

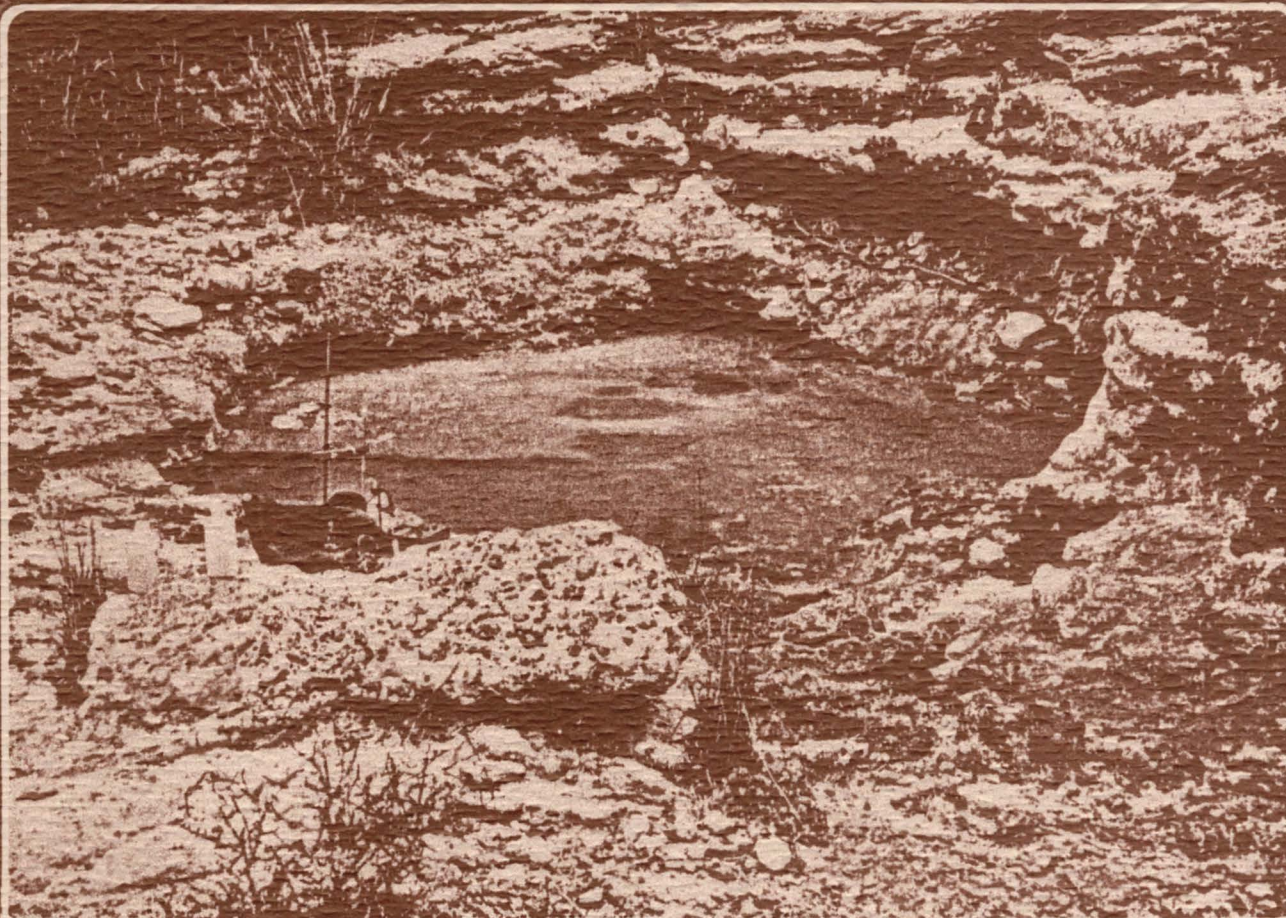
1979

**GEOLOGIC SETTING AND GEOCHEMISTRY OF  
THERMAL WATER AND GEOTHERMAL ASSESSMENT,  
TRANS-PECOS TEXAS**

*with Tectonic Map of the  
Rio Grande Area, Trans-Pecos Texas  
and Adjacent Mexico*

by

Christopher D. Henry



BUREAU OF ECONOMIC GEOLOGY ■ THE UNIVERSITY OF TEXAS AT AUSTIN ■ AUSTIN, TEXAS 78712

W. L. Fisher, Director



QAe2092



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Funded by United States Energy Research and Development Administration,  
contract number EY-76-S-05-5106.

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Tectonic map of the Rio Grande area, Trans-Pecos Texas and adjacent Mexico	in pocket
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## ABSTRACT

Hot springs and wells in West Texas and adjacent Mexico are manifestations of active convective geothermal systems concentrated in a zone along the Rio Grande between the Quitman Mountains and Big Bend National Park. Maximum temperatures are 47°C and 72°C for hot springs and wells in Texas and 90°C for hot springs in Mexico within 5 km of the border.

The area lies along the eastern margin of the Basin and Range province in what may be an extension of the Rio Grande Rift. The heat source for the thermal waters is deep circulation of ground water in an area of relatively high thermal gradient. Recent volcanism is not a source of heat because, based on both paleontologic and isotopic evidence, the youngest observed igneous activity in West Texas is Miocene.

Most hot springs lie on or immediately basinward of normal faults, at the edges of late Tertiary basins formed by east-west extension. This setting implies that faults are permeable channelways which allow thermal water to rise from below. Recharge for the thermal systems probably occurs in adjacent highlands. Hot springs are not restricted to faults with large displacement or to particular rock types. However, many faults do show evidence of recent movement, which may be important in keeping fracture systems permeable.

The setting of hot springs in basins composed of permeable sediments implies that, although thermal circulation could occur along other faults, the water does not discharge to the surface. For example, several wells tap hot water at depths of about 20 to 1,000 m.

The thermal waters fall generally into three chemical groups that can be related to subsurface host rocks. Waters circulating entirely within carbonate and clastic sediments contain moderate total dissolved solids composed of Ca, Mg, HCO<sub>3</sub>, and SO<sub>4</sub>. Waters from zeolitized, silicic volcanic rocks contain low to moderate total dissolved solids composed primarily of Na and HCO<sub>3</sub>. Evaporite waters have high total dissolved solids, Na, Cl, SO<sub>4</sub>, HCO<sub>3</sub>, and Li. According to their chemistry and geologic setting, evaporite waters have been in contact with evaporites and limestone. Gradation between groups indicates mixing of waters from different host rocks.

Interpretation of silica and sodium-potassium-calcium geothermometers is complicated by the geologic setting and

geochemistry of thermal waters. Solution of evaporites and nonequilibrium with feldspars partly negate use of the sodium-potassium-calcium method. Solution of amorphous silica and nonequilibrium with quartz complicate use of the silica method.

The best interpretation of geothermometry indicates that there is an intermediate-temperature group of springs with maximum subsurface temperatures approximately equal to their surface temperatures—around 60°C. Three thermal systems, the two Gulf Wells in Texas and Ojos Calientes in Mexico, have higher subsurface temperatures, ranging from at least 100°C up to possibly 160°C.

According to the geologic setting and geothermometry of hot spring systems, the most promising area for geothermal energy is the Presidio Graben, an actively subsiding basin along the Rio Grande. This area has the densest concentration of hot springs and wells and the highest surface and subsurface temperatures (90° and 160°C, respectively). The high temperatures result from deep circulation of meteoric water in a region with high thermal gradient caused by crustal thinning. Hueco Bolson south of El Paso is in a similar geologic setting, but hot springs are found in only one area and have moderate (approximately 60°C) subsurface temperatures.

Other areas are not as promising for geothermal development as Presidio and Hueco Boleons. Although it has many hot springs, the Big Bend area has low subsurface temperatures and no evidence of recent faulting. Hot springs in Big Bend are probably a result of relatively shallow circulation in an area of normal heat flow.

The Lobo Valley area northwest of Marfa is a deep graben and has recent fault scarps and high silica concentrations in the ground water. However, there are no hot springs or wells, and the high silica content probably reflects shallow circulation through volcanic and volcanoclastic rocks containing amorphous silica. The lack of thermal water suggests normal heat flow.

Salt Basin is an active, shallow graben without hot springs; subsurface temperatures are low. It lies at the easternmost edge of the Basin and Range province and is probably underlain by a crust of normal cratonic thickness with low heat flow.



## INTRODUCTION

Geothermal energy is one of several alternative energy sources proposed to supplement our dwindling oil and natural gas supply. In recent years there has been intense geothermal exploration throughout much of the western United States. Trans-Pecos Texas is an area for potential development of geothermal energy, although information necessary to evaluate this potential was scarce until recently. All that was known before this study was that (1) numerous hot springs and wells occur along the Rio Grande in both Texas and Mexico, and (2) the geologic setting of Trans-Pecos Texas in the Basin and Range province is similar to many geothermal areas which have proven favorable for resource development in the western United States.

Given this brief but suggestive background, a preliminary evaluation of the geothermal potential in Trans-Pecos Texas appeared warranted. Within the Trans-Pecos region, the study focused on the Rio Grande Valley, which not only has most of the known hot springs or wells, but which also has the greatest geothermal potential, according to this study. However, other parts of Trans-Pecos Texas were also evaluated using the available data.

This report summarizes existing information and presents the results of an intensive study of the area. The study proceeded through several overlapping phases: (1) compilation of existing geologic information including regional

studies of geology, structure, and geophysics, and more detailed studies of individual areas of hot springs; (2) detailed mapping of hot spring areas to determine the origin and geologic controls of the springs; (3) field measurement and sampling of hot spring or well waters for geochemical analysis; and (4) synthesis and interpretation of the data.

Most previous work consisted of basic geologic mapping. Until the last few years no studies dealing directly with geothermal energy had been made. Nevertheless, basic geologic information is necessary for understanding the origin of thermal waters, and applicable studies are summarized in the sections on regional geologic setting, source of heat, and geologic setting of hot springs. This published work was supplemented by detailed examination and mapping of hot springs and significant structural areas; the results of the field work are presented in the section on geologic setting of hot springs. All known hot springs and wells and many cold springs and wells were sampled, and the waters were analyzed to gain a better understanding of the geochemical history of the thermal waters. This information is discussed in the sections on geochemistry and geothermometry. All this available information is synthesized in the section on geothermal model and area evaluation. The report concludes with suggestions for additional work to complete evaluation of the geothermal potential of the Rio Grande region.

## REGIONAL GEOLOGIC SETTING OF TRANS-PECOS TEXAS

Trans-Pecos Texas is situated in the Basin and Range province near its eastern boundary with the Great Plains (Fenneman, 1946). Late Tertiary to Recent crustal extension and normal faulting are the most obvious structural features controlling the origin of thermal waters and location of hot springs. However, older geologic structures influenced younger structures; understanding the older structures aids in evaluating the potential for geothermal energy. Exposed rocks range in age from Precambrian to Recent. Several periods of deposition were followed and separated by major deformational events.

Precambrian rocks exposed in several areas around Van Horn and in the Franklin Mountains near El Paso comprise the structurally highest part of Trans-Pecos Texas (fig. 1). The Carrizo Mountain Group, the oldest rocks in the area, consists of as much as 5,700 m (19,000 ft) of folded, regionally metamorphosed arkose, quartzite, schist, limestone, and rhyolite (King and Flawn, 1953). Deformation and metamorphism occurred about 1,250 million years ago (Denison and Hetherington, 1969). The highly deformed Allamoore and Hazel Formations are believed to be younger than the Carrizo Mountain Group, although all contacts between them are faults. The Allamoore Formation consists of 750 m (2,500 ft) of limestone and dolomite with minor volcanic rocks. The Hazel Formation consists of 1,500 m (5,000 ft) of conglomerate and sandstone (King,

1965). Both were complexly deformed but only mildly metamorphosed about 1,000 million years ago (Denison and Hetherington, 1969). All three formations are unconformably overlain by the Van Horn Sandstone, which is believed to be late Precambrian (King and Flawn, 1953) or Cambrian in age (McGowen and Groat, 1971). The Van Horn is tilted but unmetamorphosed.

Deposition resumed in the Cambrian and continued unbroken until Late Pennsylvanian time. During this time up to 4,500 m (15,000 ft) of rocks were deposited in the area of the present Marathon Uplift (fig. 1). Ordovician and Devonian rocks are limestone, chert, and novaculite. Pennsylvanian rocks are shales and sandstones grading upward into coarse conglomerate in Upper Pennsylvanian strata. The sequence was intensely deformed, folded, and thrust faulted during the Ouachita orogeny (King, 1935). Deformation probably began in Late Mississippian or Early Pennsylvanian time, producing uplifts which were the sources for coarser Upper Pennsylvanian strata, and ended in Late Pennsylvanian or Early Permian time (Flawn and others, 1961). Folded Paleozoic rocks are also exposed in the Solitario Uplift 60 km (35 miles) to the southwest (fig. 1). Structural trends in the Marathon region and in the Solitario extend southwestward into Mexico. However, southwest of the Solitario, the Paleozoic rocks are completely covered beneath thick Cretaceous and Tertiary sequences.

On the Diablo Platform to the north (fig. 1), equivalent Paleozoic rocks are only 750 m (2,500 ft) thick and are primarily shelf carbonates. The rocks were uplifted, gently folded, and faulted during the Pennsylvanian, probably contemporaneously with intense folding in the Marathon region.

Deposition over much of Trans-Pecos Texas resumed during the Permian. Depositional environments suggest that a deep-water basin, a possible forerunner of the Chihuahua Trough (fig. 1), had already developed in the Pinto Canyon area north of Presidio (Amsbury, 1958; DeFord, 1969; Wilson, 1970). Marine platform rocks, carbonates and sandstones, were deposited over the rest of Trans-Pecos Texas.

The Chihuahua Trough (fig. 1) may have originated as early as the Ouachita orogeny. It is bounded on the east by the Diablo Platform, a Pennsylvanian feature. Datable marine deposition began in the Chihuahua Trough during the Cretaceous. Before that, possibly as early as Permian time but more probably during the Jurassic (Haenggi, 1966), evaporite deposits of halite and gypsum accumulated in much of the trough. Total thickness of evaporites is unknown because they are deformed everywhere they have been observed. Following evaporitic deposition, up to 5,500 m (18,000 ft) of Cretaceous limestone and shale were deposited in the trough. The thickness of Cretaceous strata changes dramatically northeastward from the Chihuahua Trough onto the Diablo Platform. In the Quitman Mountains (fig. 2) there are more than 4,300 m (14,000 ft) of Lower Cretaceous rocks (Jones and Reaser, 1970). In the Eagle Mountains, Cretaceous sequences average about 2,100 m (7,200 ft) in thickness (Underwood, 1963), and on the Diablo Platform there are only 600 m (2,000 ft) of equivalent rocks. In all three areas Upper Cretaceous rocks have largely been removed by erosion. Amsbury (1958) reported approximately 850 m (2,800 ft) of Lower Cretaceous strata in Pinto Canyon. An equivalent section in the Sierra de la Parra in Mexico (fig. 2) just west of Pinto Canyon is 3,650 m (12,000 ft) thick (Gries and Haenggi, 1970). Individual formations within the Cretaceous show similar differences in thickness (DeFord and Haenggi, 1970).

Laramide compression in the Chihuahuan tectonic belt may have started as early as Cenomanian time, following deposition of the massive Buda Limestone. Upper Cretaceous and Lower Tertiary strata overlying the Buda are predominantly clastic, containing thin limestone, shale, clay, and sandstone beds, which were apparently shed from rising Laramide folds.

Deformation produced a series of north-northwest-trending tight and overturned folds. Rocks in the Chihuahua Trough were thrust northeastward along decollements within the evaporite sequence. Gries and Haenggi (1970) suggest that deformation was a result of tilting of the Chihuahua Trough to the east, which allowed the thick Cretaceous sequence to slide eastward over the evaporites. In contrast, the thin sequence of Cretaceous rocks on the Diablo Platform was only mildly deformed.

Although most post-Precambrian igneous activity in West Texas and northeast Chihuahua is Late Eocene or younger, some igneous rocks were deformed during the Laramide orogeny and, consequently, are older than middle Eocene. In Big Bend Park and in Chihuahua and Coahuila to the south, gabbroic sills in the Cretaceous Boquillas Formation are folded, along with the sedimentary sequence.

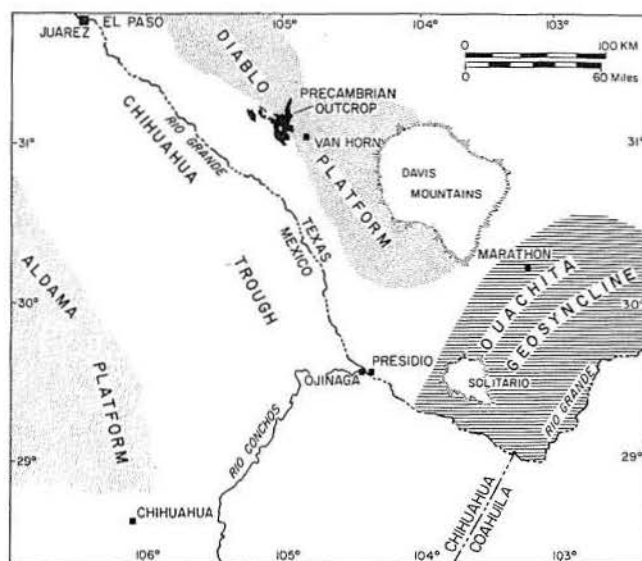


Figure 1. Major structural elements of Trans-Pecos Texas and adjacent Mexico.

Following the close of Laramide activity in early middle Eocene, volcanic activity became widespread throughout much of West Texas. Several different eruptive centers, including the Chisos, Davis, and Chinati Mountains (fig. 2), produced thick sequences of lava flow and ash-flow tuff. Between eruptive centers, thick sequences of air fall and water-laid tuffs accumulated, separated by a few relatively thin ash-flow tuffs and lava flows.

Volcanism continued in West Texas throughout the Oligocene and locally into the Miocene. Most of the volcanic rocks are silicic and alkalic (Barker, 1977). Peralkaline rocks predominate along an eastern trend from the Chisos Mountains through the Davis Mountains, and a metaluminous trend of rocks predominates from the Bofecillos Mountains through the Chinati Mountains to the Sierra Vieja and the Van Horn Mountains (fig. 2). Most of the youngest volcanic rocks are more basic, although still alkalic. The latest igneous activity which occurred during the Miocene is mafic and includes scattered basalts of the Diablo Plateau, the Rim Rock dikes of the Sierra Vieja (Dasch and others, 1969), and the youngest Rawls flows in the Bofecillos Mountains (McKnight, 1969). There is no evidence for still later igneous activity in Trans-Pecos Texas, although in southern New Mexico, 30 km (20 miles) west of El Paso, there are Pleistocene basalt cinder zones, flows, and maars (Hoffer, 1976).

During the middle Tertiary, volcanism was rare in Mexico immediately west of Trans-Pecos Texas. A thick sequence of tuffs and lavas evidently derived from a major caldera appears south of the Bofecillos Mountains in the Sierra Rica (fig. 2). Northward along the Rio Grande in Mexico, rare igneous rocks are probably derived mostly from eruptive centers in Texas. Volcanic rocks approximately 30 km north of Ojinaga are derived from and correlative with volcanic rocks in Texas, but some may have originated locally (Heiken, 1966; Gries, 1970).

