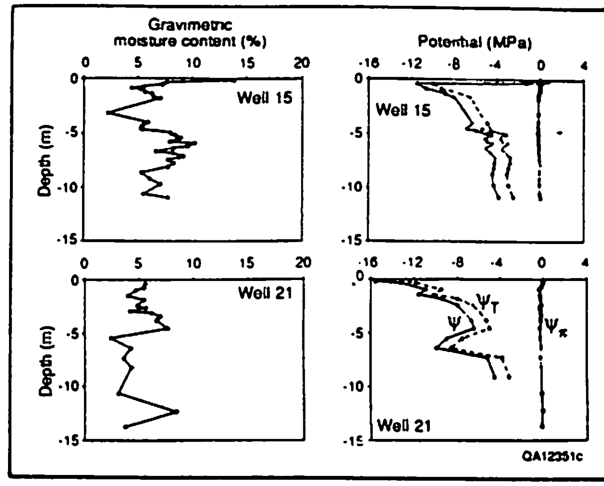


Figure 6. Profiles of gravimetric moisture content and water potential for samples from boreholes 15 and 21. All samples were collected in July and August 1988. For location of boreholes, see figures 3b and 4.

surface, approximately -5 MPa $m^{-1}$. Similar water potentials were recorded in samples from other areas (boreholes 41; -3 to -12 MPa: and borehole 15; -1 to -12 MPa). Samples collected after a rainfall event had water potentials near the surface close to 0 MPa but decreased to -12 MPa within 1.5 m. For detailed examination of the hydraulic gradient near a wetting front, samples were collected at 0.05-m intervals . A decrease in water potential from -0.1 to -13.2 MPa within a 0.05 m depth interval indicates that the wetting front is very sharp. Samples collected in the winter (boreholes 30 (-2 to -8 MPa), 31 (-3 to -7 MPa), and 54 (-3 to -8 MPa)) exhibited generally higher water potentials and a smaller range than those measured in the summer.

The shallow *in situ* psychrometers (< 1.4 m depth), which were installed in close contact with the surrounding sediments, equilibrated within a day, whereas the deeper psychrometers (2.7 to 14.3 m) installed in the borehole required approximately 20 days to equilibrate. The rate of equilibration decreased with time. After equilibration, the water potentials measured by the psychrometer pairs generally agreed within 0.2 MPa, which signified that the water potentials are reliable. Water potentials increased with depth, indicating a potential for upward movement of water. Water potentials of the shallow psychrometers (0.3 m depth) went out of range (< -8 MPa) on June 12, 1989, because the system became too dry. Water potentials in psychrometers at 0.5 and 0.8 m depth gradually decreased during the monitoring period. The moni-

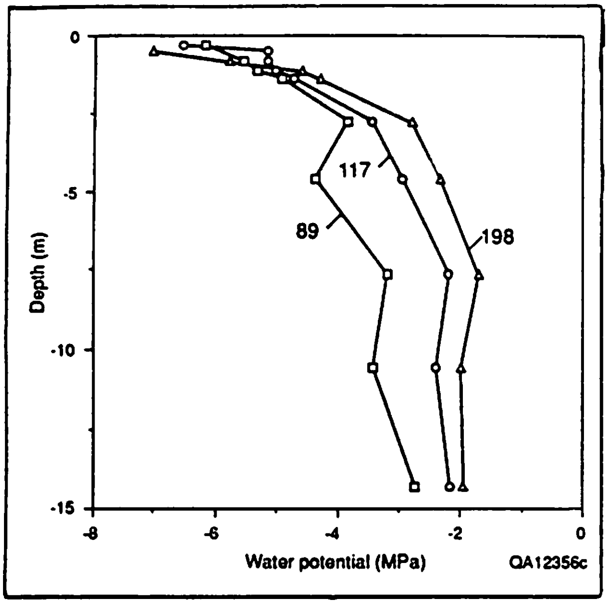


Figure 7. Vertical distribution of water potentials measured between March 30 and July 1989 in borehole 20. Profiles are labeled in Julian days. For location of borehole 20, see figures 3b and 4.

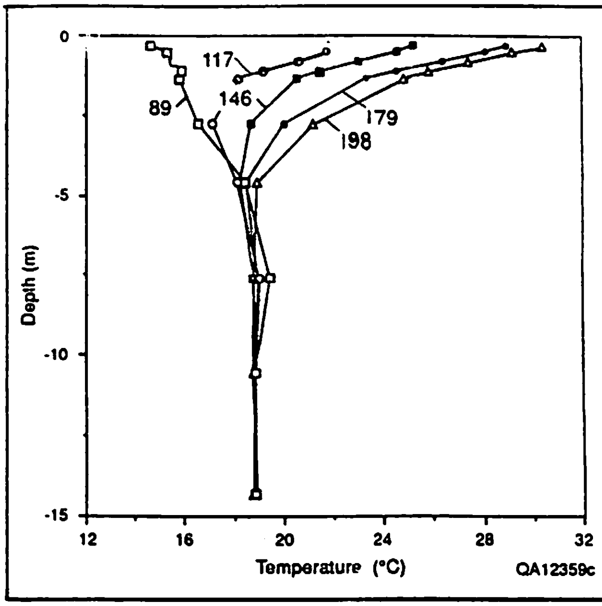


Figure 8. Vertical distribution of temperature measured between March 30 and July 17, 1989, in borehole 20. Profiles are labeled in Julian days.

toring period (March to July) was too short to evaluate seasonal fluctuations in water potential. Diurnal variations in water potential were restricted to the psychrometers at a depth of 0.3 m. The water potential gradient was steepest in the upper 3 m and approached unity at greater depth (Fig. 7), and the gradient became steeper with time at shallow depths.