Although faulting had occurred earlier, major Basin-and-Range-style block faulting began during the late Oligocene/



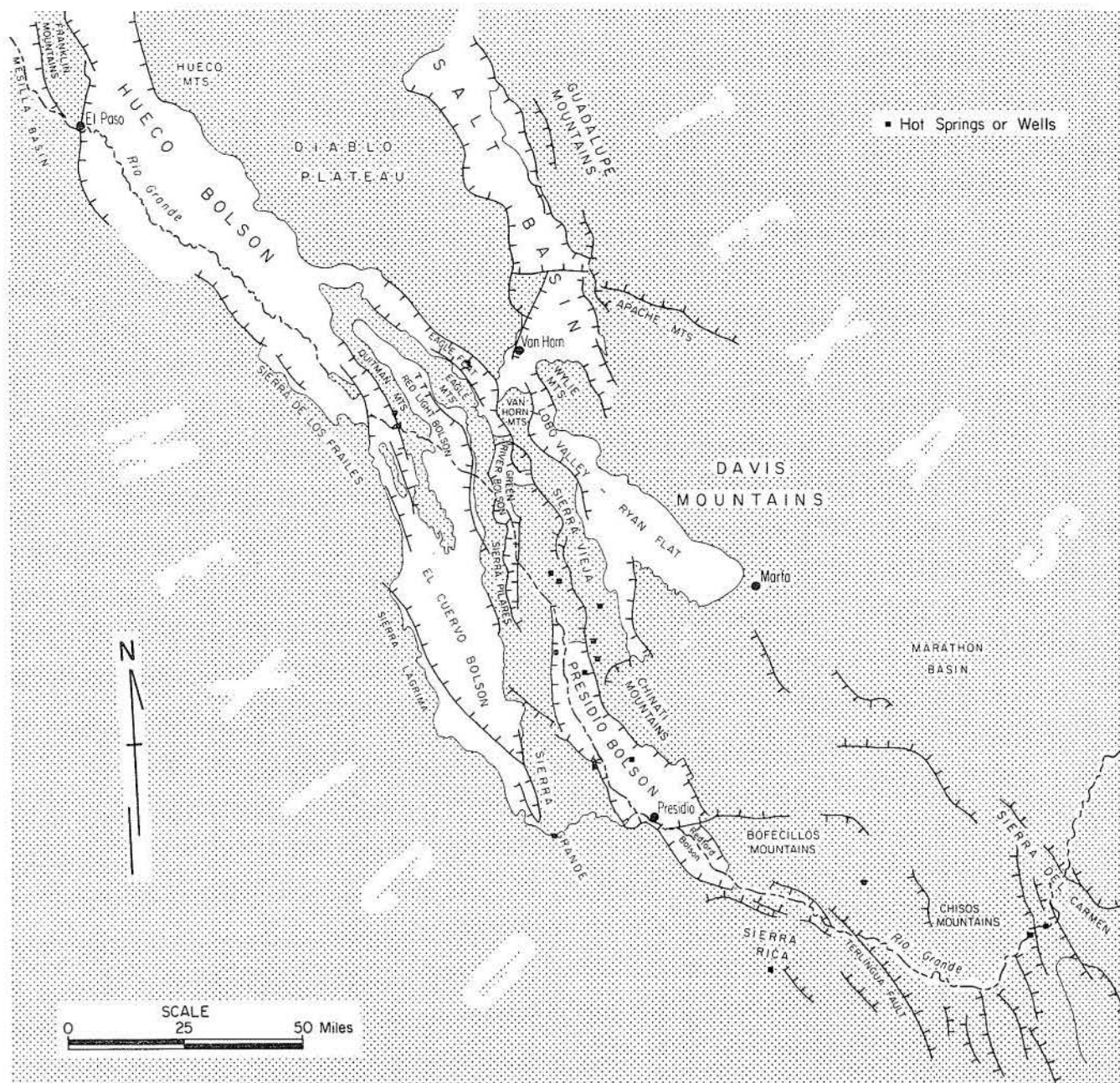


Figure 2. Basins and ranges and major normal faults, Trans-Pecos Texas and adjacent Mexico.

early Miocene (Stevens, 1969). Faulting produced a series of north- and northwest-trending mountain blocks and basins (fig. 2) which filled with debris shed off the mountains. Basins had formed and sediments accumulated in them no later than Miocene time. Stevens (1969) found Miocene mammalian fossils in basin-fill remnants in Big Bend Park. The Rim Rock dikes, a dike swarm intruded along some of the early Basin and Range Faults, are 17 million years old (Miocene), according to isotopic potassium-argon ages (Dasch and others, 1969). Paleontologic and isotopic information is thus consistent on the age of initial faulting. In the Bofecillos Mountains, the youngest Rawls basalts are intruded into or deposited on top of

sediments eroded from older Rawls flows and deposited in newly created fault-bounded basins (McKnight, 1969).

Apparently, older structures controlled some of the faults. Besides the general parallelism between Laramide and Basin and Range structures, the Rim Rock Fault bounding the west side of the Sierra Vieja follows the approximate boundary between the stable Diablo Platform and the highly deformed Chihuahua Trough.

Total offset of about 1,200 m (4,000 ft) occurred on the Rim Rock Fault and the Neal and Mayfield Faults, which bound the east side of the Sierra Vieja and Van Horn Mountains (Twiss, 1959). Similar or greater amounts of displacement occurred on the West Chinati Fault zone, the

Palo Pegado and Cipres Faults bounding Presidio Bolson in Mexico, and the Terlingua Fault and Sierra del Carmen Fault zone in the Big Bend region of Texas and Mexico (fig. 2). Many fault-bounded basins appear to be asymmetrical, with the deepest parts on the west side, including parts of Hueco Bolson east of the Franklin Mountains, Salt Basin near Cornudas, Lobo Valley south of Van Horn, and the northern part of Presidio Bolson.

Basins created by faulting have been accumulating sediments shed off the adjacent highlands since early Miocene (Stevens, 1969). There are as much as 2,750 m (9,000 ft) of sedimentary fill in Hueco Bolson east of El Paso (Gates and Stanley, 1976), but Presidio Bolson and Lobo Valley probably do not contain more than 900 to 1,370 m (3,000 to 4,500 ft) of fill. Salt Basin (fig. 2) is relatively shallow; although there is a maximum of 750 m (2,500 ft) of fill, in general there is much less (White and others, 1977).

Most basins (bolsons) are or were closed until integration of the Rio Grande drainage during late Pleistocene (Strain, 1970). Basins along the present Rio Grande, including Hueco, Red Light, Presidio, and Redford Bolsons, are currently being dissected (fig. 2). Lobo Valley and Salt Basin are still undissected and closed. Lobo Valley drains into Salt Basin, which contains several playa lakes in its lowest parts. Bedrock crops out throughout most of Big Bend, although the area is structurally low. Mostly dissected remnants of fill occur along the west side of the Big Bend adjacent to the Terlingua Fault (Stevens, 1969). Fill along the Sierra del Carmen in Mexico may be considerably thicker, but information on the depth of this basin is scarce.

Normal fault movement has continued to the present in a number of areas (Belcher and Goetz, 1977). Quaternary fault scarps are concentrated along the west side of Salt

Basin and Lobo Valley (Belcher and Goetz, 1977), along the west side of the Eagle Mountains (Underwood, 1963), and along both sides of Presidio Bolson (Gries, 1970, and this study). Scarps bordering the Quitman Mountains are not evident, but Chan and others (1977) identified recent epicenters in this area. Recent fault scarps are also not evident in Big Bend Park, but the lack of extensive Quaternary gravel surfaces makes fault identification there difficult.

Whether or not the Rio Grande Rift extends into Trans-Pecos Texas or Chihuahua is uncertain (Chapin, 1971), although Hueco and Mesilla Bolsons east and west of El Paso are considered part of the rift. Many geologic features of the graben system of Trans-Pecos Texas south of Hueco Bolson are similar to those of the Rio Grande Rift; however, they are also similar to Basin and Range structures. In fact, there may be no difference between the Rio Grande Rift and the southern Basin and Range, although the two areas can be differentiated by gravity and heat-flow data (Decker and Smithson, 1977). Bedrock relief across basin margins in the rift is as much as 11,000 m (36,000 ft) in the San Luis Valley of southern Colorado (Chapin, 1971). This relief is considerably greater than that of any of the basins in Trans-Pecos Texas, where maximum displacement probably does not exceed 2,000 m. No rift-related volcanism occurs in Texas.

If the Rio Grande Rift extended south from El Paso, it would intersect the Chihuahua Trough. Gries (1977) suggested that extension below the evaporite sequence would lead to flowage in the evaporites. Surface expression of extension would range from no visible indication to "secondary faults parallel to but not necessarily directly overlying the basement faults" (Gries, 1977). Heat-flow evidence related to the extension of the Rio Grande Rift is discussed in the following section.

## SOURCE OF HEAT TO THE THERMAL SYSTEMS

Three sources of heat for hot springs that have been proposed for other thermal systems are (1) shallow (less than 6 km), young (but not necessarily molten) magma chambers; (2) the normal thermal gradient of an area where ground water circulates deeply in a convection system; and (3) abnormally high thermal gradient due to blockage of normal heat flow by rock layers of low thermal conductivity.

Each of these sources of heat requires "deep" circulation of water. They differ in ultimate heat source—magmatic, normal thermal gradient, or enhanced thermal gradient. The word "deep" in this study does not have a precise quantitative meaning. Maximum depths of ground-water circulation, even in a tectonically active area, are probably less than 10 km. Ground-water circulation to a depth of only a few hundred meters is inadequate to produce the temperatures of hot spring waters of Trans-Pecos Texas. "Deep," then, should be interpreted as circulation to depths of approximately several hundred meters to several kilometers. More precise estimates of circulation depths are given in following sections of this report.

A shallow magma chamber is the source of heat at Long Valley, California; at Yellowstone Park; at the Jemez Caldera in northern New Mexico; and at The Geysers, California, which is the site for the only commercial geothermal-power-producing facility in the United States. Magmas, however, are unlikely sources of heat for the Texas thermal systems, because exposed igneous rocks are at least as old as Miocene; even deeply buried rocks of this age would have lost their initial heat by now. Pleistocene igneous activity occurred in southern New Mexico at the Potrillo basalt field just west of El Paso as recently as 125,000 years ago (Hoffer, 1976). Further north along the Rio Grande Rift, the Carrizo basalts are less than 10,000 years old (Smith and Shaw, 1975). Similar rocks are not exposed in Texas. Furthermore, basaltic magmas normally travel directly from the point of generation to the surface without creating shallow magma chambers as do more silicic magmas. Magmas associated with the four geothermal systems listed above are rhyolites or dacites. Finally, because the hot springs in Texas are widespread, numerous magma chambers would have to exist. If recent



magmas were so common, there should be additional evidence for their existence besides hot springs.

Ground-water circulation to depths of at least 1 km is more likely than magmas as a source of heat. Even with a magmatic source, meteoric ground water must circulate relatively deeply to be heated. The association of hot springs with normal faults shows that conduits do exist for deep circulation. However, investigations have not determined the depth necessary to produce either the measured surface temperatures or the subsurface temperatures inferred from geothermometry. The setting of Trans-Pecos Texas in the Basin and Range province or possibly in the Rio Grande Rift implies that heat flow and thermal gradients in Texas may be considerably greater than continental averages. However, an abnormal gradient is not required to produce hot springs because they can be formed as a result of sufficiently deep ground-water circulation in an area of normal heat flow (for example, Hot Springs, Arkansas) (Bedinger and others, 1974).

Blockage of vertical heat flow by low conductivity layers, the third source of heat, can produce abnormal thermal gradients in an area of otherwise normal heat flow. Rock temperatures directly below a low-conductivity layer are elevated as a function of the thickness of the layer, the difference in conductivity between the rock layers (conductivity contrast), and the normal heat flow. Theoretically, temperature differentials of 100°C can exist for layers 2 km thick in an area of high heat flow (Diment and others, 1975).

Water has low thermal conductivity. Poorly consolidated, water-saturated sediments, such as those that fill the late Tertiary basins of West Texas, exhibit low thermal conductivity. Some highly porous volcanic rocks such as ash-flow tuffs also have low thermal conductivity. In the Basin and Range province in Nevada, thermal gradients are as high as 90°C/km in basins composed of low-conductivity sediments (Hose and Taylor, 1974). Thermal gradients are only approximately 35°C/km in the ranges, even though total heat flow in the basins and ranges is similar. Basins in Texas could have similar high thermal gradients.

The latter two heat sources depend on regional heat flow. Heat flow is commonly reported in two ways: unreduced and reduced. Unreduced heat flow equals total heat flow, whereas reduced heat flow equals the total minus a component from radioactive heat generation in the upper crust. Unless otherwise noted, heat flow presented in this study is unreduced. Reduced, or mantle and lower crustal heat flow, is about 0.8 heat flow unit (HFU) (1 HFU =  $1 \times 10^6$  cal/cm<sup>2</sup>sec) in the eastern United States, and at least 1.4 HFU in the Basin and Range (Diment and others, 1975). The eastern United States is probably typical of stable continental regions (Roy and others, 1968). Heat flow east of the Basin and Range in Texas is about 1.1 HFU (fig. 3).

Comprehensive heat-flow information for the Trans-Pecos Texas area is scarce. Conclusions must be drawn from measurements in adjacent areas or from analogy to areas of similar geologic setting. The Basin and Range province is characterized by relatively high heat flow (1.5 HFU) with anomalies of over 3 HFU (Roy and others, 1972). The average is probably between 1.5 and 1.9. The Rio Grande Rift in New Mexico displays even higher heat flow (Reiter and others, 1975), with an average of over 2.5 HFU along

the western edge of the rift (fig. 3). Heat-flow measurements in the southern rift range from 2 to 3.6 HFU (Decker and Smithson, 1975). These values suggest the possibility of high heat flow in the Basin and Range of Trans-Pecos Texas and very high values along the Rio Grande in Texas and Mexico. These values contrast sharply with heat flow in the Great Plains province roughly east of the Pecos River including eastern New Mexico and Texas. Measurements in southeastern New Mexico and near Pecos, Texas (Herrin and Clark, 1956), fall in a narrow range from 0.9 to 1.3 HFU, with an average of 1.1.

Smith (1977) presented several heat-flow values in Chihuahua and Coahuila, but these values do not help clarify the question of the location of the Rio Grande Rift south of El Paso. One value, 50 km south of El Paso, is only 1.2 HFU (fig. 3). Values of 2.5 in Chihuahua City and 1.8 in western Coahuila indicate higher heat flow in a southeastern trend from New Mexico at least parallel to (but not necessarily coincident with) the Rio Grande.

Decker and Smithson (1975) determined two heat-flow values in Texas (fig. 3). One determination near Van Horn in the Basin and Range physiographic province is 1 HFU, below the Basin and Range average; the well used has a thermal gradient of only 14°C/km. A second determination near Shafter just east of Presidio Bolson is 1.5 HFU (1.2 HFU reduced). This value is much lower than the Rio Grande Rift values, is even below most Basin and Range values, and is not markedly greater than Great Plains values. The Shafter well was considerably deeper than the wells in southern New Mexico and varied significantly in rock conductivity and thermal gradient: in only the upper 560 m (which is deeper than any of the New Mexico wells) calculated heat flow is 2.1 with a thermal gradient of 25°C/km.

Kleeman (1977) assumed conductivities for well data reported by Gates and White (1976) to determine heat-flow values of 1.4 to 3.0. Thermal gradients for the wells range from 27° to 50°C/km. All of these wells were tests for water in basins where blockage of normal heat flow by low-conductivity, unconsolidated sediments might be expected to increase the thermal gradient. As an example, Guerra No. 1 (heat flow = 3.0 on Kleeman's map but listed as 2.4) penetrates to the base of the sediments, or at least to more consolidated or well-cemented sediments. The temperature profile shows a sharp increase at this point, possibly because of a change in conductivity. An alternative explanation for the high thermal gradient could be ground-water circulation.

The Big Bend region falls within the Basin and Range structural and geologic province (Fenneman, 1946). However, heat-flow determinations by Swanberg and Herrin (1976) indicate that the Big Bend region is part of the Great Plains normal heat-flow province. Measurements taken above the water table indicate heat flow of 2.3 to 4.3 HFU caused by abnormally warm water at the water table. A single deep well that penetrates the water table gives a heat flow of 1.3 HFU. Swanberg and Herrin (1976) considered only the latter to be a reliable indication of regional heat flow. If their judgment is correct, the physiographic and heat-flow boundaries of the Basin and Range do not coincide. Smith (1977) reported a value of

1.3 HFU at La Linda, slightly southeast of Big Bend Park; this information supports Swanberg and Herrin's interpretation.

The thermal gradient map of the American Association of Petroleum Geologists (AAPG, 1975) displays gradients in degrees Fahrenheit per 100 ft based on deep oil- and gas-well-temperature measurements. Unfortunately, the map shows few oil tests in Trans-Pecos Texas and almost none with thermal-gradient measurements in the Rio Grande Valley. The map does show generally increasing gradients from the midcontinent of West Texas, with values of approximately  $15^{\circ}$  to  $18^{\circ}\text{C/km}$  ( $0.8^{\circ}$  to  $1^{\circ}\text{F/100 ft}$ ), to Trans-Pecos Texas, with values of approximately  $18^{\circ}$  to  $26^{\circ}\text{C/km}$  ( $1^{\circ}$  to  $1.4^{\circ}\text{F/100 ft}$ ). Collins (1925) listed a normal thermal gradient for the Trans-Pecos area at about  $29^{\circ}\text{C/km}$  ( $1.6^{\circ}\text{F/100 ft}$ ). Unpublished temperature-log data from deep-oil tests along the Rio Grande show gradients as high as  $40^{\circ}\text{C/km}$ . Together, the data show that the Trans-Pecos region is distinct from the Great Plains and that thermal gradients are highest along the Rio Grande.

The presence of hot springs is an additional, although qualitative, indicator of heat flow and thermal gradient. Hot springs in the United States are not restricted to areas of high heat flow but are certainly concentrated in areas such as the Basin and Range province or in areas of recent igneous activity. Within the Basin and Range, hot spring activity seems to reflect heat flow. For example, Sass and others (1971) noted that an area of relatively low heat flow (low for the Basin and Range) in central Nevada has fewer, cooler hot springs than surrounding areas. With the exception of the springs in and around Big Bend Park, all the hot springs in the study area are along the nearly linear southeast-trending section of the Rio Grande from El Paso to Presidio. The presence of the hot springs in this section and the high temperatures for some of them indicate that the area has high heat flow—even higher than that of adjacent areas to the east.

This evidence indicates that Trans-Pecos Texas is a transitional region between the midcontinent normal thermal gradient and heat-flow province and the Basin and Range heat-flow province. Not all of Trans-Pecos Texas, which is physiographically part of the Basin and Range, is characterized by Basin and Range heat flow, although the Rio Grande region of Texas probably does have heat flow similar to that of the Basin and Range. However, the presence in Texas of extremely high heat flow, as is characteristic of the Rio Grande Rift in southern New Mexico, remains uncertain even though the data of Reiter and others (1975) and Decker and Smithson (1975) suggest that the rift may continue into Texas.

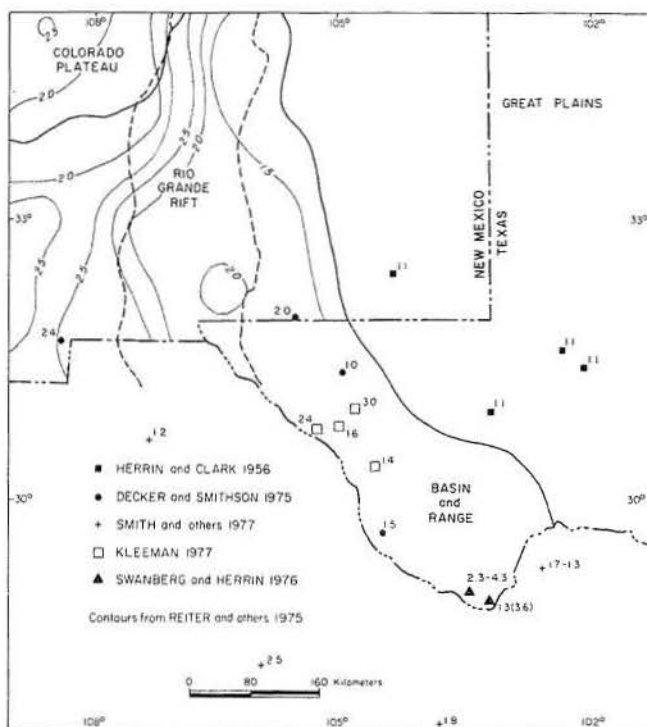


Figure 3. Heat-flow determinations in Trans-Pecos Texas and adjacent regions. Physiographic provinces from Fenneman (1946).

The source of a thermal anomaly expressed by high heat flow in the Basin and Range and the Rio Grande Rift is generally attributed to the presence in these areas of a thin crust and a shallow, high-temperature mantle (Roy and others, 1968). Late Tertiary extension was taken up by normal faults in the upper crust and by plastic thinning of the lower crust (Woodward, 1977). Crustal thicknesses are around 20 to 30 km in the Basin and Range (Healy and Warren, 1969), similar to crustal thicknesses in the rift, and about 50 km for the Great Plains (Pakiser, 1963).

An alternative explanation of the high heat flow requires the existence of a deep-seated intrusion of magma at a depth of approximately 20 km along crustal fractures extending into the mantle. Sanford and others (1973) present evidence for such a magma chamber at a depth of 18 km beneath Socorro, New Mexico. Reiter and others (1975) suggest that a series of magma chambers beneath the rift would result in areas of alternating high and low heat flow. Present data suggest a relatively continuous source of heat, but refinement of this data might reveal more discontinuity.

## GEOLOGIC AND HYDROLOGIC SETTING OF HOT SPRINGS AND WELLS

### Introduction

Chemical studies (White, 1970) and stable-isotope studies (Craig, 1963) have provided conclusive evidence that hot spring waters are derived almost entirely from local meteoric water. Stable-isotope analyses of thermal and nonthermal waters in this study demonstrate the meteoric origin of hot spring waters in Trans-Pecos Texas. The

previous section presented evidence that the source of heat for those thermal waters is deep circulation of the water in a region of relatively high thermal gradient. The origin of hot springs, then, requires that meteoric water circulate to a source of heat at an unspecified depth, be heated, and return to the surface. This pattern is clearly seen in the geologic setting of hot springs in Texas.