Temperature oscillations over time were most pronounced in the shallow zone and decreased with depth. Hourly variations in temperature were primarily restricted to psychrometers at 0.3 m depth. Temperature variations at this depth lagged surface temperature variations by ~ 12 hours. Temperature gradients were initially negative upward but reversed direction in early April (Fig. 8). During most of the monitoring period, therefore, temperature gradients opposed water potential gradients. Temperature gradients (Fig. 9) were steepest down to a depth of 5 m and were near unity below this depth. Gradients became steeper during the monitoring period, and the depth of penetration of the steep gradients increased.

## DISCUSSION

### Moisture Content and Water Potential

Spatial variability in moisture content is controlled primarily by variations in lithology. Discontinuities in moisture content across different lithologies indicate that moisture content variations with depth cannot be used to determine the direction of water movement. Temporal variations in moisture content are restricted to the upper 0.5 m of the study site. The shallow penetration of the wetting zone is attributed to the fine grain size (clay to loam) and high porosity (40 to 56%) of the surficial sediments, which are generally 10 to 20% saturated during dry periods. The absence of temporal variations in moisture content monitored in deep access tubes indicates that recharge pulses are not moving through the system. Because a constant flux could result in temporally invariant moisture content, the absence of such variations is inconclusive evidence of a lack of recharge.

The range in water potentials measured by the laboratory psychrometer (-15.6 to -0.1 MPa) is compatible with the range observed from the in situ psychrometers (< -8 to -1.7 MPa) and suggests that sample drying during collection for laboratory measurements was minimal. The low observed water potentials indicate that water in the unsaturated zone is primarily in the vapor phase. The hydraulic conductivity that would correspond to these water potentials is very low; therefore moisture fluxes are expected to be minimal. Except in the upper 0.5 m after rainfall, the hydraulic gradients indicate upward flow, probably controlled by evapotranspiration. Increased evapotranspiration

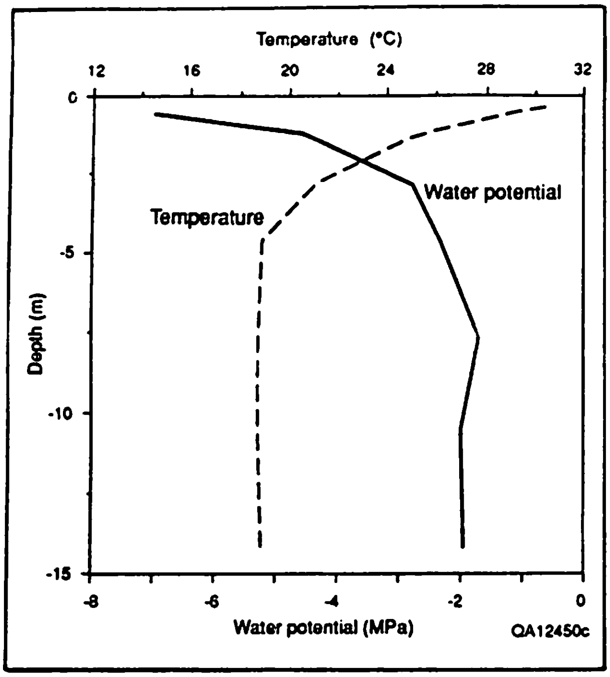


Figure 9. Comparison of water potential and temperature profiles measured on Julian day 198.

in the summer is reflected in higher water potentials of samples collected in the summer than in the winter. In addition, temporal variations in water potential of *in situ* psychrometers show gradually increasing water potential from March through July. Temperature gradients generally opposed water potential gradients and could result in downward movement of water vapor, although recent work[7] suggests that this flow component may be minor.

### Numerical Modeling Studies

As part of the Chihuahuan Desert site characterization, several hypothetical and observed situations will be evaluated using numerical models. Preliminary simulations of water flow are simplistic and limited to single-phase, isothermal flow. Steady-state upward flow was modeled with VAM2D.[8] As an initial condition, potential in the unsaturated zone was set in hydraulic equilibrium with the water table. A constant water potential of -6 MPa was imposed at the upper boundary, and the lower boundary was the water table at 0 MPa. Although information on characteristic curves is not yet available for sediments at the site, retention data [9] were used. Results of the simulations show good agreement between observed and predicted water potentials when a layered system was considered, but poor results were obtained when the site was represented as lithologically homogeneous. Lithologic variation, however, is much more complex than that represented by the simple layered system.