The summary by White (1968) of the geometry and hydrodynamics of hot springs systems is useful in understanding the origin and flow paths of springs in Trans-Pecos Texas. Two factors are responsible for circulation of hot water: one, which acts on all springs, is the difference in elevation between areas of recharge and areas of discharge; the second is the difference in density between cold recharge water and hot discharge water. As indicated by White (1968), the density of water is a function of temperature, pressure, and total dissolved solids. In thermal systems in Trans-Pecos Texas, the density difference caused by temperature differences is greater than differences caused by other density factors. Thus, a column of cold water can support a taller column of hot water, and under certain conditions the recharge area for a hot spring can be below the spring itself. The potential difference in elevation of recharge and discharge is a function of temperature contrast in the two columns and depth of circulation.

### ***Definition of Hot Springs***

Springs in the Rio Grande region of West Texas and Mexico range in temperature from 21°C (70°F), which is the approximate average annual temperature for the region, to 90°C (194°F). Springs with surface temperatures approximately 8°C (15°F) above mean annual temperatures are hot or thermal springs, according to Waring (1965). Thus, thermal springs in West Texas are those hotter than 30°C (86°F). This definition excludes springs which discharge waters of thermal origin but which have cooled below 30°C (86°F) because of mixing with nonthermal ground water or because of conductive cooling during slow discharge. These springs could be recognized by other criteria, such as their chemical composition. An example discussed below is Soda Spring in the Indian Hot Springs group.

Wells cannot be classified using the criterion of Waring (1965) because a sufficiently deep well in any region will tap hot water. For purposes of this study, a hot well produces water which is abnormally hot for that well depth and for the normal thermal gradient of the area. Because the thermal gradient in Trans-Pecos Texas is not well known, this criterion does not provide a clear distinction. Knowledge of even the normal thermal gradient is important to the study of geothermal energy; accordingly, all deep wells for which information could be obtained have been examined.

### ***Geologic Setting of Hot Springs and Wells***

Hot springs occur in three distinct areas within Trans-Pecos Texas: the southern Hueco Bolson adjacent to the Quitman Mountains, Presidio Bolson and its structural extension to the north, and the Big Bend region (fig. 2). There are probably more hot springs in this area than are discussed in this report, particularly in some of the less accessible parts of Mexico. Wells that tap anomalously hot water exist in all three areas and on Eagle Flat (fig. 2).

## **SOUTHERN HUECO BOLSON**

### **Indian Hot Springs**

Hot springs in Hueco Bolson are all at the southernmost end (fig. 2). Indian Hot Springs lies at the southwest end of the Quitman Mountains on the floodplain of the Rio Grande (fig. 4). The area has been mapped by Jones and Reaser (1970) and by Reaser (1974). When measured in September 1976, there were at least seven springs which ranged in temperature from 27°C (81°F) at Soda Spring to 47°C (117°F) at Stump Spring. In 1968, Jones reported temperatures as high as 52°C (126°F). Dorfman and Kehle (1974) reported an unconfirmed temperature of greater than 60°C (140°F) for a shallow well. When visited in 1976, Dynamite Spring and Masins Spring were not active. Dynamite Spring now is only a shallow well tapping Rio Grande alluvial water and has no discernible discharge. Jones (1968) reported that a major flood in 1962 buried several springs.

Active springs include Chief, Squaw, Stump, Beauty, and Soda Springs. All except Soda Spring emanate from an extensive travertine plateau deposited by the springs. Chief, Squaw, and Stump have precipitated travertine mounds approximately 0.5 m above the plateau (fig. 5). Rock and wood "bathhouses" have been built around Chief and Squaw Springs, whereas rock "tubs" are built around the others. Only Stump discharges to the surface; the others discharge through permeable travertine or alluvium below the travertine and flow to the Rio Grande. Total discharge is at least 400 liters per minute (l/min) and probably considerably greater because precise estimates are difficult to make.

Bell (1963) and Jones (1968) suggested that most of the springs are actually shallow wells dug into floodplain alluvium. If so, the travertine mounds have been deposited since the wells were dug. More likely, the original springs have simply been enhanced by resort owners.

Soda Spring emanates from an arroyo just north of the resort area (fig. 4). When measured in 1976, the water temperature was 27°C (81°F), below the required temperature for a thermal spring. An extensive travertine deposit has built up at Soda Spring, and the water is highly mineralized. Comparison of chemical analyses (table 4) shows that water discharged from Soda Spring is a mixture of thermal water and water in the bed of the arroyo.

Indian Hot Springs appears just basinward of the Caballo Fault, a major northwest-trending normal fault dividing the Quitman Mountains and Hueco Bolson. The fault separates Cretaceous sediments on the northeast from downdropped bolson fill on the southwest (figs. 4 and 6). The fault trace follows a narrow bench along the range front 6 m (20 ft) above the travertine plateau (fig. 6). Bolson sediments exposed in roadcut along this bench are cemented with calcite. A cap of travertine that overlies the sediments may be from earlier spring activity at the higher elevation of the bench. Subsequently, downcutting of the Rio Grande lowered the hydrologic base level; the thermal waters now emanate from a lower level. Continuation of spring activity during downcutting of the Rio Grande suggests that the springs and thermal circulation are as old as some of the terraces.

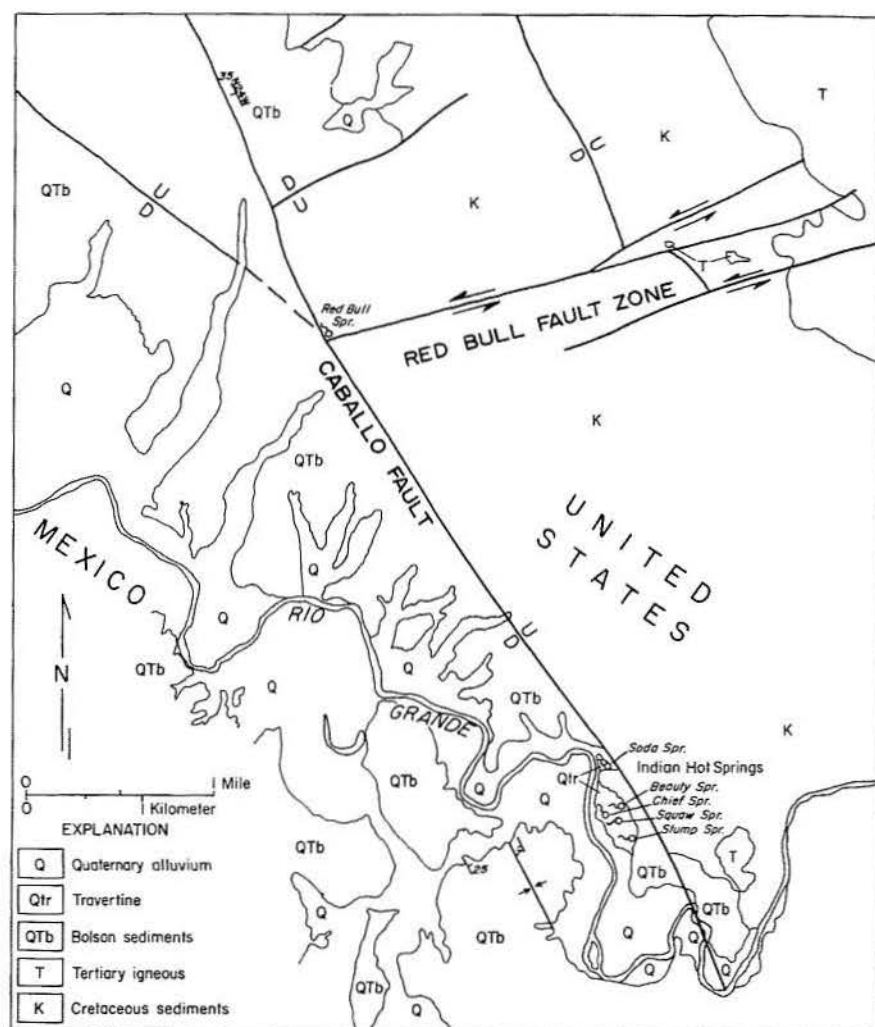


Figure 4. Geologic map of Indian Hot Springs and Red Bull Springs area, southern Hueco Bolson, Texas. Geology from Jones and Reaser (1970) and this report.

Movement on the Caballo Fault started in early Miocene (Jones and Reaser, 1970); the fault may still be active. Bolson sediments are brecciated and strongly tilted from drag along the trace of the fault. Jones and Reaser (1970) show that the Caballo Fault does not displace Quaternary terrace gravels approximately 2 km northwest of Indian Hot Springs. However, at that location the terrace gravels abut against but do not continue across the fault; thus the age of most recent displacement cannot be determined. Chan and others (1977) reported a magnitude 2 to 3 earthquake epicenter within about 20 km of Indian Hot Springs. Uncertainty about the precise epicenter location prohibits correlation with any known fault, but the occurrence of the earthquake demonstrates that faults in the area are still active.

#### Red Bull Spring

Red Bull Spring (fig. 4) discharges 37°C water approximately 50 l/min from fractures in red calcareous claystone of the Cretaceous Mountain Formation approximately 5 m updip from the fault trace. Jones and Reaser (1970) reported from local ranchers that before a major earthquake

in 1931 the discharge of Red Bull Spring was considerably greater than it is now.

The spring lies at the intersection of at least two and possibly three faults (fig. 4). The major fault is the northwest-trending Caballo Fault. Both claystone and bolson sediments near the fault are highly fractured and cut by numerous calcite veinlets. Bolson sediments dip as much as 60° southwest away from the fault, apparently because of drag along the fault. The east-trending, left lateral Red Bull fault zone terminates against the Caballo Fault at the spring. Although not as evident on the ground, a lineation on aerial photographs may represent a third fault cutting bolson fill west of the Caballo Fault. The fault trends directly towards Red Bull Spring, but the fault cannot be traced all the way to the spring. An indistinct scarp with a down-dropped southwest side follows the fault trend for about 0.5 km.

Jones and Reaser (1970) suggest two episodes of movement on the Red Bull fault zone: a post-Laramide period of left lateral displacement and a late Tertiary episode of normal faulting. The inferred fault presumably cuts bolson fill and must result from normal faulting during the late Tertiary.



Figure 5. View to southwest of Indian Hot Springs. Flat area in middle distance is underlain by travertine deposited by spring waters. Chief Spring (center, inside bathhouse on left) has built up a low travertine mound above plateau. Tilted bolson sediments form cliff face on far side of Rio Grande in Mexico. Photograph taken from approximate trace of Caballo Fault.



Figure 6. View north along Caballo Fault above Indian Hot Springs. Fault runs diagonally across photograph from lower right to upper left along narrow bench below rounded hills. Road cut exposes calcite-cemented and travertine-capped bolson sediments. Flat area to left is travertine plateau of active springs.



## PRESIDIO GRABEN

Much of the present thermal activity in Trans-Pecos and Mexico occurs within the structural low that includes Presidio Bolson (figs. 2, 7, 8). Several springs occur along major boundary faults; others occur on minor faults within the basin, and several springs lie on faults in bedrock north of the bolson.

### Capote Springs

Capote Springs lies in volcanic rocks approximately 15 km north of the sediment-filled portion of Presidio Bolson (figs. 7 and 8) in an area mapped by Buongiorno (1955) and Twiss (unpublished). The springs appear in the wall of a canyon about 20 m above the bed of Capote Creek where the creek is cut by one of a series of minor normal faults. The springs discharge at a temperature of 37°C (98°F) from numerous fractures in a rhyolite lava flow on the downdropped side of the fault (fig. 9). Total discharge is approximately 400 l/min according to estimates of stream flow below the springs. There are no spring deposits.

The fault is one of several normal faults that trend approximately north and dip steeply to the west. Displacement is about 50 m down to the west. Movement on this fault system started in late Oligocene or early Miocene (Dasch and others, 1969). Evidence of more recent movement is not available because the fault is entirely within Tertiary volcanic rocks.

### Nixon Spring

Nixon Spring, approximately 8 km southwest of Capote Springs (figs. 7 and 8), appears on a minor branch of the Candelaria Fault in an area mapped by Buongiorno (1955) and Twiss (unpublished). The spring consists of several small seeps from colluvium in a narrow canyon eroded along the fault scarp. Total discharge is only a few l/min; maximum temperature is 32°C (approximately 90°F). No deposits are evident other than some unidentified powdery salts probably deposited by evaporation of the spring water.

The fault cuts Tertiary volcanic rocks, trends approximately north, and is down to the west. At the spring location, tuffaceous sediments appear on both sides of the fault. Total offset is only a few tens of meters here but increases to the north. Because the fault is entirely within Tertiary rocks, the most recent movement cannot be determined. The Candelaria Fault, however, offsets older Quaternary terrace gravels with as much as 20 m of displacement.

### Hot Springs - Ruidosa

Hot Springs lies at the northeast corner of Presidio Bolson approximately 10 km northeast of Ruidosa (figs. 7 and 8). Water at a temperature of 45°C (113°F) emanates from a concrete enclosure on a terrace about 3 m (10 ft) above the bed of Hot Springs Creek. Natural discharge was

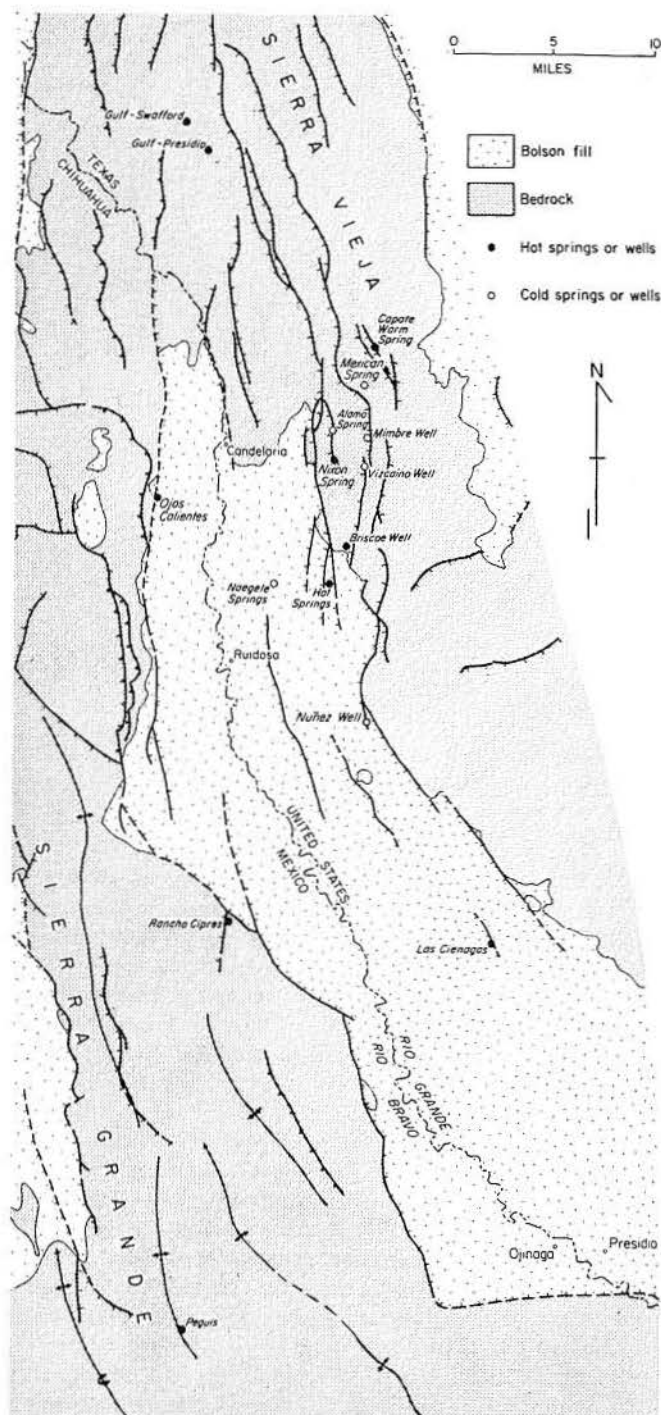


Figure 7. Presidio Bolson and location of sampled hot and cold springs and wells.

evidently from gravels along a small bluff overlooking the terrace. The gravels are uncemented, unlike bolson sediments in the area, and are probably Quaternary terrace gravels. No spring deposits are apparent, and it is unlikely that recent human disturbance could have completely removed them. Discharge is about 75 l/min.

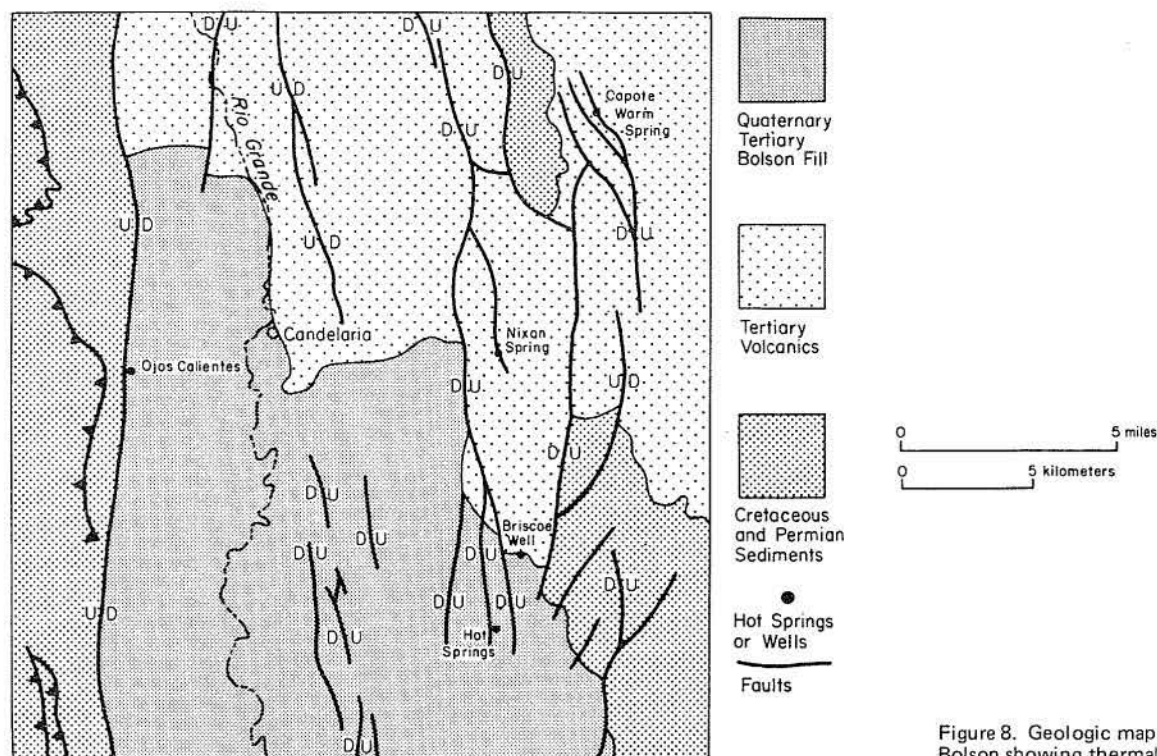


Figure 8. Geologic map of northern Presidio Bolson showing thermal springs and wells.

The area around Hot Springs has been mapped by Amsbury (1958), Dickerson (1966), and Groat (1972). All exposed rocks within approximately 2 km of the spring are basin fill, conglomerate, and coarse sandstone of basin-margin facies (Groat, 1972). Hot Springs lies between two branches of the Candelaria Fault (fig. 8). The faults are well exposed in Hot Springs Creek about 200 m downstream and 300 m upstream from the spring. Both faults are down to the basin, according to observable offset on the faults and drag folding of the bolson sediments. The faults cannot be followed south of Hot Springs Creek on the ground or on aerial photographs. Recent fault movement is not evident near the spring. As discussed for Nixon Spring, the Candelaria Fault shows considerable Quaternary displacement. No faults were observed at Hot Springs, but exposures of bolson sediments are not adequate to discount completely the presence of a fault.

Bedrock may be fairly shallow in the spring area. Cretaceous limestone crops out approximately 2 km to the east but on the upthrown side of two normal faults, both of which are down towards the springs. To the north, Tertiary volcanic rocks crop out in the same fault block as the spring and plunge gradually towards the springs.

#### Las Cienagas

Las Cienagas (the bogs) is a group of springs and seeps on the east side of Presidio Bolson (fig. 7). Maximum temperature of the springs is 30°C (86°F). Many are near or at average annual air temperature but are probably cooled thermal waters. Total discharge of Las Cienagas, which was determined from creek-flow measurement below the springs,

is approximately 1,000 l/min. However, some discharge probably percolates into the ground, and some of the creek water may not be hot-spring discharge. Consequently, creek flow is only an approximation of spring discharge. Soils around the springs are loosely cemented with calcite or travertine that may have been deposited by the spring.

A detailed geologic map constructed for this study is presented in figure 10. Previous mapping of the spring area was by Rix (1953) and Groat (1972). The springs discharge at the base of a rhyolite knob exposed in the side of a hill cut in fine-grained basin-fill sediments and capped by Quaternary terrace gravels. The springs discharge from colluvium, but the spring location is probably controlled by fractures in the rhyolite where the rock is surrounded by relatively impermeable bolson sediments. The rhyolite is probably a shallow intrusion related to the Chinati Mountains caldera (Cepeda, 1977). The rhyolite is older than the bolson sediments that were deposited around it. A major normal fault bounds the Presidio Graben approximately 2 km to the northeast. Las Cienagas occurs in the downthrown block (fig. 7). The presence of the rhyolite hill at Las Cienagas and similar erosional remnants nearby indicates that bedrock is shallow throughout the area. Just below the springs, several small normal faults cut the bolson. These faults trend N. 30° W. and are downthrown to the west. Bolson sediments are folded, presumably by drag along the faults. Observable offset along these faults is minor. Total displacement of either bolson or the prebolson basement is unknown.

Bolson sediments just above the springs are sandstone facies, and sediments at and below the springs are mudstone facies, both described by Groat (1972). Presence of

impermeable clay accounts for the discharge. In coarser, more permeable sediments the thermal waters could discharge into the subsurface. The change in bolson lithology could be interpreted as a fault; however, no fault plane is exposed, and the sediments are undeformed. The difference in sediment composition probably represents a change in depositional facies within the bolson.

Timing of displacement on both the basin-edge fault and the fault below Las Cienagas can be identified only as post-bolson in age. Faults do not cut late Quaternary terrace gravels. However, less than 15 km northwest, in an equivalent position near the edge of the basin, several normal faults do cut youngest Quaternary terrace gravels with as much as 5 m of displacement down to the west.

### Ojos Calientes

Ojos Calientes, the hottest springs observed during this study, are located on the Mexican side of Presidio Bolson approximately 7 km southwest of Candelaria (figs. 7 and 8). The area was mapped by Haenggi (1966) who speculated that the thermal waters arose along the Palo Pegado Fault, a major normal fault which bounds the west side of the Presidio Graben. The springs appear in a zone 200 to 600 m basinward of the Palo Pegado Fault. There are no large single springs but innumerable small springs and seeps which emerge from bolson and Quaternary gravels in the bed of an arroyo. Measured water temperatures range from approximately 60°C (140°F) to a maximum of 90°C (194°F). The lower temperature water may result from mixing of thermal water with nonthermal ground water or from shallow cooling of water of low-discharge springs.

Several springs continuously spurt to a height of approximately 0.5 m like miniature geysers. The driving force for those springs may be degassing of CO<sub>2</sub> dissolved in the thermal water rather than steam pressure. From the flow of the arroyo below the springs, total discharge of all the springs is estimated as more than 1,000 l/min.

The springs have built up extensive travertine deposits, including many small knobs and several broad mounds as high as 3 m above the arroyo bottom (figs. 11 and 12). Discharge occurs from the tops and sides of the mounds. Rate of discharge changed significantly between October 1976 and February 1977 when several former discharge areas were reduced in flow or dried up entirely.

The trace of the Palo Pegado Fault crosses the arroyo above the springs but is not exposed until the next arroyo, 2 km to the south, where it is well exposed. The fault trends N. 15° E., dips 55° to the east, and displaces bolson gravels against highly sheared Cretaceous shales. Bolson sediments are turned up along the fault and dip as much as 20° to the east. Within 100 m, bolson beds return to the normal 3° western dip. The fault zone is filled with finegrained, massive calcite which could have been deposited by thermal water.

Haenggi (1966) estimated 900 m of displacement of the Palo Pegado Fault in an area approximately 10 km north of the springs where Cretaceous sediments appear on both sides of the fault. Displacement increases toward the springs, so 900 m is a minimum estimate of displacement in the springs' vicinity. No other faults appear basinward of the Palo Pegado Fault; all displacement on the west side of Presidio Bolson apparently was taken up by this fault. There is no evidence of recent fault movement in this area.



Figure 9. A part of Capote Springs emerging from fractures in rhyolite lava flow along a minor normal fault.



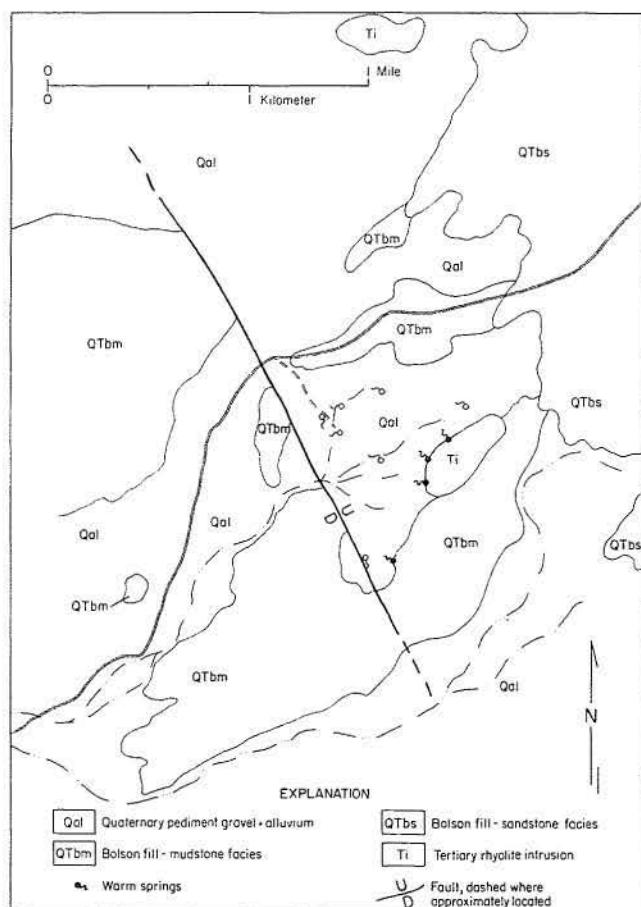


Figure 10. Geologic map of Las Cienagas area, Presidio Bolson, Texas. Geology from Groat (1972) and this report.

However, to the south in an area mapped by Gries (1970), a similar boundary fault cuts Quaternary deposits.

#### Spring at Rancho Cipres

Another hot spring in Presidio Bolson in Mexico is found near Rancho Cipres (fig. 7), approximately 40 km northwest of Presidio and 20 km south of Ruidosa in an area mapped by Heiken (1966) and Gries (1970). A single spring appears near the top of a large, partially dissected travertine mound overlying Tertiary volcanic rock and Cretaceous sandstone and shale. The spring forms a pool approximately 2 m in diameter within a shallow basin at the top of the mound (fig. 13). Temperature in the pool is 35°C (95°F). The pool emits a strong odor of H<sub>2</sub>S and bubbles constantly, probably from degassing of CO<sub>2</sub> and H<sub>2</sub>S dissolved in the thermal water. Water does not flow out of the pool but percolates through permeable travertine and discharges as cool seeps around the edge of the mound. The size of the travertine mound, approximately 700 m by 200 m in area and as thick as 10 m, indicates that the spring must have been much more active in the past.

The travertine deposit straddles a north-trending fault which displaces upthrown Cretaceous rocks on the east against downthrown Tertiary volcanic rocks on the west. The fault is a branch of the Cipres Fault, a major

basin-bounding normal fault on the west side of Presidio Bolson (fig. 7). The presently active pool is almost directly over the fault trace. There is no evidence of post-Tertiary movement on the fault, but to the south the Cipres Fault cuts Quaternary features (Gries, 1970).

#### Briscoe Well

Three wells in the structural trough north of Presidio Bolson also tap hot water. The first is at the Briscoe Ranch where a domestic well penetrated hot water at a depth of 27 m (figs. 7 and 8). The water temperature was 42°C (108°F) when measured in 1976 for this study; White and others (1977) reported a temperature of 51°C (124°F). The well lies at the northern edge of the sediment-filled portion of the basin, approximately 3 km north of Hot Springs - Ruidosa and 1 km east of the Candelaria Fault. A thin cover of bolson gravels crops out at the surface, but volcanic rocks crop out near the well. The water comes either from gravels or from volcanic rocks just beneath the sediments. A 12-m-deep well near the hot well produces only cold water. If the two water sources were hydraulically connected, the hot water should either rise to the top of the water table or at least mix with the cold water. Therefore, the two wells must be hydraulically isolated. Two minor faults mapped by Amsbury (1958) trend toward the well location, but their relation to the well and thermal water is uncertain.

One other well, approximately 4 km southeast of Hot Springs, is reported to produce water at about 34°C (93°F) (White and others, 1977). The location of this well is not precisely known, but it appears to be approximately on the trace of the normal fault bounding the east side of Presidio Bolson. A well in that area is abandoned; nearby wells are windmill driven and pump too slowly for a meaningful temperature measurement.

#### Gulf Wells

Two artesian wells drilled 4 km apart in 1965 by Gulf Oil Corporation produce the hottest water in Trans-Pecos Texas. These wells occur about 30 km north of Presidio Bolson but in the same structurally downdropped block west of the Sierra Vieja (fig. 7). Depth, temperature, and flow data are shown in table 1. Both wells have been plugged back to the water-producing horizon, a massive limestone that is the Georgetown equivalent in the area (table 2). The variation in reported temperatures probably reflects sampling conditions because the water cannot be tested directly at the well head, and air temperature at the time of this study was about 7°C (45°F). Temperature logs for Gulf-Swofford show a temperature inversion below the hot-water-producing horizon. Below that level, the temperature drops and reaches 80°C again only at a depth of about 2,500 m (8,200 ft). Thus the high-temperature water of the shallow producing horizon must be carried from greater depths by thermal convection. From readings of maximum bottom hole temperatures, thermal gradients in the two wells have been determined as 36°C/km for Gulf-Swofford and 43°C/km for Gulf-Presidio. How much thermal convection has altered the gradients from true regional gradients is unknown.



Figure 11. View to northeast across travertine mound deposited by spring waters at Ojos Calientes. Sediments inside of arroyo are Quaternary terrace gravels. In background are ridges of dark volcanic rocks and white volcaniclastic sediments of the Vieja Group in Texas.



Figure 12. Low travertine knobs with active travertine deposition by thermal waters of Ojos Calientes.

Hot water would flow to the surface if permeable channelways were available. That the water does not reach the surface is somewhat surprising because the area is highly faulted. However, Cretaceous formations above the producing horizon are largely impermeable shales that may prevent surface flow.

#### Hot Springs at Peguis

Two hot springs with temperatures of about 36°C (97°F) lie on opposite sides of the Rio Conchos at Peguis, 35 km west of Presidio, where the river flows out of a canyon of the Sierra Grande (fig. 7). The springs are near to but outside of Presidio Bolson. Construction of an irrigation dam has considerably altered the natural setting, but the description of the spring area by Gries (1970) agrees substantially with observations of this study. Discharge is about 1,000 l/min, and much additional water may flow unobserved into the river. The springs do not deposit travertine.

The springs appear unrelated to faulting. They discharge on the eastern limb of the Peguis anticline at the Buda Limestone - Ojinaga Formation contact (fig. 14, table 2). Gries (1970) stated that the springs appeared to issue from fractures in the anticline. The Buda is massive, highly fractured, cavernous limestone. The Ojinaga Formation, although fractured, consists of impermeable flaggy limestone, sandstone, and shale. Thermal waters rise through the permeable Buda and discharge at the water table, which

is the Rio Conchos, at the Buda - Ojinaga contact, a ground-water barrier. Some water could continue through fractures in the Ojinaga, but most would be diverted.

### BIG BEND AREA

#### Big Bend National Park

At least six hot springs lie along the Rio Grande in Big Bend National Park or on the south side of the river in Mexico (R. A. Maxwell, personal communication, 1976) (fig. 15). All were visited, but only three were sampled in this study. Most of the other springs occur in impenetrable cane thickets and were impossible to reach without disturbing the vegetation. Chemical analyses of the three springs sampled are nearly identical and probably representative. The three sampled are Hot Springs (41°C, 105°F), an unnamed spring (40°C, 104°F) surrounded by cane thickets approximately 300 m downstream from Hot Springs, and another unnamed spring (36°C, 97°F) used to supply water for Rio Grande Village (fig. 15). There are no spring deposits.

The spring at the village and all unsampled springs lie on normal faults in the Santa Elena Limestone (table 2). One of the sampled springs discharges at the Buda - Boquillas contact (table 2); the other (Hot Springs) discharges from northeast-trending fractures in the Boquillas Formation near the contact. Presumably, thermal water can move

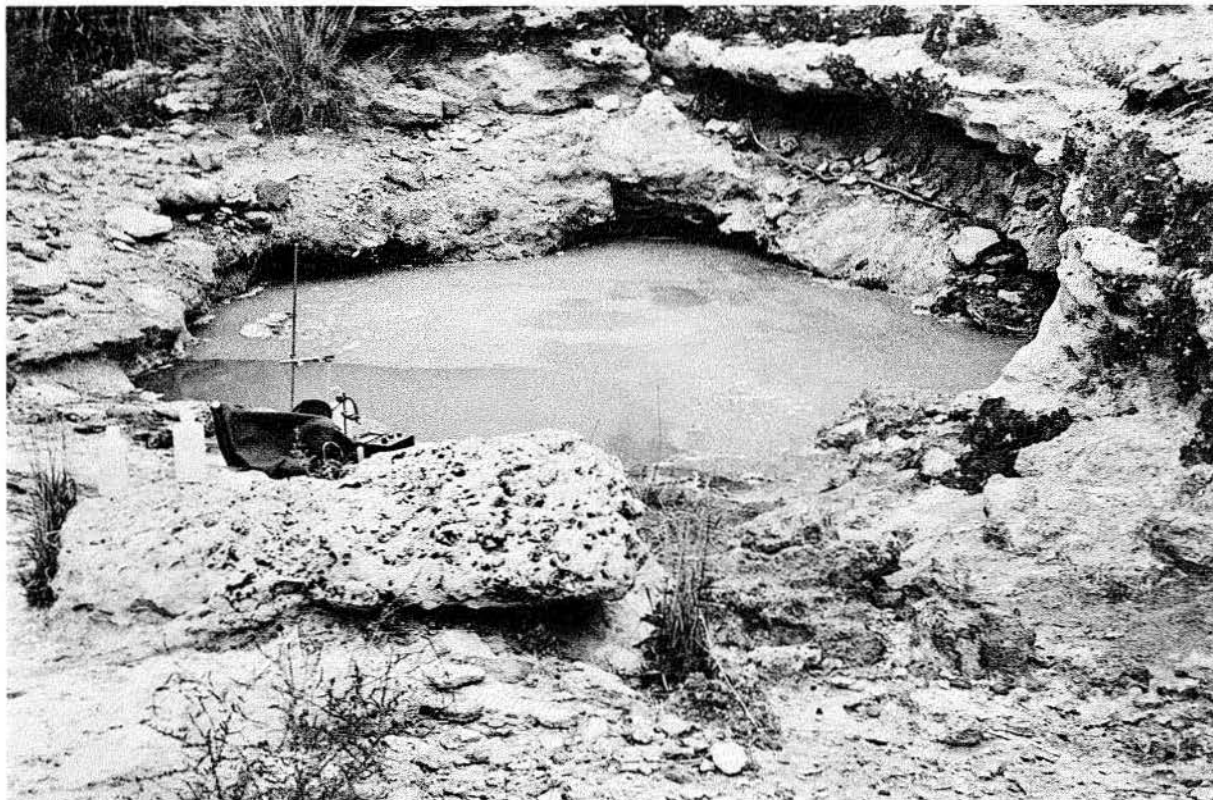


Figure 13. Spring at Rancho Cipres surrounded by travertine deposited by spring waters. Spring surface is coated with scum except where bubbling of CO<sub>2</sub> and H<sub>2</sub>S has disrupted the surface.



Table 1. Gulf wells: depth, temperature, and hydrologic data.<sup>a</sup>

	GULF WELLS	
	Presidio Trust #1	R.P. Swafford #1
Total depth	1,893 m (6,208 ft)	2,688 m (8,815 ft)
Water-producing depth	958 m (3,142 ft) 29 m above MSL	874 m (2,868 ft) 193 m above MSL
Temperature, this study	72°C (162°F)	69°C (156°F)
Temperature, reported	77°C (171°F) 82°C (180°F)	79°C (174°F) 82°C (180°F)
Flow	8,300 l/min (2,200 gpm)	5,700 l/min (1,500 gpm)

<sup>a</sup>Data from Gulf Oil Corporation and White and others (1977).

Table 2. Correlation of Cretaceous formations discussed in text.

Big Bend National Park (Maxwell and others, 1967)	Peguis, Chihuahua Benavides, Chihuahua (Gries, 1970) (Hernandez-Rios, 1974)	Sierra Vieja (Buongiorno, 1955) (Wolleben, 1966)
Pen Formation	San Carlos Formation	San Carlos Formation
Boquillas Formation	Ojinaga Formation	Ojinaga Formation
Buda Limestone	Buda Limestone	Buda Limestone
Del Rio Clay	Del Rio Clay	Del Rio Clay
Santa Elena Limestone	Loma Plata Limestone	Georgetown Limestone

through massive, cavernous (therefore permeable) Buda and Santa Elena Limestone but not through the thin-bedded, nonfractured part of the Boquillas Limestone. Hot water thus discharges either at the contact or through fractures only slightly above the contact, a setting similar to the springs at Peguis. The Del Rio Clay, which should be impermeable, is very thin at Peguis and could be permeable if fractured. No evidence for recent fault movement was observed in the spring area by either this author or Belcher and Goetz (1977), who more thoroughly examined the whole park area.

Several warm springs occur along the Rio Grande downstream from the park. The springs are accessible only by a multiple-day boat trip and were not visited. K. E. Smith (personal communication, 1978) measured a temperature of 32°C (90°F) for three of the springs. Geologic maps (International Boundary Commission, 1951) show a number of springs along the river but do not identify warm springs; most springs lie on north-northwest-trending normal faults. These warm springs are probably similar to those in Big Bend Park.

#### Terlingua Wells

Local inhabitants reported that abandoned mercury mines in the Terlingua mining district are flooded with

45°C (113°F) water at a depth of approximately 275 m (900 ft). In addition, at least one deep well (265 m, 875 ft) taps hot water (43°C, 110°F) at a similar depth. These reports could not be verified because the well currently has no pump, and the mines, which are collapsing, are unsafe to enter. During January, when the outside air temperature was only 10°C (50°F), hot air (32°C, 90°F) was flowing out the main shaft of the Rainbow Mine, which is reportedly 180 m (600 ft) deep. Presumably an air-convection system is forming which is analogous to a hydrothermal system.

#### Springs near San Carlos

A number of warm springs (maximum temperature 32°C, 89°F) discharge into the canyon of the Rio San Carlos approximately 2 km west of the town of Manuel Benavides (formerly San Carlos), approximately 70 km southeast of Presidio (fig. 2). Travertine deposited by the springs lines canyon walls and slopes below the springs (fig. 16). Hernandez-Rios (1974) mapped a large area which includes the springs, and Chacon (1972) mapped in detail an area adjacent to but not including the springs.

The geologic setting of these springs is unusual for hot springs in the Rio Grande area. The springs lie at the edge of what is apparently a large caldera, 60 km in diameter. All

the springs are in the upper plate of a high-angle reverse fault, which has thrust Lower Cretaceous Loma Plata Limestone (table 2) over Upper Cretaceous rocks (figs. 17 and 18). The Loma Plata Limestone is massive, cavernous, and highly fractured. Although fractured near faults, shales in the Ojinaga Formation are otherwise impermeable. The springs thus discharge at the contact between permeable and impermeable units. Cretaceous strata in the study area are complexly folded, similar in many respects to Laramide folds to the northwest. Laramide folding may be represented, but much of the deformation evidently occurred contemporaneously with caldera development, possibly during resurgence. Nearby volcanic units are also folded, and all the folds generally coincide with the caldera outline (fig. 18). Resurgence and doming probably created the large anticline shown on the west side of the map and thrust the Lower Cretaceous rocks over the Upper Cretaceous sequence at the edge of the caldera. Volcanic units could have been folded during the same structural events.

In a brief reconnaissance of the spring area, no evidence of recent igneous activity was found. If the caldera-related rocks are as old as other igneous units in the region, they are too old to have retained residual heat. Thermal waters are related to the caldera only in that fracturing during caldera development created permeable channelways for water circulation.

### Sotano de Sauz

An unusual hot cave, Sotano de Sauz, is situated in limestone in Mexico approximately 10 km north of Benavides and about 5 km south of the Rio Grande (Sprouse, 1977). The cave was not visited during this study, but from Sprouse's description and the writer's knowledge of the geology of the area, the cave must be in the Loma Plata Limestone and possibly in Buda Limestone (table 2). According to Sprouse, much of the cave is joint controlled. The bottom of the cave is a large mudflat that evidently acts as a sump through which water drains from the cave.

Temperature in the cave reaches 41°C (106°F) at 220 m, which is the maximum depth of the cave. The thermal gradient of about 100°C/km could result from the presence of hot water at the water table below the cave. The cave is apparently a point of recharge, however, so the high thermal gradient is anomalous.