Future simulations will consider transient downward movement of water in the unsaturated zone. Because of the difficulty in simulating very dry initial conditions, sensitivity analyses will be conducted to determine the accuracy to which initial conditions must be represented. Potential pathways for radionuclides will be evaluated. The effect of the sand-clay boundary on water flow will be examined, and the ability of dry sands, which are interbedded with the clays, to prohibit downward movement of radionuclides will be assessed. A series of increasingly complex simulations will be conducted, including an analysis of nonisothermal vapor flow.

### Other Approaches

The physical approach described in this paper provides information on the potential for water movement but does not provide direct evidence that water actually moved in a particular direction. In addition to a physical approach, chemical tracers, such as $^{35}Cl$, $^{36}Cl$, and $^{3}H$, are being used to evaluate water movement. These tracers reflect an integrated history of flow during a much longer period (30 to 10,000 years) than that considered in the physical studies, and assist in prediction of flow conditions for a 500-year period. Chemical tracers provide information on the net water movement, which is difficult to assess in the shallow unsaturated zone because of cycling between infiltration and evaporation. Comparison of tritium and $^{36}Cl$ data will be used to evaluate the relative importance of liquid and vapor flow.

### Implications for Low-Level Radioactive Waste Disposal

Preliminary results of unsaturated flow studies indicate that the site is suitable for low-level radioactive waste disposal. Installation of the disposal facility may significantly alter the natural hydrologic system. If water leaches from the facility, it should be attenuated before penetrating very deep because of the low moisture contents and water potentials. Very dry sands (down to 1% volumetric moisture content) interbedded with the clays would serve as an additional barrier to water movement. Because it is critical to preserve the natural system, future work will evaluate cap designs in an attempt to replicate the natural system, and numerical modeling and field experiments will test various cap designs.

## CONCLUSIONS

Variations in moisture content are related to the distribution of sediment types in the unsaturated zone. Low moisture contents were measured in the shallow (12 to 15 m) sandy and gravelly loam. These sediments are 10 to 20% saturated. Although moisture contents within the underlying clays (140-m thick) are much higher, interbedded sands within the clays have very low moisture contents (≥ 1%). The clays are close to saturation. Monthly monitoring of moisture content in 41-m deep access tubes indicates that recharge pulses are not moving through the system. Measured water potentials are low (≥ -15.6 MPa) and indicate that water is primarily in the vapor phase. Water potential gradients are generally positive upward, except in the shallow subsurface after rainfall events. These upward gradients indicate that water movement is primarily upward, probably controlled by evapotranspiration. Thermal gradients generally oppose water potential gradients and may result in some downward movement of water. Numerical simulations of steady upward flow in a simplified layered system matched observed water potentials. Future modeling will include analysis of nonisothermal vapor flow. Preliminary interpretations of these data indicate that the attributes of the unsaturated zone at the site are suitable for waste disposal because of the low water potentials and upward gradients. Future studies will evaluate cap designs that will attempt to replicate the hydrology of the natural system.

## ACKNOWLEDGMENTS

This research was supported by the Texas Low-Level Radioactive Waste Disposal Authority. The manuscript was reviewed by A. R. Dutton and T. F. Hentz. Publication authorized by the Director, Bureau of Economic Geology. Climatic data were provided by J. F. Griffiths, Meteorology Department, Texas A & M University. I would like to thank R. W. Brown, G. W. Gee, and P. J. Wierenga for providing technical information during the study.

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Eagle Springs.

# REGIONAL GROUND-WATER SYSTEMS IN NORTHERN TRANS-PECOS TEXAS

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## INTRODUCTION

Northern Trans-Pecos Texas (Fig. 1) includes the Diablo Plateau, the Salt Basin and its extensions, the Delaware and Apache Mountains, the Wylie Mountains, the Davis and Barilla Mountains, the Rustler Hills, and the Toyah Basin. The area is bordered on the northeast by the Pecos River and is transitional eastward to the Edwards Plateau. Surface water is essentially nonexistent in this area; the Pecos River is too salty to use, and the locally important springs are widely scattered. Consequently, ground water is of paramount importance; ground-water systems are both local and regional. Extensive irrigation districts have been established in the Toyah Basin, Balmorhea, Dell City, Wild Horse Flat, Michigan Flat and Lobo Flat areas (Fig. 1). Hydrological studies of this area are numerous, but it is only in the last several years that a regional understanding of the ground-water systems and the processes which control them has been assembled.

Geological studies of the area are also numerous and it is the geology (see Barnes, 1983), which varies from basin-and-range to stable platform and intracratonic basins, that gives these Trans-Pecos ground-water systems their flavor. Uplift, faulting, salt dissolution, and Tertiary volcanism were the significant geological processes in the evolution of this area. Recent data also indicate the importance of paleoclimatic events. Some of the ground water is probably of Late Pleistocene in age (Lambert and Harvey, 1986; Kreitler and others, 1987; LaFave and Sharp, 1987), and modern recharge is minimal and irregularly distributed. The three most important aquifers are in the Toyah Basin alluvium, in Wild Horse Flat (a combined alluvium/limestone aquifer), and on the Diablo Plateau in the vicinity of Dell City. Numerous other minor aquifers are used, mostly for domestic or livestock use.