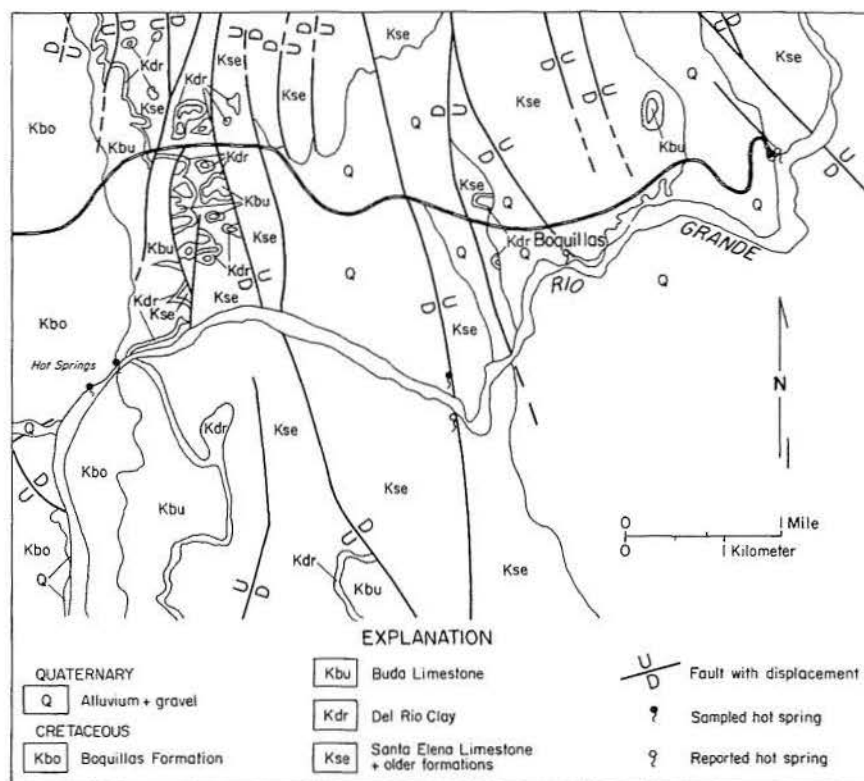
### HOT WELLS

Hot Wells was drilled in 1907 by the Southern Pacific Railroad in bolson fill in Eagle Flat (fig. 2). Total depth of the well is 305 m (1,000 ft); water level is 206 m (675 ft) below the surface. The maximum surface temperature of the water was 40°C (104°F) after approximately 1 hour of pumping with a submersible pump. The water probably had



Figure 14. Hot springs at Peguis on east limb of anticline at Buda Limestone (massive limestone at left)—Ojinaga Formation contact (thin-bedded limestone and shale overlying Buda).

Figure 15. Geologic map of hot-spring area of Big Bend National Park and adjacent Mexico. Geology in Texas from Maxwell and others (1967) and in Mexico from Smith (1970).



not cooled significantly from the temperature at depth. The thermal gradient in the well is at least  $75^{\circ}\text{C}/\text{km}$ . Because such a high gradient without convection is unlikely, the water probably rose by hydrothermal convection from greater depths.

Hot Wells is near the extension of the Rim Rock Fault indicated by the gravity data of Wiley (1972), although there is no surface expression of the fault. Perhaps hot water is circulating along the Rim Rock or a related fault and discharging into permeable bolson fill at the top of the water table. A U. S. Geological Survey water test well, J. C. Davis No. 1 (Gates and White, 1976), drilled to a total depth of 613 m (2,013 ft) 8 km southeast of Hot Wells near the trace of the Rim Rock Fault, encountered a more normal thermal gradient of  $32^{\circ}\text{C}/\text{km}$  and a maximum temperature of  $38^{\circ}\text{C}$  ( $100^{\circ}\text{F}$ ).

### Shallow Ground-Water Flow within Basins

Water-level measurements in wells in dissected basins and adjacent highlands along the Rio Grande in Trans-Pecos Texas show that the water table slopes toward basin centers and the Rio Grande and generally follows the present topography (White and others, 1977). Recharge occurs partly in the highlands from rainfall and partly in the basins from rainfall and runoff from the highlands.

Ground-water movement to the Rio Grande through the bolson fill is probably negligible because permeable beds needed to transmit the water do not exist near the basin centers (Groat, 1972). Coarse-grained facies of the late Tertiary basin fill are generally permeable to horizontal water flow, but interlayered fine-grained facies probably restrict vertical permeability. Fine-grained basin-center

facies are generally impermeable. Cold springs appear in bolson fill along faults which apparently act as barriers to ground-water flow. Ground water is dammed behind the faults and discharges in numerous springs above the faults. (No hot springs occur in such a setting.) This water can then move to the Rio Grande as underflow in tributary channel deposits. Otherwise, the Rio Grande is hydrologically isolated by its location within the fine-grained basin-centered sediments. Only where the river crosses coarse-grained basin-margin sediments, bedrock, or major faults (all of which commonly occur together) is it hydrologically connected to the ground-water system.

### Geologic Control of Hot Springs and Geometry of Flow

The presence of hot springs indicates that there are permeable channelways which allow circulation of ground water to a source of heat at depth where the water is heated and then returned to the surface. Most hot springs in this study area are on or near the normal fault system created during late Tertiary extension. A second, smaller group of hot springs appears at the contact of permeable and impermeable rocks. One group of springs is adjacent to a reverse fault.

The springs on normal faults within bedrock, Capote and Nixon Springs, emanate directly from fractures within fault zones. The fractures provide permeability in otherwise impermeable rocks and are the primary channelways for the rise of hot water from source reservoirs. Indian Hot Springs, Red Bull Springs, and Ojos Calientes discharge within or just basinward of fault zones separating bedrock from basin fill. Faults and fracture systems are the major conduits for





Figure 16. Travertine deposited by warm spring waters coats Loma Plata Limestone on far wall of canyon of Rio San Carlos.



Figure 17. View to west of high-angle reverse fault with overturned Loma Plata Limestone thrust over Upper Cretaceous shales which crop out in Rio San Carlos at mouth of canyon. Warm springs discharge into canyon upstream from fault.

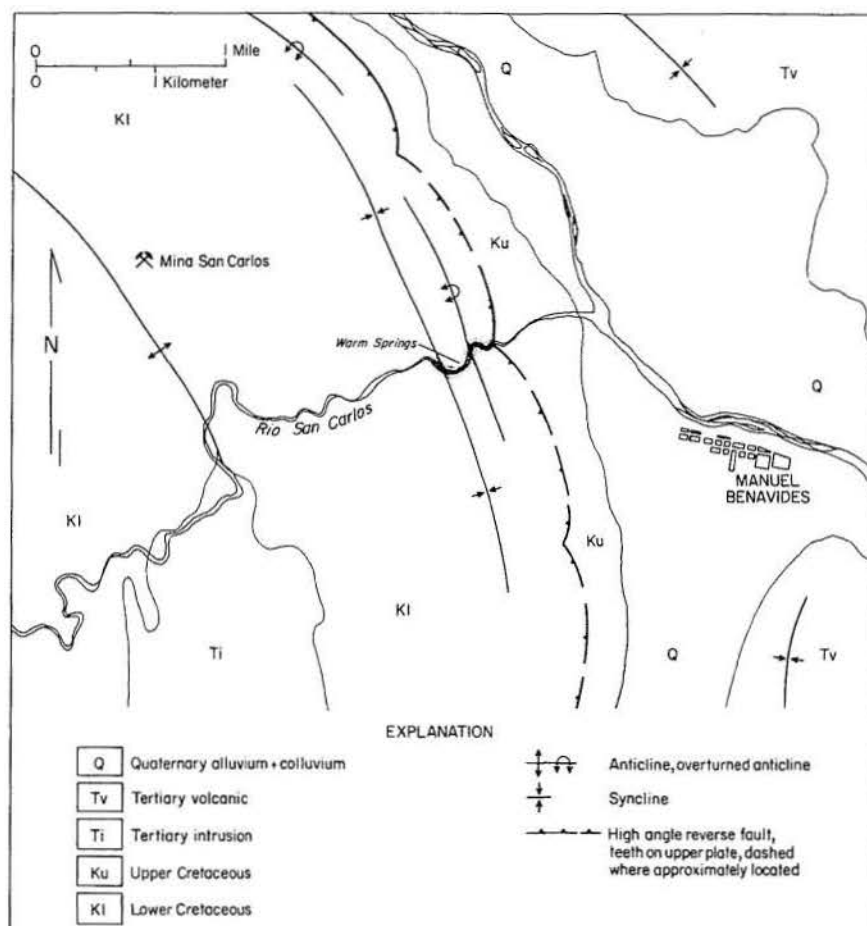


Figure 18. Geologic map of San Carlos warm-spring area, Chihuahua, Mexico. Geology from Chacon (1972), Hernandez-Rios (1974), and this report.

hot water, but at Indian Hot Springs and Ojos Calientes the water is probably diverted into permeable bolson or terrace gravels near the surface at the water table. The water moves laterally in the coarse sediment at the top of the water table, discharging where the water table intersects the surface. Hot springs within bedrock cannot discharge into shallow, permeable horizons because sufficient permeability for ground-water movement exists only within the fracture system.

Only two hot springs, Las Cienagas and Hot Springs - Ruidosa, appear in basins more than a few hundred meters from basin-margin faults. For each spring, shallow bedrock may be important in allowing hot water to rise near the surface without intersecting permeable basin fill and in creating a shallow water table. Las Cienagas discharges along the edge of a bedrock outlier. Bedrock may also occur at shallow depths at Hot Springs - Ruidosa, where Tertiary volcanic rocks crop out and dip gradually toward the spring.

No hot springs appear near basin centers, where faults are uncommon, and the unfaulted fine-grained basin fill has low permeability, providing no channelways for the rise of hot water. Even where faults do appear, there are no hot springs. Bedrock is at considerable depth, and the clay fault gouge in the basin centers may be no more permeable than undisturbed clay.

Some hot water rising along fault zones does not reach the surface. Thermal water in the Briscoe Well, for example,

was discovered only by chance. There are probably many as yet undiscovered thermal systems within the late Tertiary basins. Thermal water flow to the surface within the basins may be an unusual occurrence. Springs of any kind are rare within the undissected basins (Lobo Valley and Salt Basin) and are rare even within the dissected basins (Presidio and Hueco Bolsons).

Recent fault displacement has occurred in several areas of hot-spring activity in West Texas. Continued brecciation by recent fault movement may be important in keeping fracture zones open and permeable. Otherwise, precipitation of calcite from thermal water might seal fractures. Recent fault movement, however, is apparently not a prerequisite for deep circulation of ground water because several hot springs do appear in areas with no evidence of recent fault activity.

The springs at San Carlos are also fault controlled; they differ only in style of faulting. Greater brecciation and opportunity for solution permeability to develop in thrust-faulted, massive limestone may account for the location of the springs.

Springs at Peguis and Hot Springs in Big Bend are unrelated to faults. At both locations, water discharges at the contact of Buda Limestone and the overlying Boquillas or Ojinaga Formations. The overlying formations are time-equivalent shales and flaggy limestones (table 2) and are relatively impermeable. The massive Buda Limestone is cavernous and, therefore, highly permeable. Hot water

rising from below reaches the contact and is diverted along the contact to the water table (fig. 19). Where the water table intersects the surface, the hot water discharges as springs analogous to cold springs, which commonly occur at the contact of permeable and impermeable rock types. Recharge water for a cold spring flows from a higher elevation through confined permeable rock and emerges where the contact intersects land surface. The driving force for this cycle is gravity. The additional driving force created by the difference in density of thermal water and cooler water allows the thermal water to discharge above its recharge area.

Recharge for all the hot springs probably occurs in the adjacent highlands and is spread over a broad area, whereas discharge is concentrated in one area. Faults in the otherwise impermeable igneous rocks and faults and solution cavities in Cretaceous limestone allow meteoric water to circulate to depth. Recharge waters circulate downward through numerous separate channelways and coalesce at depth to follow a single, more restricted return conduit. Recharge through the same basin-margin faults, which are discharge conduits, is possible but less likely near hot springs. Downward-moving cold water would mix with upward-welling hot water and the hot spring would not occur. Hot springs, therefore, probably appear where recharge to a fault system is unfavorable. Recharge is still less likely in the basin interior for the same reasons that hot springs do not appear there: unfaulted, fine-grained basin fill exhibits extremely low permeability.

Recharge to Indian Hot Springs and springs in Big Bend National Park, which lie along the Rio Grande, and to the springs at Peguis, which lie along the Rio Conchos, could come from the rivers. At each location the rivers cross major structures and bedrock. The Rio Grande crosses the

Caballo Fault and the southern end of the Quitman Mountains at Indian Hot Springs and crosses numerous faults and limestone bedrock in the hot-spring area of Big Bend National Park. At Peguis, the Rio Conchos crosses cavernous limestone exposed in an anticline.

The two Gulf Wells, Briscoe Well, Hot Wells, and the Terlingua mines and Terlingua Well tap hot water that does not reach the surface because, for all but the Gulf Wells, the water table is below the land surface. Whether or not hot water is rising from greater depth by convection, the hot water remains at the top of the water table far below the surface. Hot water in the Gulf Wells is far below the water table. The temperature reversal in the wells indicates that the water is rising from still greater depths—at least 2,500 m according to the well temperature log. Water must be rising to the bottom of an impermeable layer, possibly shales of the San Carlos or Ojinaga Formations that overlie the reservoir Georgetown Limestone (table 2), displacing cold water. If permeable channelways exist, the hot water could rise still nearer to the surface, but evidently it does not discharge to the surface.

The true flow paths are undoubtedly more complicated than those discussed in this report. Intersecting fractures and other permeable zones probably allow considerable mixing of cold and hot convecting water. A hot spring appears where a somewhat fortuitous combination of geologic and hydrologic features allows the hot water to reach the surface. The geologic setting of hot springs and wells, along with facets of thermal-water chemistry discussed in the next section, provide some information about the geometry of thermal-water flow. However, the deeper parts of the thermal convection systems in West Texas remain poorly understood.

## **GEOCHEMISTRY OF THERMAL WATERS**

### **Sampling and Analysis**

In this study, water samples were taken from locations as close as possible to the discharge point. Consequently, samples were collected where water discharges from fractures in bedrock, from pipes or cisterns in artificially enhanced springs, from well discharge, or from collection pools where discharge consisted of many small seeps. Any of these samples, especially the slowest discharging waters, could have undergone considerable chemical change before testing, particularly changes in pH and alkalinity.

Temperature, pH, and alkalinity titrations were measured in the field, and other analyses were performed in the laboratory. General analytical methods are given in table 3.

Several different kinds of samples were collected for different analyses. All samples were pressure filtered through 0.45-micron filter paper and stored in polyethylene bottles. For analysis of most major constituents 1 liter remained unacidified, and 1 liter, acidified to a pH of 2 with distilled 70-percent  $\text{HNO}_3$ , was used for trace element analysis. The bottle for standard analyses was washed and rinsed in distilled, deionized water. The bottle for trace elements was rinsed overnight with 10-percent  $\text{HNO}_3$ , then with 10-percent

HCl, then rinsed with distilled, deionized water. A third sample, for  $\text{SiO}_2$  determination, was collected during the initial phase of sampling and again from selected sites in later sampling. A 125-ml bottle was washed, rinsed, dried, and then partly filled with 90 ml of distilled water. During sampling, 10 ml of spring water was pipetted into the bottle to dilute the sample and to prevent precipitation or polymerization of dissolved silica. Additional silica analyses were performed on the undiluted sample both with and without NaF treatment for depolymerization. Initial results with all three methods were identical, probably because of the relatively low silica content in these waters. During later testing, samples were diluted from only a few springs and wells that were suspected of having high silica concentration.

### **Results**

#### **INTRODUCTION**

The source of dissolved constituents in hot spring waters has been a long-standing question. Ellis and Mahon (1964, 1967), Ellis (1970), and Mahon (1970) have demonstrated that the chemistry of many thermal systems is dominated



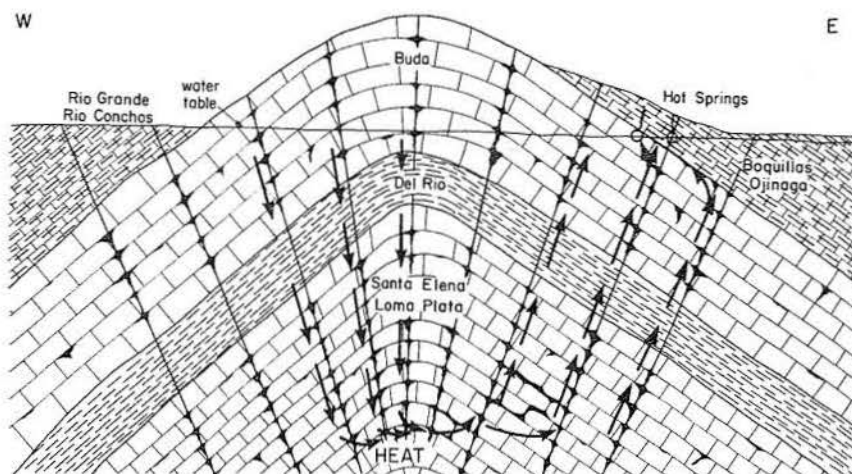


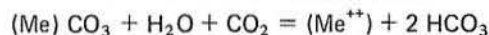
Figure 19. Postulated flow paths of hot-spring waters at Peguis and Big Bend National Park.

by water-rock reaction. Nevertheless, White (1970) concluded that additional sources, such as fluids derived from igneous magmas, supplied some soluble elements (for example, chloride) that are minor components of most rocks.

Chemical analyses of spring and well waters from Trans-Pecos Texas and adjacent Mexico are given in table 3 and are presented graphically in figure 20. Although there is wide variation in the chemistry of thermal waters, correlation with host and subsurface rock types shows that there are different groups of waters. Generally, the thermal waters fall into the following three groups with at least some gradation between each: (1) waters dominated by calcium, magnesium, bicarbonate, and sulfate with moderate total dissolved solids; (2) waters with low to moderate total dissolved solids composed of sodium and bicarbonate; and (3) waters containing high total dissolved solids dominated by sodium, chloride, and sulfate.

#### CALCIUM-MAGNESIUM-BICARBONATE-SULFATE WATERS

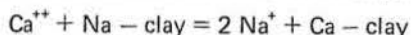
One group of spring waters has moderate total dissolved solids (600 to 900 mg/l), the highest proportion of calcium and magnesium (approximately 54 percent of cations), and the highest magnesium and the lowest calcium-magnesium ratios (table 3). Springs falling distinctly into this category are in Big Bend National Park, at San Carlos, and at Peguis. All these springs emanate from Cretaceous carbonate rocks or from adjacent alluvium. Available published analyses of hot spring waters show only one spring that definitely emanates from carbonate rocks. The water from Mammoth Hot Springs in Yellowstone Park is hotter and slightly higher in total dissolved solids but is similar in ion proportions to springs in this study (fig. 20b) (Thompson and others, 1975). The geologic setting and a reasonable estimation of water/host-rock reactions demonstrate that spring chemistry is determined mostly by solution of calcite and dolomite. Calcium, magnesium, and bicarbonate enter into solution by the following reaction:



The proportion of calcium and magnesium represented by the metal ion (Me) is determined by the particular carbonate mineralogy of surrounding rocks.

Cold spring waters emanating from carbonate rocks are composed almost entirely of calcium, magnesium, and bicarbonate with generally minor amounts of sodium, sulfate, or chloride (table 5 of White and others, 1963). The compositions cluster near the calcium-magnesium-bicarbonate corner on a trilinear plot and off the field to the left of the plot in figure 20b. Compared to cold springs, the hot springs of the Rio Grande area have considerably greater sodium, sulfate, and chloride contents and would therefore require an additional source for these solids besides solution of carbonate materials.

Sodium, chloride, sulfate, and additional calcium could be contributed by dissolution of gypsum and halite in minor evaporites in Cretaceous rocks. Because molecular calcium to sulfate plus bicarbonate ratios are considerably less than 1 (table 3), either some calcium must be lost or additional sources of sulfate and bicarbonate must exist. Sodium-chloride ratios are greater than 1, so either sodium is being added or chloride is being lost. Most likely, calcium is being exchanged for sodium contained in clays in Cretaceous marine shales. The reaction is



Solution of sodium-bearing minerals other than halite is doubtful. Feldspars are not abundant in the Cretaceous rocks, and solution of feldspars could not contribute sulfate or chloride. Other sources of sulfate do exist, however, such as oxidation of reduced sulfur ( $FeS_2$  or  $H_2S$ ). Oxidation reactions are probably occurring, but water in equilibrium with air contains only sufficient oxygen to produce by this process about 15 mg/l sulfate. Gypsum solution remains the only reasonable source of large amounts of sulfate, and gypsum is present in the host rock. Other sources for sulfate and chloride, such as magmatic gases, cannot be excluded, but their presence is unnecessary.