## HYDROSTRATIGRAPHY

The oldest hydrogeologically significant rocks are Permian, although Pennsylvanian through Precambrian rocks are present at depth (McMahon, 1977), and crop out in the Diablo Plateau and south and west of the study area. Permian strata are divided into

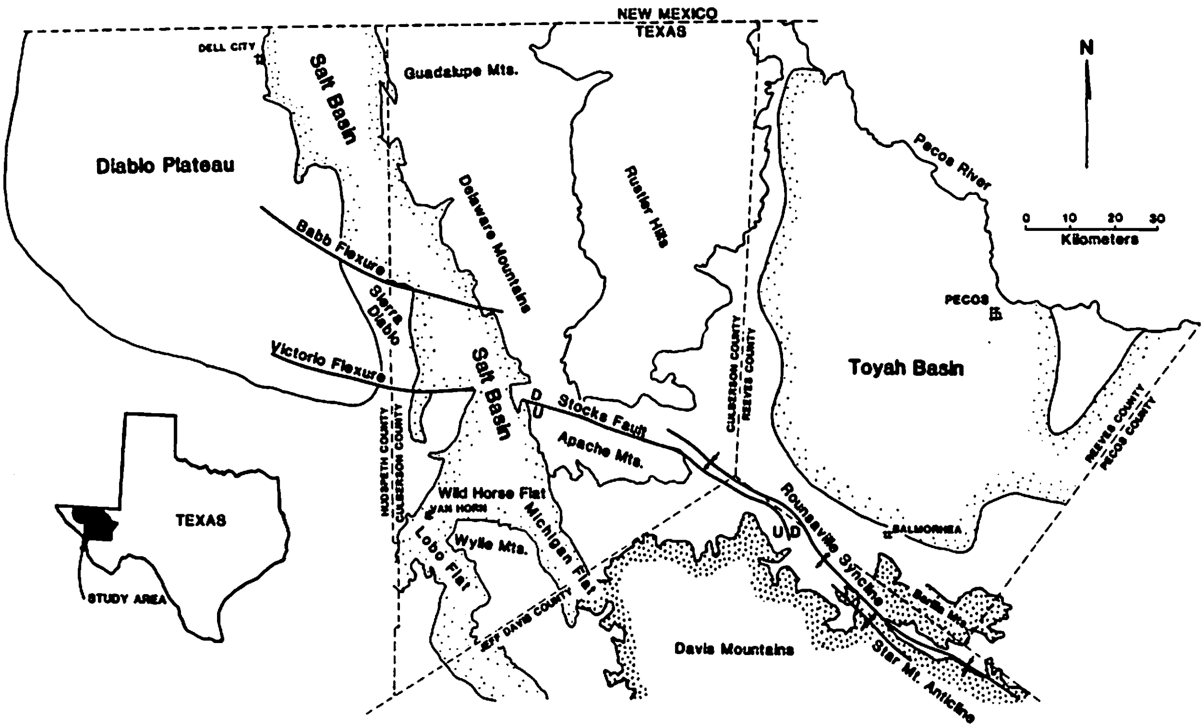


FIGURE 1. Major geologic features of northern Trans-Pecos Texas. The Davis Mountains consist mainly of extrusive igneous rocks. The Toyah Basin and the Salt Basin with its extensions contain thick sequences of Cenozoic fluvial, alluvial, and lacustrine fill. Cropping out over the remainder of the study area are mostly Permian but also Cretaceous sedimentary rocks.

four series--Wolfcampian, Leonardian, Guadalupian, and Ochoan*. These units can be subdivided into three hydro-geologic facies (Fig. 2; 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 the low-permeability Guadalupian and Ochoan Series basin facies, which fills the Delaware Basin.

Shelfal facies rocks form the aquifer that serves both the Dell City irriga-tion district and ranches throughout the Diablo Plateau. 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 which extends circum-ferentially around the Delaware Basin (Adams, 1944; Hiss, 1980). The reefal facies exerts a major control on regional ground-water flow systems in Trans-Pecos Texas.

The basinal facies are located east of the Guadalupe Mountains. These Guadalupian and Ochoan formations are low in permeability, except where exposed at the surface, and water quality is generally poor. Guadalupian carbonate/evaporite formations are the Seven Rivers, the Yates, and the Tansill; no water is produced from these formations in the study area. Guadalupian siliciclastic formations of the Delaware Mountain Group are the Brushy Canyon, Cherry Canyon, and Bell Canyon, which yield minor amounts of ground water. Included in the Ochoan Series are the Castile, Salado, and Rustler Formations. The Castile 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-

*See stratigraphic columns elsewhere in this volume.

quality water which is used for irrigation and livestock. The outcrop of Ochoan rocks corresponds to the outline of the Rustler Hills on Figure 1.