Recharge for the springs in Big Bend could come from the Rio Grande and from the Rio Conchos for the springs at Peguis. Because almost all the flow of the Rio Grande below Presidio is from the Rio Conchos, the water chemistry of the two rivers is similar. Published analyses of

Table 3. Chemical analyses of thermal and nonthermal waters (Rio Grande area of Trans-Pecos Texas and adjacent Mexico).

Analyzed Constituents <sup>a</sup> General analytical methods	1	2	3	4	5	6	7	8	9	10	11	11A	12	13	14
	INDIAN HOT SPRINGS					Red Bull Spring	BIG BEND NATL. PARK			San Carlos Springs	PEGUIS		Las Cienagas	Hot Springs—Ruidosa	Briscoe Well
	Stump	Chief	Squaw	Beauty	Soda		Spring at Rio Grande Village	Hot Springs	Big Bend #2		Peguis	Peguis #2			
Sodium (Na) Flame photometry	2185	2340	2375	2610	725	312	98	108	107	108	126	123	228	148	186
Potassium (K) Flame photometry	134	158	160	200	40.9	11	5.4	5.8	5.8	4.4	4.2	4	6	14.5	17.7
Calcium (Ca) Atomic absorption	150	145	160	175	62	15.5	125	133	133	134	62.5	67	27	27.5	65.8
Magnesium (Mg) Atomic absorption	27	34	29	36	14.4	1.7	36.4	36.4	36.4	9.6	16.8	17.5	2.4	4.6	8.4
Lithium (Li) Flame photometry	2.2	2.5	2.5	2.8	0.6	0.2	0.2	0.2	0.2	0.2	0.1	0.1	0.2	0.2	0.2
Strontium (Sr) Atomic absorption	3.8	3.4	—	3.5	1.9	0.7	4.2	—	4.4	2.1	1.8	1.8	0.5	0.6	1.3
Bicarbonate (HCO <sub>3</sub> ) Electrometric titration at spring	906	902	962	1034	623	474	280	273	266	207	243	244	415	290	281
Sulfate (SO <sub>4</sub> ) Thorin titration	1090	1150	1190	1270	440	259	346	371	366	392	128	131	119	100	192
Chloride (Cl) Mercurimetric titration	2680	2950	3000	3380	605	87.2	64.3	70.2	69.7	29.4	127	126	87.8	72.2	140
Nitrate (NO <sub>3</sub> ) Cd reduction to nitrite, nitrite analysis by diazotization method	0.1	5.4	1	0.9	0.4	<0.1	1	0.5	0.7	3.1	4.4	5.5	0.9	<0.1	10.6
Fluoride (F) Ion selective electrode	2.8	2.7	2.8	3.1	3.6	3.2	2.2	2.3	2.3	3.2	1.5	2.2	7.2	3.8	5.1
Boron (B) Carmine method	4.7	5	5.2	5.7	2	—	0.2	0.3	0.3	0.4	0.4	0.2	0.4	0.6	0.8
Silica (SiO <sub>2</sub> ) Molybdate blue colorimetric	40	40	35	35	20	36	21	22	22	37	22	21	39	35	39
pCO <sub>2</sub> Calculated from analysis	0.237	0.107	0.111	0.170	0.017	0.0058	0.015	0.018	0.024	0.0087	0.0094	0.0084	0.0084	0.012	0.010
Saturation index (SI) Calculated from analysis	0.05	0.31	0.20	0.20	0.34	0.28	0.30	0.29	0.15	0.22	0.17	0.25	0.20	0.07	0.33
pH pH meter at spring	6.6	7	6.9	6.8	7.5	8	7.3	7.2	7.1	7.4	7.5	7.5	7.7	7.5	7.5
Temperature °C Maximum reading thermometer at spring	47.2	44.4	33.6	39.7	27.2	37	35.6	40.6	40	31.7	35.6	36.1	30.2	45	41.1
Total dissolved solids (TDS) Calculated from analysis	6766	7280	7434	8231	2223	960	842	884	879	723	614	620	723	549	805
Atomic Ratios															
Cl/Ca	20.2	23	21.2	21.8	11	6.37	0.58	0.6	0.59	0.25	2.30	2.12	3.68	2.97	2.40
Cl/Mg	68	59.5	71.4	64.5	28.8	35.2	1.21	1.32	1.31	2.10	5.18	4.94	25.1	10.8	11.4
Cl/B	190	167	170	192	85.5	—	90	66	66.7	20	90	90	50	33.3	55.7
Na/Cl	1.26	1.22	1.22	1.19	1.85	5.52	2.35	2.37	2.37	5.66	1.53	1.50	4	3.16	2.05
Na/K	27.7	25.2	25.3	22.2	30.2	48.2	30.9	31.7	31.4	41.8	51	52.3	64.6	17.4	17.9
Na/Li	320	260	260	290	350	450	140	160	160	160	550	530	330	210	270
Ca/SO <sub>4</sub>	0.33	0.30	0.32	0.33	0.34	0.14	0.87	0.86	0.87	0.82	1.17	1.23	0.54	0.66	0.82
Ca/HCO <sub>3</sub>	0.25	0.25	0.25	0.26	0.15	0.05	0.68	0.74	0.76	0.99	0.39	0.42	0.10	0.14	0.36
Ca/Mg	3.37	2.59	3.35	2.95	2.61	5.52	2.08	2.22	2.22	8.47	2.26	2.32	6.83	3.63	4.75

<sup>a</sup>Concentrations in milligrams per liter.

Table 3.—Concluded.

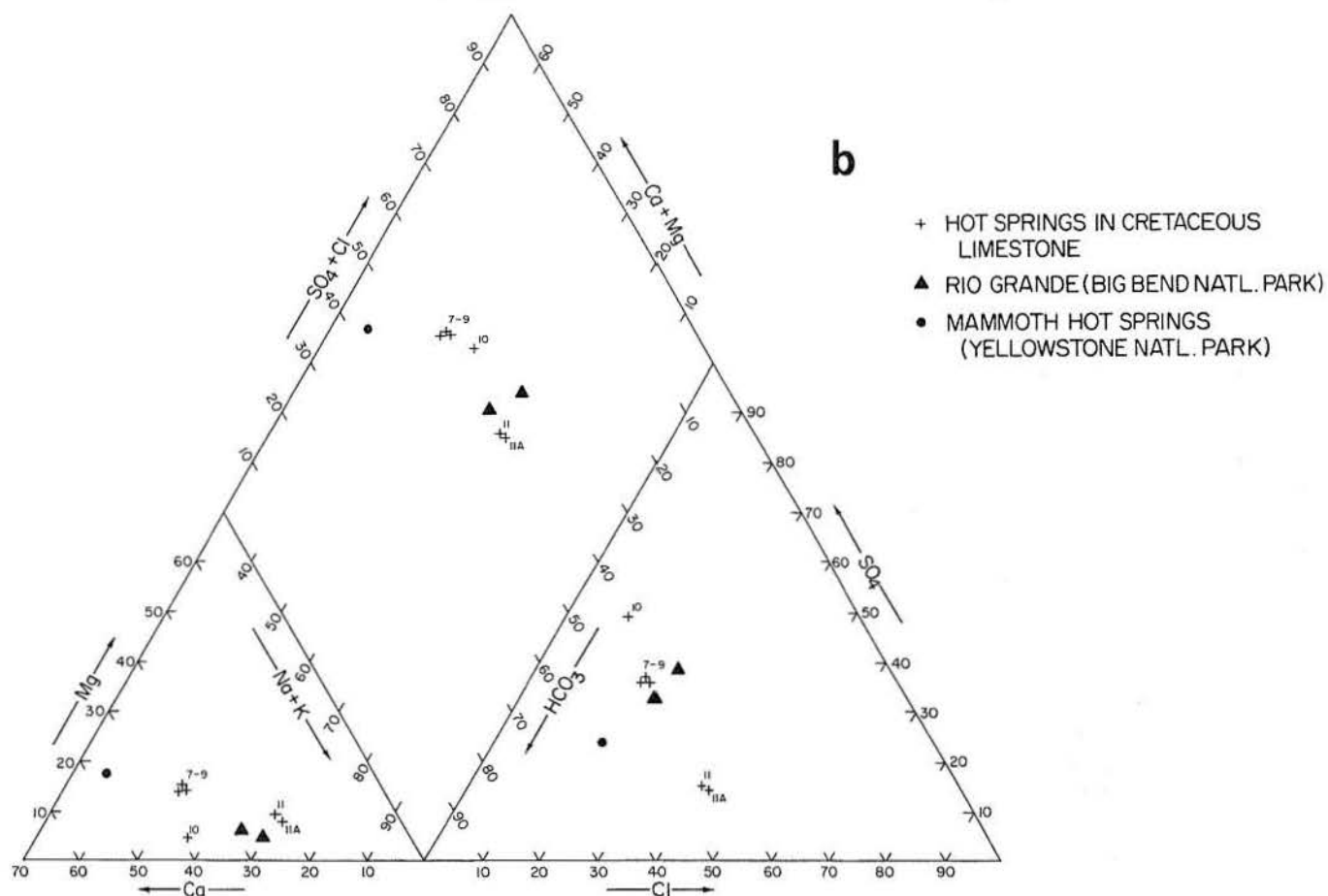
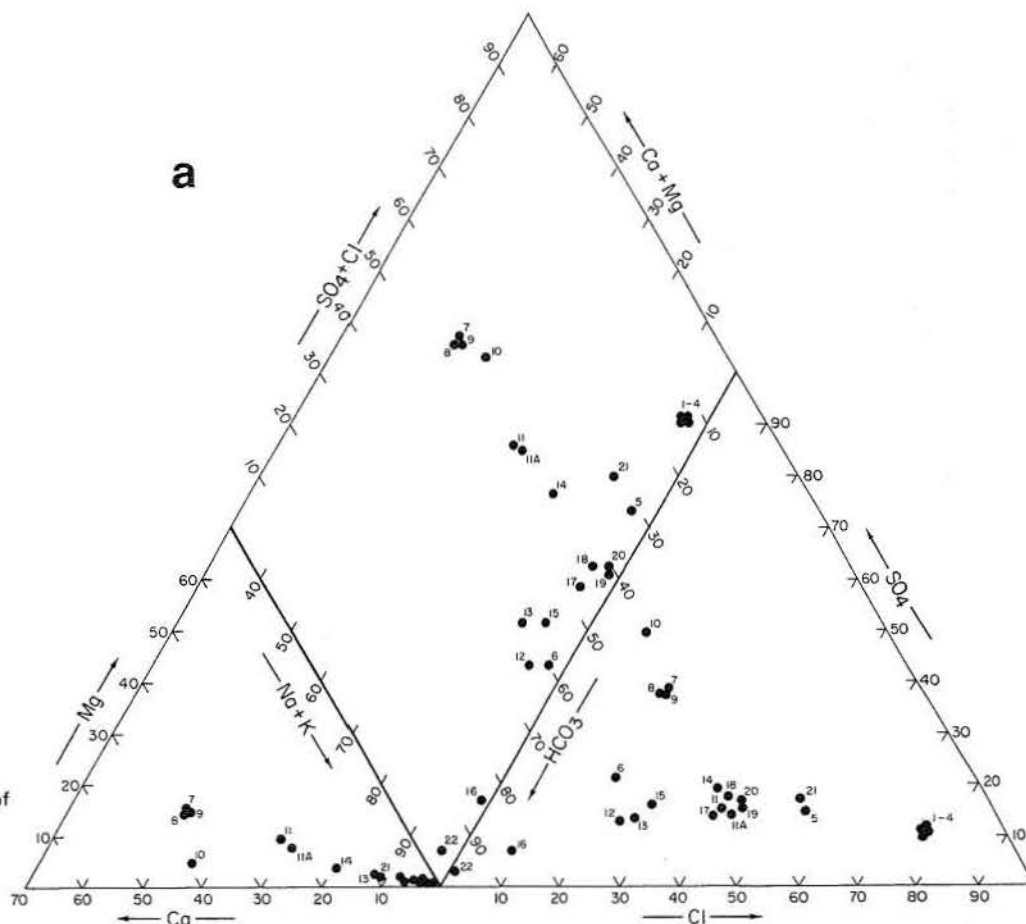
Analyzed Constituents <sup>a</sup> General analytical methods	15	16	17	18	19	20	21	22	23	24	25	26	27	28
	Nixon Springs	Capote Warm Spring	Gulf— Presidio	Gulf— Swafford	OJOS CALIENTES		Rancho Cipres	Hot Wells	Mimbres Well	Vizcaino Well	Naegele Springs	Mexican Springs	Nuñez Well	Alamo Springs
					#3	#4								
Sodium (Na) Flame photometry	160	120	368	500	750	780	1140	105	95	202	195	212	91	190
Potassium (K) Flame photometry	5.5	0.6	42.9	60	68	68	58	1.6	3.6	10	21.5	1.6	3.3	0.8
Calcium (Ca) Atomic absorption	20.5	1.6	37	46.4	25	32	221	4	30	57.5	55.1	12	63.3	3.8
Magnesium (Mg) Atomic absorption	2.3	0.09	2	1.2	2.1	2.7	20	1.8	2.4	3.6	6.7	0.66	21	0.04
Lithium (Li) Flame photometry	0.1	<0.1	0.5	0.6	0.9	1	1.3	<0.1	<0.1	<0.1	0.2	<0.1	0.1	<0.1
Strontium (Sr) Atomic absorption	0.2	<0.1	0.7	0.7	1.4	1.5	4.6	0.3	0.1	0.1	1	—	—	—
Bicarbonate (HCO <sub>3</sub> ) Electrometric titration at spring	220	226	497	563	756	783	1048	293	247	477	390	462	274	237
Sulfate (SO <sub>4</sub> ) Thorin titration	97.5	31.8	230	355	458	481	898	16	58.4	119	161	54.4	99	115
Chloride (Cl) Mercurimetric titration	63	12.8	239	300	452	469	1010	<2.0	19.6	57.2	98.9	35.2	65.1	69.6
Nitrate (NO <sub>3</sub> ) Cd reduction to nitrite, nitrite analysis by diazotization method	2.7	5.3	0.2	<0.1	0.8	0.9	0.1	2.5	6.2	3.4	0.2	0.3	21.6	9.4
Fluoride (F) Ion selective electrode	3.4	2.5	10.6	10.6	11.8	12.4	5.6	2.4	3.8	4.1	5.9	3.2	2.4	3.8
Boron (B) Carmine method	0.3	0.4	1	1.2	1.8	1.9	2.6	0.4	0.3	0.6	0.4	0.2	0.5	0.5
Silica (SiO <sub>2</sub> ) Molybdate blue colorimetric	43	37	76	144 <sup>b</sup>	95	87	46	20	51	60	43	39	52	22
pCO <sub>2</sub> Calculated from analysis	0.0093	0.00022	0.064	0.084	0.014	0.045	0.20	0.00048	0.0042	0.011	0.0090	0.028	0.0053	0.00042
Saturation Index (SI) Calculated from analysis	-0.42	0.05	0.19	0.18	0.79	0.46	0.28	0.45	0.07	0.41	0.28	-0.64	0.26	0.02
pH pH meter at spring	7.4	9	7.2	7.1	8.1	7.5	6.7	8.8	7.8	7.6	7.6	7.2	7.7	8.7
Temperature °C Maximum reading thermometer at spring	31.7	36.7	72 <sup>c</sup>	69 <sup>c</sup>	90	69	35	40	24.4	23.9	24	24.4	22	25
Total dissolved solids (TDS) Calculated from analysis	507	329	1253	1697	2239	2323	3924	300	392	753	781	586	554	532
Atomic Ratios														
Cl/Ca	3.47	9.01	7.30	7.31	20.5	16.6	5.16	<0.28	0.74	11.3	2.03	3.32	1.16	20.7
Cl/Mg	18.8	80	82	172	147	119	34.6	<0.38	5.6	10.9	10.1	36.6	2.13	1190
Cl/B	60	10	74.4	85	640	665	715	<0.75	16.7	26.7	70	50	36	40
Na/Cl	3.92	14.5	2.37	2.57	2.56	2.56	1.74	>16.2	7.47	5.45	3.04	9.29	2.15	4.21
Na/K	49.5	340	14.6	14.2	18.8	19.5	33.4	11.2	44.9	34.4	15.4	220	46.9	404
Na/Li	690	<520	230	240	330	340	260	>460	>410	>880	280	>920	400	>830
Ca/SO <sub>4</sub>	0.50	0.12	0.39	0.31	0.13	0.16	0.59	0.60	1.23	1.16	0.82	0.53	1.53	0.08
Ca/HCO <sub>3</sub>	0.14	0.01	0.11	0.13	0.05	0.06	0.32	0.02	0.19	0.18	0.22	0.04	0.35	0.02
Ca/Mg	5.40	10.8	11.2	23.5	7.25	7.19	6.70	1.35	7.58	9.69	4.99	11	1.83	57.6

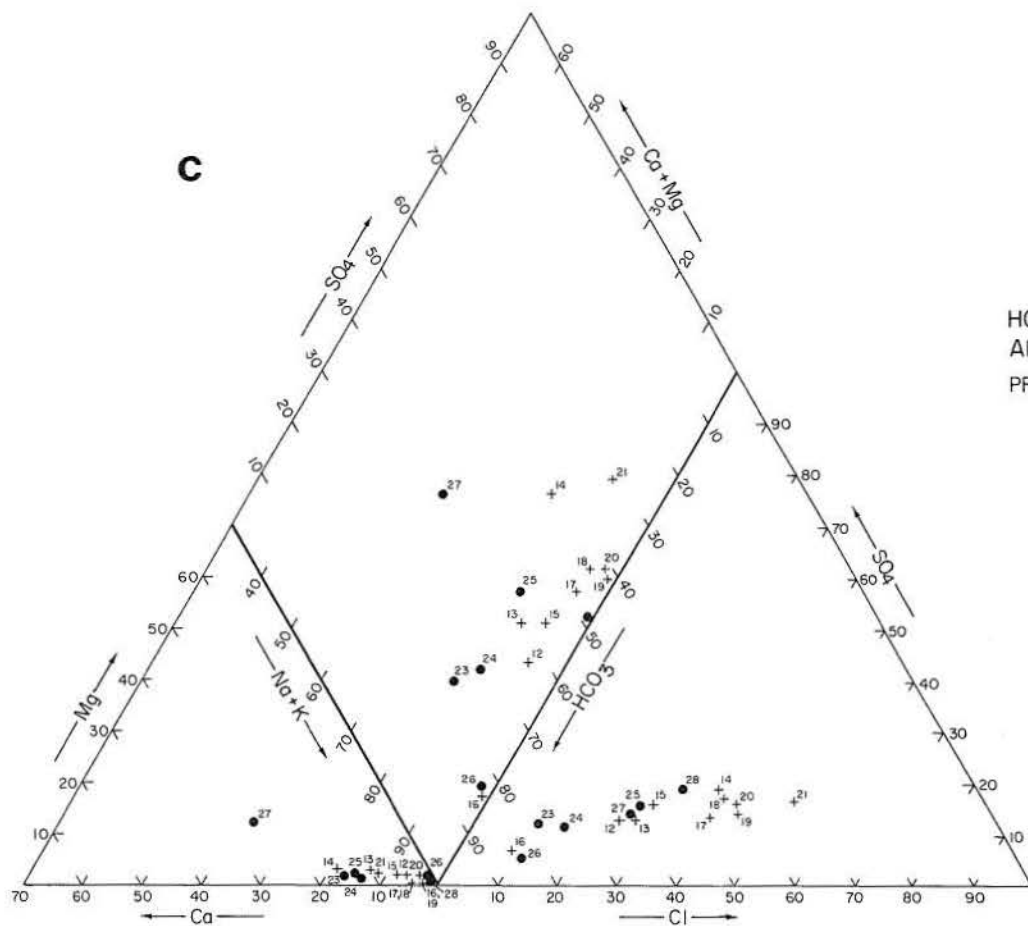
<sup>a</sup>Concentrations in milligrams per liter.<sup>b</sup>187 reported.<sup>c</sup>82°C reported.