Overlying the Permian is the Triassic Dockum Group which yields good-quality water in the southeastern part of the Toyah Basin. The Dockum dips to the southwest and pinches out about 12 miles southwest of the Pecos River. Overlying the Dockum are Cretaceous Comanchean and Gulfian carbonates and sandstones. Hydrogeologically important units are the Cox Sandstone and the Edwards, Georgetown, and Boquillas limestones. The Cox Sandstone unconformably overlies either the Permian Dewey Lake or the Cretaceous Yearwood Formation in the study area. It 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 south and east of the study area. The Cretaceous carbonate units provide water to ranches north of the Davis Mountains. Paleokarst features are present in these formations.

Rocks of Tertiary age form the Davis Mountains and include the McCutcheon volcanics, consisting of the basal Jeff Conglomerate, interbedded lava flows, tuffs, and nonmarine sedimentary rocks. These are capable of yielding minor quantities of water for domestic and livestock use.

Of great hydrogeological significance are Quaternary and Tertiary alluvial sediments which veneer much of the study area and attain thicknesses of over 2400 feet in the Salt Basin (Gates and others, 1980) and over 1500 feet in the Toyah Basin. These sediments are dominantly clastic, although gypsum and caliche are probably present (Ogilbee and others, 1962). The alluvium provides water for irrigation in Wild Horse Flat and the Toyah Basin, 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). In the vicinity of Balmorhea, some shallow wells produce small amounts of water from undifferentiated alluvium/Cretaceous 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.

## STRUCTURAL SETTING

The northern Trans-Pecos is a transitional area from the Cenozoic 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 feet of Paleozoic sediments and is bounded by Capitan Reef rocks which are exposed in the Guadalupe Mountains, Sierra Diablo, and Apache Mountains (Fig. 1). The reef trend continues north-northeastward into New Mexico and southeastward in the subsurface. Basin-and-range style tectonics downdropped the Salt Basin in the Cenozoic. Some fault movement has con-tinued into the Quaternary (Goetz, 1977, 1980, and 1985).

The major geological processes have been extensional normal faulting, salt dissolution which created the Toyah Basin, and Tertiary volcanism which formed the Davis Mountains. Second-order structural features (Fig. 1), the importance of which will be discussed, include the Babb flexure, the Victorio flexure, and the Stocks Fault/Rounsaville syncline trend. The Babb and Victorio flexures existed in the earliest Permian time; they now transect and offset the Salt Basin.

The Stocks Fault is one of a set of east-trending brittle fractures which are evident north of the Davis Mountains (LaFave and Sharp, 1987), and may, possibly, also be present beneath the Tertiary volcanics. The Stocks Fault bounds the north-northeastern flank of the Apache Mountains. DeFord (pers. comm., 1986) and Wood (1965) state that the great throw of the Stocks Fault is the result of subsurface dissolution of Delaware Basin evaporites. Along this fault, permeable strata of Permian and Cretaceous age are in hydrogeologic contact. The Rounsaville Syncline and Star Mountain Anticline parallel the Stocks Fault to the south. The large springs near Balmorhea are located near the syncline (Fig. 3).

The orientation of regional fractures changes across the study area (King, 1949). The eastern margin of the Salt Basin is intensely faulted; faults parallel the north-trending basin boundaries. In the Rustler Hills a few miles to the east, in contrast, the fractures and minor fold axes have a strong easterly trend which changes to an east-southeast orientation near the Stocks Fault and adjacent folds. The bounding faults on

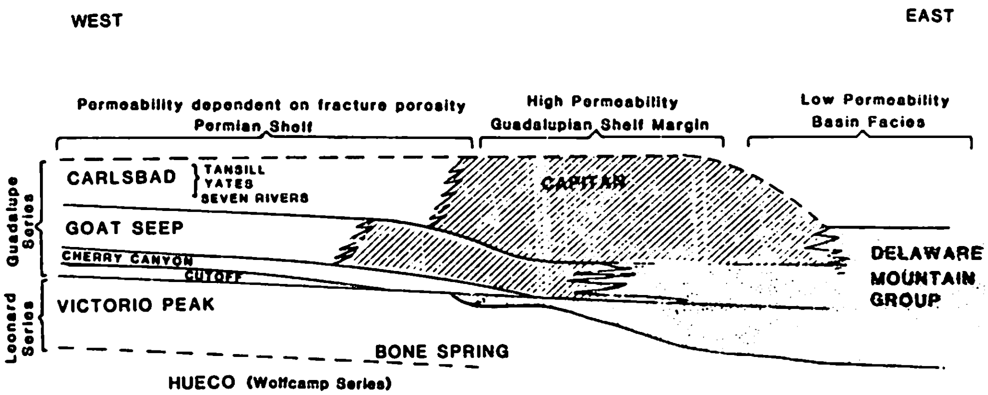


FIGURE 2. Generalized Permian hydrostratigraphic facies (from Nielson and Sharp, 1985, with permission of the West Texas Geological Society). The highly permeable shelf-margin or reefal facies separates the shelfal facies (permeable where fractured) from the low-permeability basinal facies. The underlying Wolfcampian Series is not shown, nor is the overlying Ochoan Series, which is present only in the Delaware basin.

the west side of the Salt Basin are more narrowly confined. Muehlberger (pers. comm., 1988) suggests that the graben is tilted to the west and that many faults, suspected to parallel the western boundary of the Salt Basin, are covered by bolson fill. Just west of the basin, a general NW-trending structural grain is pre-valent; this grain is subparallel to the Babb and Victorio flexures.