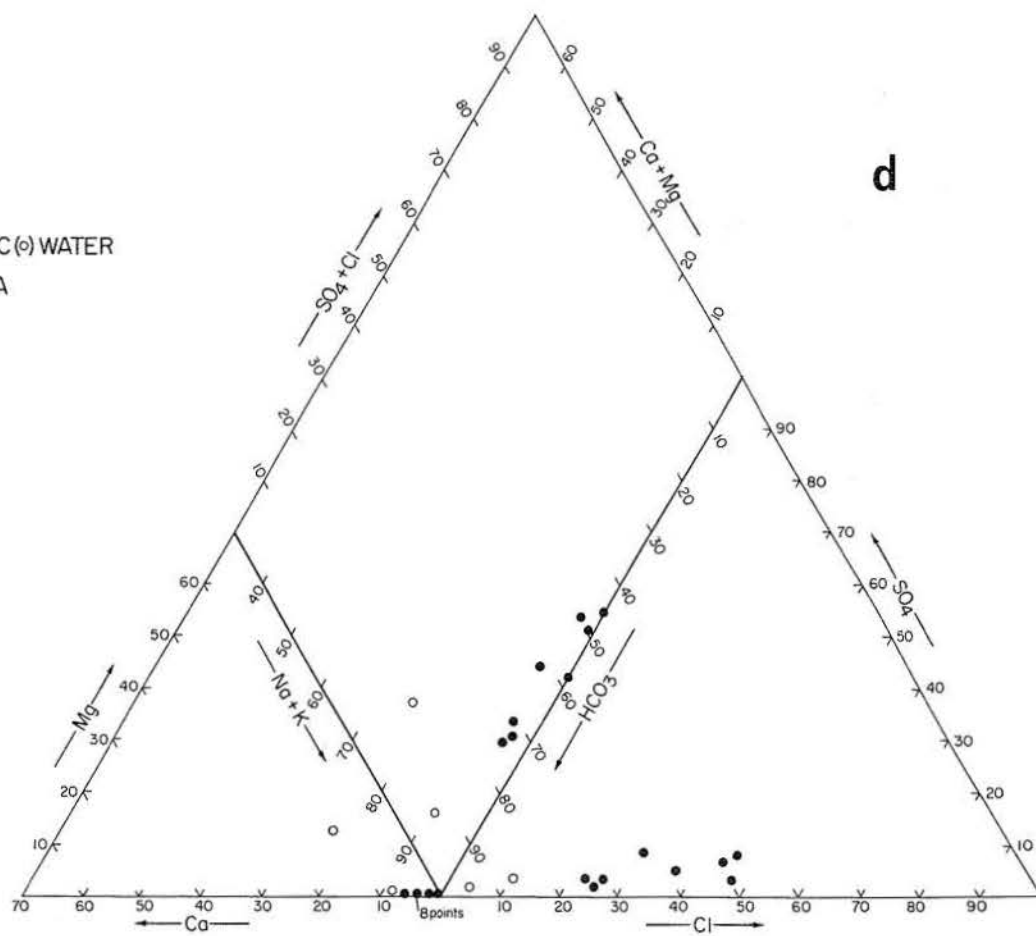
HOT SPRINGS AND WELLS  
TRANS-PECOS TEXAS AND  
ADJACENT MEXICO

Figure 20a-e. Trilinear diagram of  
thermal and nonthermal waters  
discussed in text. Proportions  
are in moles per liter.





THERMAL (•) AND METEORIC (o) WATER  
LONG VALLEY, CALIFORNIA







of the volcanic section is zeolitized with high sodium but low potassium concentrations. Spring waters are high in sodium and low in calcium and magnesium, but many are also low in potassium with high sodium-potassium ratios (table 3). Possibly, zeolitization removed most of the potassium that could go into solution. Sodium-potassium ratios in thermal waters can be controlled by temperature-dependent equilibrium with sodium-potassium feldspars. However, there is no correlation of sodium-potassium ratios with measured or estimated temperatures of the thermal waters. Cation exchange of calcium for sodium may also be significant in keeping calcium concentrations low and sodium concentrations high. Hall (1963) suggested that cation exchange within a rhyolitic breccia converts calcium bicarbonate water to sodium bicarbonate water which discharges as hot springs in the Socorro, New Mexico, area. In any event, the cation chemistry of the spring waters is consistent with the chemistry of the surrounding rock.

Hydrolysis also releases silica to solution. Cold ground waters listed in table 3 contain as much or more silica as many of the thermal waters. Because silica content of thermal water is used as a geothermometer, this reaction must be considered when interpreting silica geothermometers.

Nearly complete gradation exists between sodium bicarbonate waters and the calcium-magnesium-bicarbonate waters whose chemistry is derived from solution of limestone. The proportion of calcium and magnesium increases from approximately 1 mole percent in Capote Springs to 19 mole percent in Briscoe Well to 37 mole percent in Nuñez Well, a cold well. Sulfate and chloride, as a percentage of total anions, increase from 4 percent in Hot Wells and 16 percent in Capote Springs to 57 percent in Briscoe Well. The presumed recharge area for the intermediate-composition springs includes both Cretaceous or Permian limestones and Tertiary silicic volcanic rocks. Mixing of waters in contact with different rock types probably accounts for the mixed chemistry of these spring waters.

The chemical similarity between cold and hot waters demonstrated in this study is evidence that the cold waters recharge the thermal convection systems and determine the chemistry of the thermal waters. Deep circulation and heating of water have little effect on water chemistry.

### SODIUM-CHLORIDE-SULFATE WATERS

A third group of waters is characterized by high total dissolved solids and high concentrations of sodium, potassium, chloride, sulfate, bicarbonate, lithium, and boron (table 3; fig. 20e, nos. 1, 2, 3, 4, 5, 17, 18, 19, 20, 21). Calculated total dissolved solids for Indian Hot Springs are approximately 8,000 mg/l; for Rancho Cipres, approximately 4,000 mg/l; and for Ojos Calientes, approximately 2,200 mg/l (table 3). Even though 8,000 mg/l is relatively dilute compared to sea water or many brines associated with evaporites (White and others, 1963, table 27), the total dissolved solids in these waters are considerably higher than in other thermal waters in the area.

The chemistry and geologic setting of these springs demonstrate that the major source of dissolved ions is solution of evaporites. Ions cited above should be concen-

trated in evaporite deposits. The three spring groups listed fall either within the Jurassic evaporite basin described by DeFord and Haenggi (1971) or at its edge, where water could contact evaporites (fig. 21).

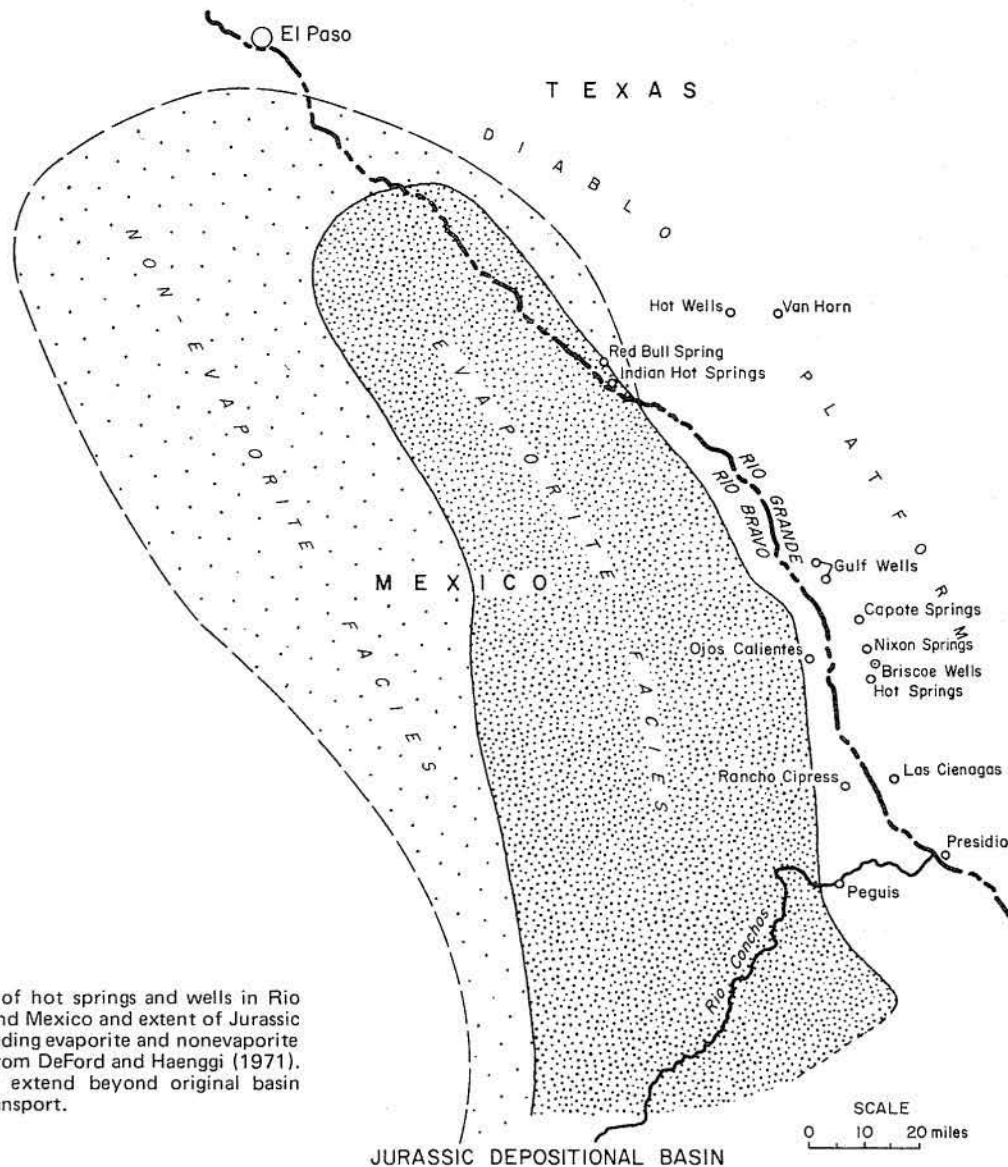
None of the hot spring waters is saturated with either halite or gypsum. Either the waters are not in contact with evaporites long enough to reach saturation or, more likely, waters circulating in other rock types are mixing with evaporite waters before discharge. If water in contact with evaporites were saturated before mixing, the amount of water coming from evaporites could be small compared with nonevaporite water, and yet the dissolved solids would still be dominated by the constituents of the evaporite source. The regular decrease in total dissolved solids, sodium, chloride, and sulfate from Indian Hot Springs waters to Rancho Cipres waters to Ojos Calientes waters probably reflects a decrease in the proportion of water from evaporites compared with water from other rocks.

Waters that are derived from evaporites and that do not undergo later chemical alteration ought to have molecular sodium-chloride and calcium-sulfate ratios near 1. None of the evaporite-related waters exhibits this ratio (table 3). Calcium-sulfate ratios for these and almost all other springs are less than 1, whereas all sodium-chloride ratios are more than 1. Only Indian Hot Springs waters have sodium-chloride ratios near 1; other evaporite spring waters have sodium-chloride ratios that range up to approximately 2.6. Calcium could also be supplied by solution of limestone which would further increase calcium-sulfate ratios. Depletion of calcium and enrichment in sodium was also noted for the carbonate waters. In this example, however, the evaporite component is dominant, and the shift in sodium-chloride ratios is considerably less than for the carbonate waters. Simple mixing of nonevaporite waters with low dissolved solids would not produce these waters. Chemical reactions with other host rocks, such as the cation exchange postulated for carbonate waters, more readily account for the difference. Calcium may also be removed by precipitation of calcite at depth because of its reduced solubility at elevated temperatures, or by travertine deposition near the surface outlets where the water releases carbon dioxide. For example, even considering their low total dissolved solids content in relation to other springs, Ojos Calientes waters have lower calcium and magnesium contents than either Indian Hot Springs or Rancho Cipres. All clearly evaporite-related springs deposit travertine at the surface and probably precipitate calcite in the shallow subsurface.

All the springs discharging evaporite waters are near outcrops of Cretaceous limestone. Limestone solution is undoubtedly one source of dissolved constituents. For example, many of the hot spring waters have high calcium-magnesium contents and low calcium-magnesium ratios (table 3) equal to those of the Big Bend area springs. However, chemical contributions from other sources are less important than the high sodium, chloride, and sulfate contents derived from evaporites.

Recharge of the Indian Hot Springs thermal system from the Rio Grande is possible considering the springs' physical setting. Indian Hot Springs may also incorporate the chemical characteristics of Rio Grande water. Analyses of Rio Grande water taken monthly for 1 year (U. S.

Figure 21. Location of hot springs and wells in Rio Grande area of Texas and Mexico and extent of Jurassic depositional basin including evaporite and nonevaporite facies. Jurassic basin from DeFord and Haenggi (1971). Evaporite rocks now extend beyond original basin because of tectonic transport.



Geological Survey, 1976) are similar in ionic proportions to those of Indian Hot Springs waters (fig. 20e). Rio Grande water is more like Indian Hot Springs water than most other surface or ground water in Trans-Pecos Texas. During periods of low discharge in late spring and early summer, total dissolved solids in the Rio Grande exceed 5,000 mg/l, approximately 65 percent of the amount found in Indian Hot Springs; however, with increased discharge during early fall, the amount of total dissolved solids decreases. The annual average amount of total dissolved solids is only about 2,600 mg/l or about one-third that of the spring waters. There are also subtle chemical differences. The river water, although high in sodium, is low in potassium (10 to 20 mg/l—well below spring concentrations of 130 to 200 mg/l). River water may recharge the thermal system and contribute part of the dissolved solids, but the major portion of dissolved solids must come from evaporite solution.

The Gulf Wells probably also discharge water that has been in contact with evaporites. The total dissolved solids (approximately 1,200 and 1,700 mg/l) in Gulf Wells water

are slightly greater than the dissolved solids in nonevaporite waters but considerably less than the dissolved solids in waters obviously derived from evaporites. Likewise, the sodium, chloride, and sulfate contents of Gulf Wells water are relatively high but are lower than those in the distinctively evaporite waters. Other major ions do not distinguish the Gulf Wells, but the wells share some characteristics with evaporite waters, which clearly do distinguish them. In ion proportions, they are similar to Ojos Calientes waters and dissimilar to most others. Also, potassium, lithium, and boron concentrations are distinctly greater than those of any of the nonevaporite springs and are only slightly diluted compared with Indian Hot Springs, Rancho Cipres, or Ojos Calientes. The Gulf Wells produce from Cretaceous limestone; none of the other springs from limestone have high concentrations of the ions in question. Although the wells lie outside the Jurassic evaporite basin (fig. 21), folding and thrust faulting along slip planes in evaporites during Laramide deformation have transported evaporite rocks beyond the original basin margin (DeFord and Haenggi, 1971). Thrust faults in the vicinity of the Gulf

Wells indicate that evaporites exist in the subsurface. Water discharged from the Gulf Wells probably has been in contact with and derived a major part of its dissolved solids from evaporites.

Red Bull Springs, approximately 5 km northwest of Indian Hot Springs, should also fall within the evaporite basin (fig. 21). Its concentration of total dissolved solids (1,000 mg/l), although slightly greater than that of most obviously nonevaporite springs, is lower than the concentration of Indian Hot Springs. It has moderately high sulfate, sodium, and bicarbonate contents, but low chloride and lithium contents. Only a minor percentage of its water and dissolved salts could be derived from evaporites. Possibly its circulation system is too shallow to contact evaporites.

The hot springs at Peguis are also at the edge of the evaporite basin (fig. 21), but the chemistry of Peguis water falls within that of the carbonate group discussed previously. A 3,660-m-deep (12,000 ft) Pemex exploration well drilled 7 km north of the springs did not penetrate evaporites, although it was interpreted as having reached Paleozoic strata (Diaz, 1964). Either Peguis is outside the Jurassic basin or its convection system is too shallow to reach evaporites.

No cold spring or well waters are chemically similar to the evaporite waters. Evaporites apparently do not occur in the shallow subsurface where cold, shallow ground water could reach them. The chemistry of the evaporite waters is determined more deeply within the circulation system than is the chemistry of other thermal waters.

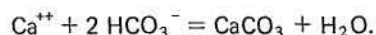
### CALCITE SATURATION

Except for one hot spring and one cold spring, all the waters sampled in the study are in equilibrium or oversaturated with calcite. Saturation indices (table 3) are equal to the  $\log_{10}$  of the activity products of calcium and carbonate divided by the solubility product for the spring temperature. These indices were calculated by WATEQF, a Fortran version of a computer program for interpreting water analyses (Truesdell and Jones, 1964). Water with indices between  $\pm 0.1$  are generally considered saturated; waters with indices above or below this range are oversaturated and undersaturated, respectively.

Oversaturated waters were probably in equilibrium in the subsurface but loss of dissolved carbon dioxide near or at the surface raised the pH, leading to oversaturation. The reaction is



Precipitation of calcite then could occur by the following reaction:



Some spring waters oversaturated with respect to calcite do not precipitate calcite, presumably because of slow reaction ratio and rapid dispersal or dilution of the water. Only the evaporite waters, which have extremely high partial pressures of carbon dioxide, precipitate travertine, with the exception of the springs at San Carlos and possibly Las Cienegas. Rapid loss of carbon dioxide from evaporite waters leads to oversaturation and precipitation, as occurs in Indian Hot Springs waters. Most of the springs discharge at the bottoms of pools that bubble gas, presumably carbon

dioxide. Stump Spring discharges rapidly to the surface and has the highest temperature, lowest pH, and highest partial pressure of carbon dioxide of the Indian Hot Springs. Its saturation index indicates that the water is in equilibrium with calcite (table 3). The other springs are oversaturated with calcite, have lower temperatures and partial pressures of carbon dioxide, and have higher pH values. All these factors reflect the loss of carbon dioxide.

There are some important implications of calcite equilibrium. Calcite solubility decreases with increasing temperature (Blount and Dickson, 1969); therefore, many high-temperature thermal waters have low calcium concentrations and are undersaturated with calcite when they cool near the surface. Increased partial pressure of carbon dioxide increases calcite solubility so that there are actually two opposing factors in calcite solubility. Springs that are in equilibrium with calcite and have not lost much carbon dioxide must have reached equilibrium approximately at the spring's surface temperature. These thermal waters either were never much hotter than their surface temperatures or re-equilibrated near the surface during cooling. In the latter case, any record of equilibrium reached at higher temperature at greater depth would have been lost.

Partial pressures of carbon dioxide in West Texas waters range from  $2.2 \times 10^{-4}$ , near the atmospheric equilibrium value of  $3.2 \times 10^{-4}$  (Keeling, 1958), to  $2 \times 10^{-1}$ , greater by several orders of magnitude than atmospheric equilibrium. The source of this excess carbon dioxide is uncertain. High partial pressures of carbon dioxide in ground water are commonly attributed to water percolation through a soil zone enriched in carbon dioxide from the decay of organic matter (Garrels and McKenzie, 1967; Hem, 1970). Soils in Trans-Pecos Texas are thin and rocky with little or no vegetation, and high partial pressures of carbon dioxide in these soils are unlikely. Metamorphism of limestone, a source of carbon dioxide which is commonly cited in the Russian literature (Hem, 1970), implies extremely high temperatures at depth. Other possible sources of carbon dioxide are unlikely (for example, reduction of sulfate by hydrocarbons).

### Stable Isotope Geochemistry

#### INTRODUCTION

Hydrogen and oxygen each have more than one naturally occurring stable isotope. Hydrogen has an isotope of mass 1, common hydrogen ( $\text{H}$  or  $\text{H}^1$ ), and of mass 2, deuterium ( $\text{D}$  or  $\text{H}^2$ ). Oxygen has three isotopes, mass 16 ( $\text{O}^{16}$ ), mass 17 ( $\text{O}^{17}$ ), and mass 18 ( $\text{O}^{18}$ ). Only  $\text{O}^{16}$  and  $\text{O}^{18}$  are usually reported. Isotope ratios of both hydrogen and oxygen vary in nature and provide information about the origin and history of the natural materials in which they occur.

For practical reasons, hydrogen and oxygen isotope compositions are expressed as deviations from a standard, which for water is Standard Mean Ocean Water (SMOW) (Craig, 1961b). The deviation ( $\delta$  or  $\delta$ ) is reported as a parts per thousand (per mil or ‰) difference between the isotope ratio of the sample and SMOW.

Natural waters are depleted of  $\text{D}$  and  $\text{O}^{18}$  compared with SMOW. The lighter isotopes are concentrated in the vapor during evaporation; during condensation, the heavier



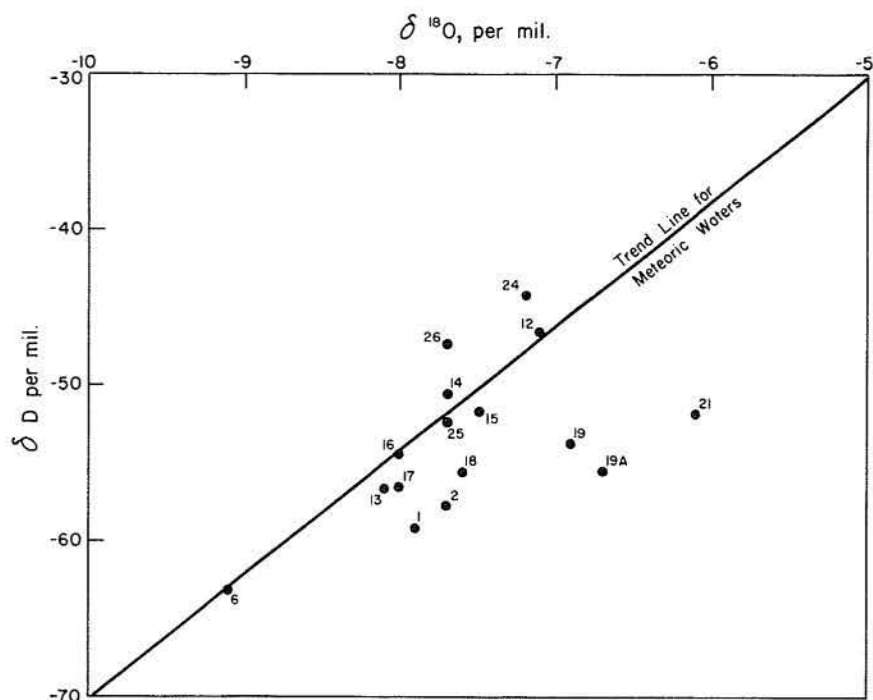


Figure 22. Isotopic composition of thermal and nonthermal waters, Rio Grande region of Texas and Mexico.

isotopes are concentrated in the condensate. Atmospheric moisture is thus continuously being depleted of D and  $O^{18}$  as it moves away from the ocean source. The amount of depletion increases with distance from the source. Meteoric waters demonstrate a constant relationship between  $\delta$  D and  $\delta$   $O^{18}$ , shown in figure 22 as the meteoric water line. Sheppard and others (1969) have mapped the distribution of  $\delta$  D in meteoric water of North America.