East of the Rustler Hills, the Toyah Basin contains fluvial Cenozoic fill; the bulk of this fill is thought to be Quaternary, but some is certainly also Tertiary (Maley and Huffington, 1953). The Toyah Basin was created by Late Tertiary and Quaternary dissolution of Permian evaporites--salts of the Castile and Salado Formations and anhydrite and gypsum from the Rustler Formation.

The Davis Mountains (Fig. 1) are composed of volcanic tuffs, rhyolites, and intercalated sediments. These rest nonconformably over Cretaceous sedimentary rocks. The volcanic rocks have been undergoing slope retreat in the Cenozoic and probably were present much farther to the north and east than at present (Halamicek, 1951).

## REGIONAL FLOW SYSTEMS

While there was no comprehensive study of area ground-water systems prior to their municipal and agricultural development, but it is possible to piece together approximate predevelopment potentiometric surfaces from a series of reports and unpublished data in the files of the Texas Water Commission. These are for the Toyah Basin in 1940 (Lang, 1943; see also LaFave and Sharp, 1987), for the Salt Basin including Wild Horse, Lobo, and Michigan Flats in the late 1950s (Hood and Scalapino, 1951; Nielson and Sharp, 1985), and for the Dell City area (Scalapino, 1950). Data for the Diablo Plateau are presented by Kreitler and others (1987). 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 Aquifer. Ground water in the Diablo Plateau has not been extensively developed, except near Dell City, so its present potentiometric surface is probably similar to predevelopment conditions. Flow systems in the Rustler Hills are still not well delineated.

Figure 3 shows the interpretation of these surfaces. There are two regional predevelopment discharge areas--the northern and middle sections of the Salt Basin and the Pecos River on the northeastern boundary of the study area. The Salt Basin is divided into three flow systems: the northern section, the middle section, and the southern section or Wild Horse Flat. There are playas in the northern and middle sections, indicating evaporative discharge. This inference is borne out by water-chemistry studies (Gates and others, 1980; Boyd, 1982; and Chapman, 1984) which show increasing salinity in the direction of flow. In these sections, gypsum and halite are precipitated from ground water in the capillary fringe.

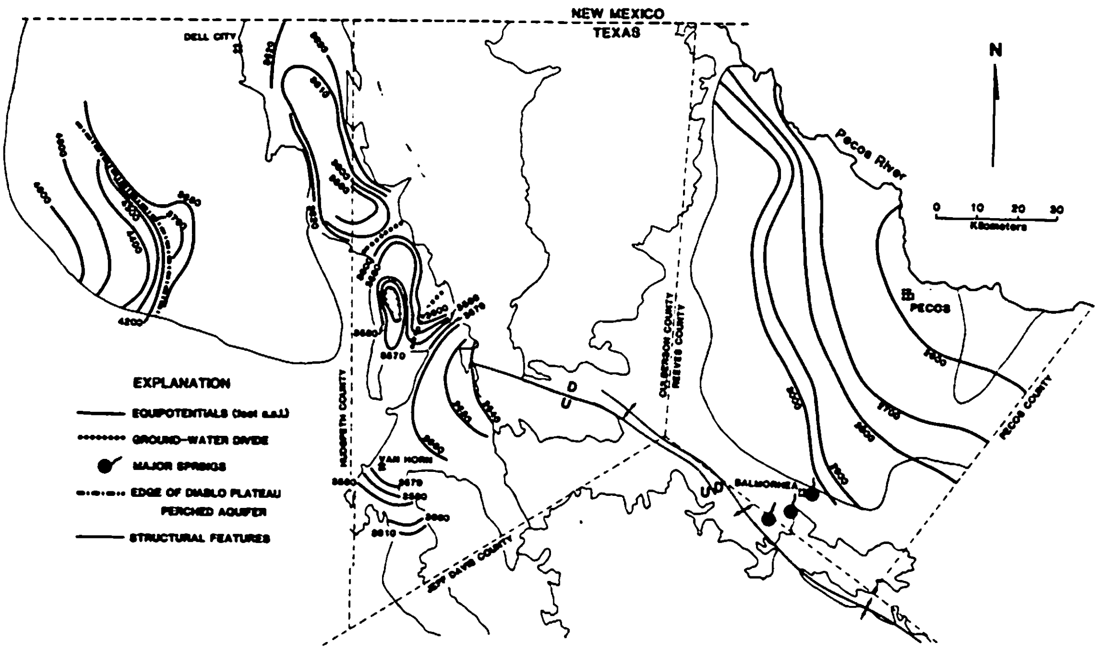


FIGURE 3. Ground-water systems approximating conditions before their municipal and agricultural development. Contour intervals are variable: 10 feet for the Salt Basin, and 100 feet for the Toyah Basin and the Diablo Plateau.

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 permeablė shelf margin carbonates. On the eastern flank of the Salt Basin, slightly brackish water can be found at depths of more than 2000 feet in Capitan Reef rocks (Reed, 1965). In contrast, near Dell City, on the western flank, there is very little topographic relief on the western margin of the graben which merges imperceptibly with the Diablo Plateau. Ground-water flow on the Diablo Plateau was and is generally northeast and east to the Salt Basin. Irrigation pumpage near Dell City has created a lasting 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 (Ashworth, 1988, pers. comm.). The low gradient in the Plateau is created by very low rates of recharge (probably concentrated along the drainage ways) and the constriction of subsurface flow to a few highly permeable fractures.

The existence of a perched aquifer in the southwest part of the Diablo Plateau is a new finding. Kreitler and others (1987) suggest that it is perched within Cretaceous limestone; and steep water-table gradient implies a lower hydraulic conductivity than in the underlying Permian Bone Spring-Victorio Peak aquifer. Fracturing associated with the Babb flexure contributes to the higher permeability of these Permian units. Ground-water salinities in the Diablo Plateau, except for the Dell City area, range from 900 to 3,000 mg/l, the lower values associated with larger amounts of recent infiltration.

Ground-water divides separate the three sections of the Salt Basin. Nielson and Sharp (1985) noted the proximity of the divides to the Babb and Victorio flexures and suggested that tectonic movement and resultant trends of sedimentation created permeability barriers, in addition to controlling the locations of the surface drainage systems and the alluvial fans that serve as the prime recharge sites.

In Wild Horse Flat, in the southern section of the Salt Basin, predevelopment flow was from the southwest and northwest to the east. Recharge occurs primarily on alluvial fans on the western flank of the flats and on the northern margin of the Wylie Mountains.

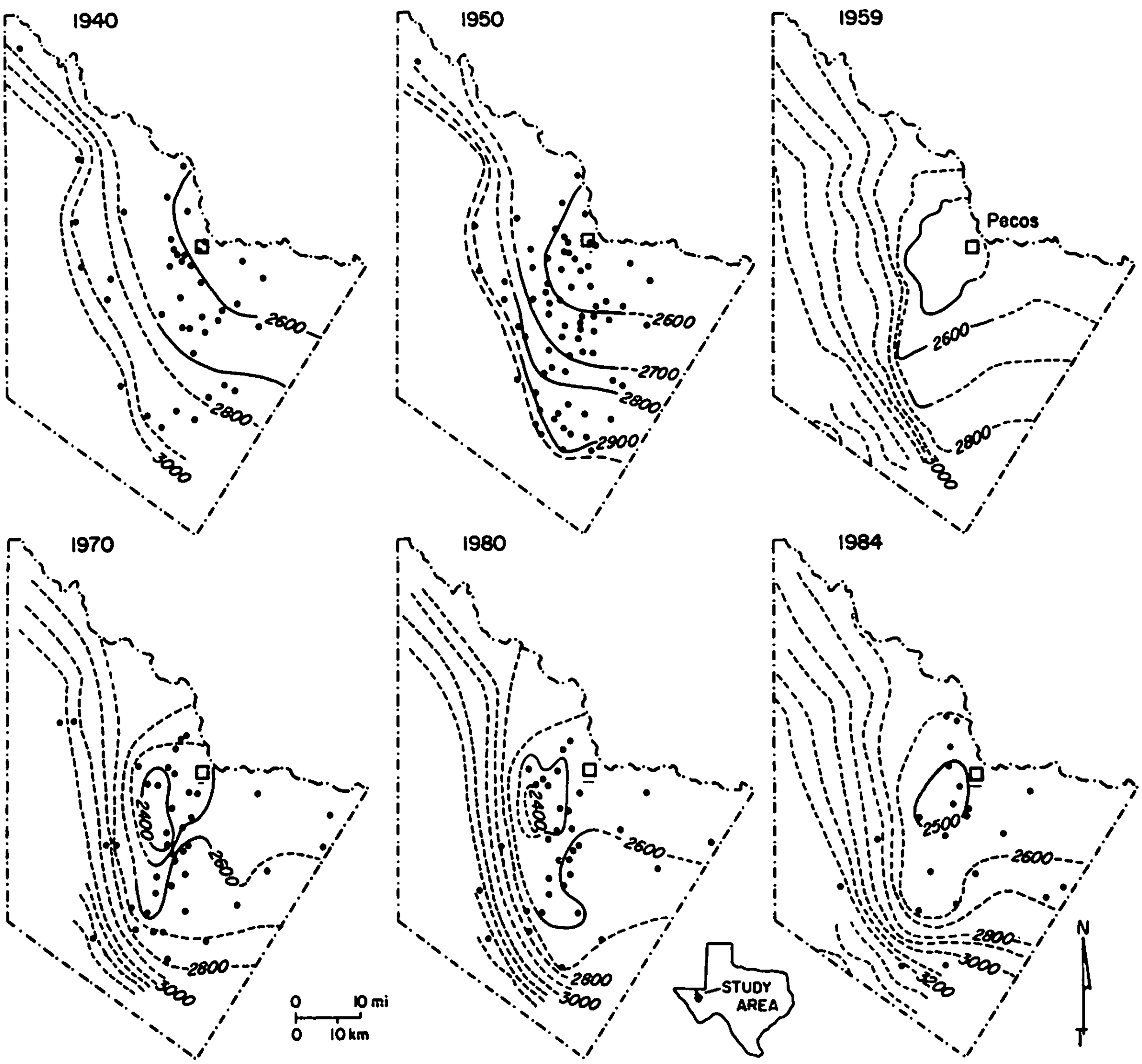


FIGURE 4. Water table maps of the Toyah Basin, Reeves County, in 1940, 1950, 1959, 1970, 1980, and 1984 from LaFave (1987). Data are from the files of the Texas Water Commission (control points are shown), except for the 1959 map which is adapted from Ogilbee and others (1962).

There is also intrabasin flow from the south; in particular, Lobo Flat contributes, but the magnitude of flow from Lobo Flat is uncertain. The steepening of the water table gradient south of Van Horn is coincident with east-trending faults (Hay-Roe, 1958; Twiss, 1959). Recharge also occurs from precipitation along ephemeral streams, such as Wild Horse Creek. The predevelopment water table in Wild Horse Flat was about 100 feet beneath the surface in contrast to the evaporative discharge systems in the basin to the north. The shelf margin (reef) facies rocks of the Apache Mountains act as a drain for Wild Horse Flat. The main predevelopment discharge was interbasinal and eastward towards the Toyah Basin.

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. It is logical to assume that these hydrostratigraphic units are highly anisotropic, and that the direction of greatest permeability would be parallel to the Stocks Fault (Figs. 1 and 3).

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. The springs issue from orifices at an elevation of about 3,300 feet. This would imply 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 Toyah Basin aquifer produces from the Cenozoic alluvium and, in the eastern section, from undifferentiated alluvium and Cretaceous limestones. Recharge to the aquifer is from the ephemeral streams which drain the Davis and Barilla Mountains and the Rustler Hills. In addition, there is apparently a significant subsurface component of interbasin flow. LaFave (1987) noted that the aquifer produces a 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 ground waters produced from Capitan Reef aquifers and from Balmorhea Springs. Regional flow would include water recharged in the Delaware Mountains as well as interbasin flow from Wild Horse Flat. This hypothesis is in concurrence with the reports of Hiss (1980), Mazzulo (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 if these waters recharge the Toyah Basin aquifer chiefly by subsurface flow or by infiltration along the many draws which drain eastward from the Rustler Hills. Finally, on the eastern margins of Reeves County, better-quality (less than 1500 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 man on these ground-water systems is dramatically shown in Figure 4. Irrigation pumpage in Reeves County increased rapidly after 1945. Many of the large springs in the study area have since ceased to flow (Brune, 1981). Figure 4 shows how irrigation pumpage from the Toyah Basin aquifer in Reeves County lowered water-table elevations to create a cone of depression southwest of Pecos. This pumpage totally altered the regional flow-system discharge zone from the Pecos River to irrigation wells within Reeves County. 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. Recent declines of pumpage because of increased costs has allowed partial recovery, evident in the 1984 map. It seems doubtful, however, that predevelopment conditions will again be approached throughout the study area.

## CONCLUSIONS

Geologic processes of faulting, folding, and dissolution in semiarid Trans-Pecos Texas have created the controlling framework for regional ground-water flow systems. There are two major discharge areas--the northern and middle sections of the Salt Basin, and the Pecos River. The fractured Permian shelf-facies aquifers of the Diablo Plateau discharge in the playas as does water recharging through the fringing alluvial fans. Flow in Wild Horse Flat, on the other hand, leaves the Salt Basin by interbasin flow through the permeable Permian shelf-margin-facies rocks of the Apache Mountains. Physical, geological, and chemical data indicate that regional, interbasin flow discharges at the springs of Balmorhea and in the Toyah Basin. The Toyah Basin, created by subsurface salt dissolution, receives recharge also from the Rustler Hills and the Delaware Mountains. This aquifer formerly discharged into the Pecos River, but intensive irrigation pumpage has halted the natural discharge.

## ACKNOWLEDGMENTS

J. A. Ashworth, R. K. DeFord, P. W. Dickerson, S. Ellison, and W. R. Muehlberger read manuscript drafts. I also thank my former students, Pam Nielson and John LaFave, and my colleagues at the Texas Bureau of Economic Geology, Charles Kreitler and Bill Mullican, for fruitful discussions and for introducing me to the Trans-Pecos.

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