Many high-temperature geothermal waters plot to the right of the meteoric water line. Thermal waters have the same hydrogen-isotope composition as meteoric waters in the area but are enriched in  $O^{18}$ . This oxygen shift is attributed to isotopic exchange between heated meteoric water and  $O^{18}$ -rich wall rocks. There is no shift in hydrogen-isotope composition because hydrogen concentrations in rocks are low compared with those in water.

## RESULTS

Table 4 and figure 22 show the isotopic composition of thermal and nonthermal waters from the Rio Grande area. The numbers in both figure 22 and table 4 are the same as those used for chemical analyses. Most waters sampled in this study plot along the meteoric water line. A few (significantly, only evaporite-related waters) plot off the line to the right.

All the analyzed thermal and nonthermal sodium bicarbonate waters and calcium-magnesium-bicarbonate-sulfate waters fall along the trend line. The slight scatter is within natural limits. Along with the previous evidence of water chemistry and geologic setting, the isotopic compositions show that thermal waters are heated meteoric waters. The isotopic compositions are consistent with those mapped by Sheppard and others (1969).

A slight geographic variation is shown by the D and  $O^{18}$  depletion of Red Bull and Indian Hot Springs water relative

to those in the Presidio Graben. The 10-per-mil scatter in  $\delta$  D for the Presidio waters cannot be attributed to geographic variation. However, the scatter is not analytical because the waters do follow the meteoric water line. The lack of an oxygen shift indicates that the waters either have a short residence time underground or, more likely, were never heated to temperatures sufficient for measurable exchange—which is consistent with geothermometry determinations presented in this section.

The hydrogen-isotope composition of Peguis water is slightly different from the composition of all other waters sampled for this study. Because the oxygen composition has not been determined, this water is not plotted on figure 22. Local meteoric water cannot be the source of the Peguis spring water. The springs at Peguis lie within the floodplain of the Rio Conchos, which drains the Sierra Madre Occidental. Meteoric water from the Sierra Madre Occidental, a mountain chain along the west coast of Mexico, should be isotopically light compared with Peguis-area meteoric water. Although analyses of Rio Conchos water are not available to substantiate this hypothesis, the lower  $\delta$  D appears to demonstrate that the Rio Conchos is recharging the Peguis thermal system. The isotopic composition of Indian Hot Springs water, however, indicates that recharge is probably not from the Rio Grande.

Most of the evaporite waters (fig. 21, nos. 1, 2, 19, 19a, 21) do not plot on the meteoric water line. Gulf-Presidio water (fig. 22, no. 17) cannot be distinguished from meteoric water; Gulf-Swofford water (fig. 22, no. 18) is at the approximate limit of natural variation. The evaporite waters are enriched in  $O^{18}$  like the high-temperature waters previously discussed, but the amount of exchange is much less than in most high-temperature spring waters. Geothermometry calculations for the waters from Indian Hot Springs (fig. 22, nos. 1 and 2) and Rancho Cipres (fig. 22, no. 21) indicate subsurface temperatures no greater than

Table 4. Hydrogen and oxygen isotope compositions of thermal and nonthermal waters.

#	Spring or well	$\delta D$ SMOW ‰	$\delta O^{18}$ SMOW ‰
1	Indian Hot Springs—Stump	- 59	-7.9
2	Indian Hot Springs—Chief	- 58	-7.7
6	Red Bull Springs	- 63	-9.1
11	Peguis	-120	
12	Las Cienagas	- 47	-7.1
13	Hot Springs—Ruidosa	- 57	-8.1
14	Briscoe Well	- 51	-7.7
15	Nixon Springs	- 52	-7.5
16	Capote Springs	- 54	-8.0
17	Gulf—Presidio	- 57	-8.0
18	Gulf—Swafford	- 56	-7.6
19	Ojos Calientes #3	- 54	-6.9
19A	Ojos Calientes #1	- 56	-6.7
21	Rancho Cipres	- 52	-6.1
24	Vizcaino Well	- 44	-7.2
25	Naegle Springs	- 52	-7.7
26	Mexican Springs	- 47	-7.7

approximately 60°C. The oxygen shift must have some cause other than high-temperature exchange.

There are several possible explanations for the oxygen shift. Meteoric water near Indian Hot Springs probably has an isotopic composition like that of Red Bull Spring water (fig. 22, no. 6). If so, Indian Hot Springs waters show both an oxygen and a hydrogen shift unlike high-temperature water, which shows only an oxygen shift. A line connecting Red Bull and Indian Hot Springs waters follows a trend line produced by evaporation in an enclosed basin (Craig, 1961a). Residual basin waters are enriched in both D and  $O^{18}$ .

There is no reason to suspect that Indian Hot Springs recharge water evaporated partially before entering the system, so the isotopic shift must have occurred during deep circulation, possibly by mixing meteoric water with an isotopically heavy formation water within the evaporites. The persistence of formation water in Jurassic or older rocks during considerable structural deformation is surprising.

Meteoric recharge water for Ojos Calientes, Gulf-Swafford, and Rancho Cipres ought to have an isotopic composition within the range of that of Presidio Graben waters (fig. 22, nos. 12, 13, 14, 15, 16, 24, 25, 26). If so, the oxygen shift for these waters does not follow an evaporation line. Mixing of formation water to produce the shift is even less likely. Exchange of oxygen with wall rocks, observed for many high-temperature thermal systems,

is possible for Ojos Calientes and Gulf-Swafford because geothermometry calculations for these waters indicate temperatures greater than about 100°C, but is not possible for either Indian Hot Springs or Rancho Cipres waters.

In summary, all the thermal waters in this study area are entirely or dominantly meteoric. All except the evaporite waters have retained meteoric isotopic composition. Mixing of meteoric water and an isotopically heavy formation water could produce the observed isotopic compositions of Indian Hot Springs waters, but is less likely for Presidio Graben evaporite waters. High-temperature exchange of wall rocks and meteoric water could produce the isotopic composition of Ojos Calientes and Gulf-Swafford waters but probably not the composition of Indian Hot Springs or Rancho Cipres waters.

### Conclusions

The chemistry of most individual hot springs waters of Trans-Pecos is similar to the chemistry of cold waters near the hot springs. The composition range and stable isotope ratios of the sodium bicarbonate and calcium-magnesium-bicarbonate-sulfate waters overlap with those of cold spring and well waters. These thermal waters must be local meteoric water whose overall chemistry is governed by the rocks with which they are in contact during the cycle of recharge, deep circulation, and discharge. This chemical similarity indicates that most of the dissolved constituents

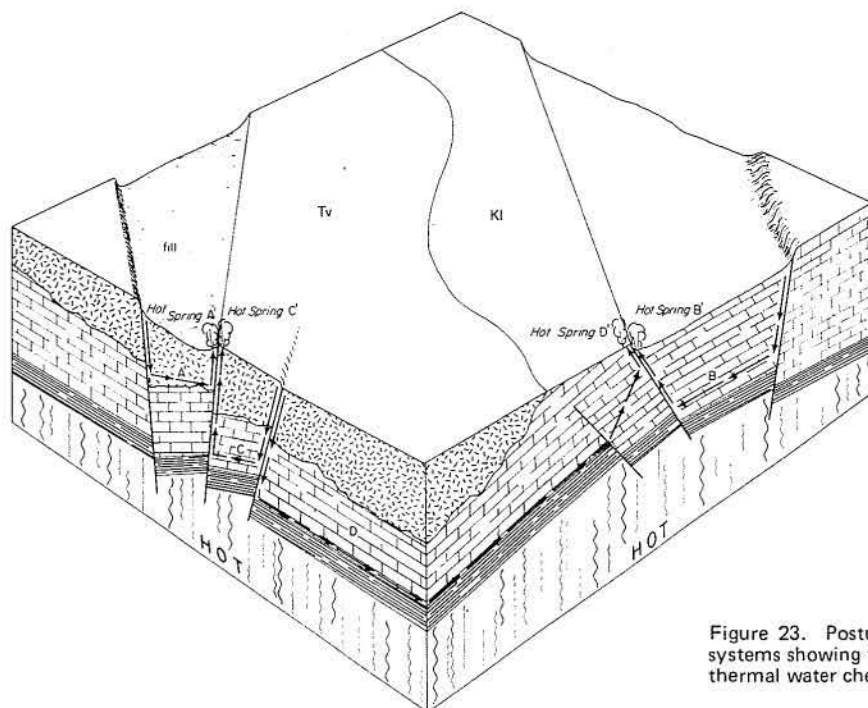


Figure 23. Postulated circulation of water in thermal convection systems showing fault control of flow paths and lithologic control of thermal water chemistry.

are acquired early in circulation history and that deep circulation and heating have little effect on water chemistry.

There are no cold spring waters chemically similar to the sodium-chloride-sulfate waters with high total dissolved solids. Because evaporites rarely crop out, circulation to a greater depth is required before the water can start dissolving evaporite minerals. Initially, recharge waters for these springs probably pass through limestones where they collect calcium, magnesium, and bicarbonate. Additional dissolved solids are acquired during deeper circulation and contact with evaporites. Stable isotope ratios support the view that these waters are also meteoric. The slight isotopic shift may be a result of mixing with a small component of an isotopically heavy formation water or, for Ojos Calientes and Gulf-Swofford, may result from high-temperature oxygen exchange with wall rocks.

All chemical compositions can be related to reasonable water/rock reactions. Intermediate compositions are created where recharge waters from different host rocks mix before returning to the surface or where a single flow path intersects more than one rock type. With the possible exception of carbon dioxide, no extra source besides rock leaching is required to provide all the dissolved constituents. Carbon dioxide may also be derived from rock leaching, but the exact chemical reaction is unknown. Proposed magmatic sources for some constituents, such as chloride and sulfate, are unnecessary and unlikely.

The presence of distinct chemical groups of thermal waters in a complex geologic terrain implies that flow paths are highly localized—that is, springs must be discharging water which was recharged and circulated within a very restricted geographic area. If circulation were broader, the springs would discharge waters that were more homogeneous. For example, in the northern part of Presidio Bolson

in Texas, the outcropping rock consists of Cretaceous carbonates bordered by basin fill on the south and rhyolitic volcanic rocks on the north (fig. 8). Four hot springs or wells separated by a total distance of 20 km—Capote Springs, Nixon Spring, Hot Springs, and Briscoe Well—have distinctive water chemistry derived from rock types exposed in the shallow subsurface near the spring and in the presumed highland recharge area adjacent to the springs. That they are not identical implies that the flow paths must be localized. The total recharge area for each spring must be no more than several tens of square kilometers. Circulation must also be shallow because vertical changes in rock type would also tend to homogenize the thermal water.

The concept of local circulation is illustrated schematically in figure 23. Flow path A is entirely in volcanic rocks; water discharged from hot spring A' would be a sodium bicarbonate water similar to the water of Capote Springs. Flow path B is entirely in limestone; water from spring B' would be composed of calcium-magnesium-bicarbonate-sulfate like those discharging from springs in Big Bend National Park. Flow path C intersects both volcanic rocks and carbonates; water discharged from spring C' should have an intermediate composition, like that of Briscoe Well, between a sodium bicarbonate water and a calcium-magnesium-bicarbonate-sulfate water. Mixing waters from flow paths A and B would also produce a water with intermediate composition. Probably no convection systems have long flow paths like path D, nor do any springs tap a homogeneous reservoir. The origin of these thermal waters differs from that of some other thermal waters, such as those at Yellowstone Park. Truesdell and Fournier (1976) interpret the variation in spring chemistry in Yellowstone as arising from the action of steam loss, dilution with cold water, and wall-rock reaction on a homogeneous reservoir water.



# GEOTHERMOMETRY

## Introduction

One of the few well-developed and commonly used tools in geothermal exploration is geothermometry. Geothermometry is based upon the assumption that some aspect of the chemical composition of the geothermal water is controlled by maximum temperature and that the composition can be used to determine maximum subsurface temperatures. Although the water may have cooled from its maximum temperature, the water chemistry will reflect that temperature.

Fournier and others (1974) presented five assumptions used in interpreting subsurface temperatures: (1) temperature-dependent reactions occur at depth; (2) chemical and mineralogical constituents involved in the reactions are abundant in rocks; (3) equilibrium occurs at the reservoir temperature; (4) reservoir composition is retained with little or no re-equilibration or change in composition at lower temperature; and (5) thermal water does not mix with cooler shallow ground water. Fournier and others (1974) also warn that these assumptions are not always valid and that interpretation of results is simpler for springs with high temperatures or discharges than for springs with low temperatures or discharges.

There are several qualitative geothermometers, but only two are sufficiently calibrated to be quantitative: the silica method, based on the solubility of silica, and the sodium-potassium or the sodium-potassium-calcium methods, based on equilibrium ratios of the elements.

## Quantitative Geothermometers

### SILICA METHOD

The solubility of silica in water increases with increase in temperature; silica concentration thus reflects maximum water temperature. Absolute solubility, however, is also a function of the silica phase in contact with water. Quartz is least soluble, followed by chalcedony, cristobalite, and amorphous silica, which is the most soluble.

The silica method, developed by Fournier and Rowe (1966) and Mahon (1966), assumes that equilibrium with quartz, attained at high reservoir temperatures, controls aqueous silica concentrations. For high-temperature systems (higher than 150°C) in most geologic settings, this assumption is reasonable. Quartz is abundant in most rock types, and reaction kinetics are rapid at high temperatures. For systems at lower temperatures, however, equilibrium may not occur or may be with another silica phase. At temperatures of 100°C or rarely to 180°C (Arnorsen, 1975), equilibrium may be with chalcedony. Water in contact with rocks containing amorphous silica may have very high concentrations of silica—above the solubility limits of either quartz or chalcedony (Klein, 1976). Given sufficient time and higher temperatures to increase the reaction kinetics, such water could precipitate silica to reach equilibrium with quartz.

Other problems with the method are precipitation of silica during cooling of thermal water and dilution by low-silica cold ground water. White (1970) indicates that

silica precipitates rapidly as water cools to about 180°C, but that the rate of precipitation drops greatly below that temperature.

### SODIUM-POTASSIUM-CALCIUM METHOD

The sodium-potassium-calcium method (Fournier and Truesdell, 1973) and the sodium-potassium method (White, 1965; Ellis, 1970) from which it was derived are based on ratios of ions. Equilibrium constants for reactions between sodium-, potassium-, and calcium-bearing phases are temperature dependent. Feldspars are generally believed to be the minerals governing equilibrium. Experimental work with alkali feldspars and chloride solutions (Orville, 1963; Hemley, 1967) and empirical studies of natural thermal waters (White, 1965; Ellis, 1970) demonstrate a relationship between sodium-potassium ratios and presumed equilibrium temperatures. Fournier and Truesdell (1973) developed an empirical sodium-potassium-calcium geothermometer which incorporates calcium but still assumes equilibrium with feldspars. They found a linear relationship between the function  $\log \frac{Na}{K} + \beta \log \frac{\sqrt{Ca}}{Na}$  and inverse temperature. The factor Beta ( $\beta$ ) is derived from the stoichiometry of feldspar reactions and is either 4/3 if the water equilibrated below 100°C, or 1/3 if the water equilibrated above 100°C.

Problems with the sodium-potassium and sodium-potassium-calcium methods can arise if equilibrium does not occur, or if equilibrium occurs with minerals other than feldspars. Although equilibrium can eventually be established at low temperatures (less than 50°C), equilibrium probably should not be assumed for waters cooler than 100°C. Equilibrium with minerals other than feldspars, such as zeolites (which are found in rocks of this study), does not follow the same pattern of temperature relationships. Subsurface temperatures for such waters cannot be predicted by the sodium-potassium-calcium method. Precipitation of calcite by travertine-depositing springs lowers calcium concentrations, and without re-equilibration of sodium and potassium the sodium-potassium-calcium method gives temperatures that are too high. Dilution affects sodium-potassium-calcium temperatures less than silica temperatures because sodium-potassium-calcium temperatures are based partly on ratios of ions.

## Qualitative Geothermometers

Many qualitative geothermometers indicate whether thermal waters are part of relatively high or low temperature systems or of hot-water or vapor-dominated systems (White, 1970). These geothermometers include calcium-bicarbonate ratios, magnesium concentrations and magnesium-calcium ratios, sodium-calcium ratios, chloride-fluoride ratios, and stable isotope ratios; an additional geothermometer is siliceous sinter deposition as opposed to travertine deposition. Furthermore, chloride concentrations and chloride-bicarbonate plus carbonate ratios can be used as indicators of mixing between thermal and nonthermal waters in groups of related springs.

Table 5. Calculated subsurface temperatures (°C)  
by Si and Na-K and Na-K-Ca methods.<sup>a</sup>

#	Location	Quartz	Chalcedony	Na-K	Na-K-Ca ( $\beta = 1/3$ )	Na-K-Ca ( $\beta = 4/3$ )
1	IHS Stump	92	62	133	182	208
2	IHS Chief	92	61	142	189	221
3	IHS Squaw	86	55	142	188	218
4	IHS Beauty	86	55	155	196	230
5	IHS Soda	64	32	125	166	155
6	Red Bull Spring	87	56	88	141	125
7	Spring at Rio Grande Village	65	33	123	128	42
8	Hot Springs—Big Bend	67	36	121	128	44
9	Big Bend Spring #2	68	36	122	129	44
10	San Carlos Springs	88	57	99	117	37
11	Peguis	67	35	84	116	50
11A	Peguis #2	66	34	82	114	48
12	Las Cienagas	91	60	68	120	84
13	Hot Spring—Ruidosa	86	55	181	174	111
14	Briscoe Well	90	60	178	169	99
15	Nixon Springs	95	65	86	128	84
16	Capote Warm Spring	88	57	14	70	64
17	Gulf—Presidio	122	94	202	198	163
18	Gulf—Swafford	159(176) <sup>b</sup>	135(154) <sup>b</sup>	206	203	178
19	Ojo Caliente #3	134	107	173	201	217
20	Ojo Caliente #4	129	102	168	197	208
21	Rancho Cipres	98	67	117	158	139
22	Hot Wells	63	30	35	101	74
23	Mimbres Well	90	60	178	169	99
24	Vizcaino Well	110	81	114	140	83
25	Naegele Springs	95	65	195	179	112
26	Mexican Springs	91	60	2	79	59
27	Nuñez Well	104	74	90	115	41
28	Alamo Springs	68	36	21	65	60

<sup>a</sup>See text discussion for interpretation of subsurface temperatures.

<sup>b</sup>Values in parentheses use published SiO<sub>2</sub> concentrations.

## Results

Table 5 presents subsurface reservoir temperatures calculated by the silica method assuming equilibrium with quartz and chalcedony and by the sodium-potassium and sodium-potassium-calcium methods using  $\beta$  equals 1/3 and 4/3. It is best to interpret the geothermometry results in relation to the geologic setting and geochemistry of the thermal waters.

### GEOOTHERMOMETRY OF COLD- TO INTERMEDIATE-TEMPERATURE SPRINGS AND WELLS

Subsurface-temperature estimates range up to 159°C assuming quartz equilibrium and up to 135°C assuming

chalcedony equilibrium. This is within the temperature range suggested by Arnorson (1975) as a transitional range between quartz and chalcedony equilibrium. The hottest springs and wells (Ojos Calientes and Gulf Wells) have the highest silica concentrations (76 to 144 mg/l) and reservoir temperatures (122° to 159°C, quartz equilibrium). However, cold springs and wells in the Presidio Basin have silica concentrations ranging from 22 to 60 mg/l (table 3; fig. 24). Equilibrium temperatures for this range of concentrations are 67° to 110°C for quartz equilibrium and 35° to 80°C for chalcedony equilibrium (table 5). With the exception of Ojos Calientes and Gulf Wells, the silica content of most thermal waters ranges only from 20 to 45 mg/l (fig. 24).

This seemingly contradictory relationship may be explained in several ways.

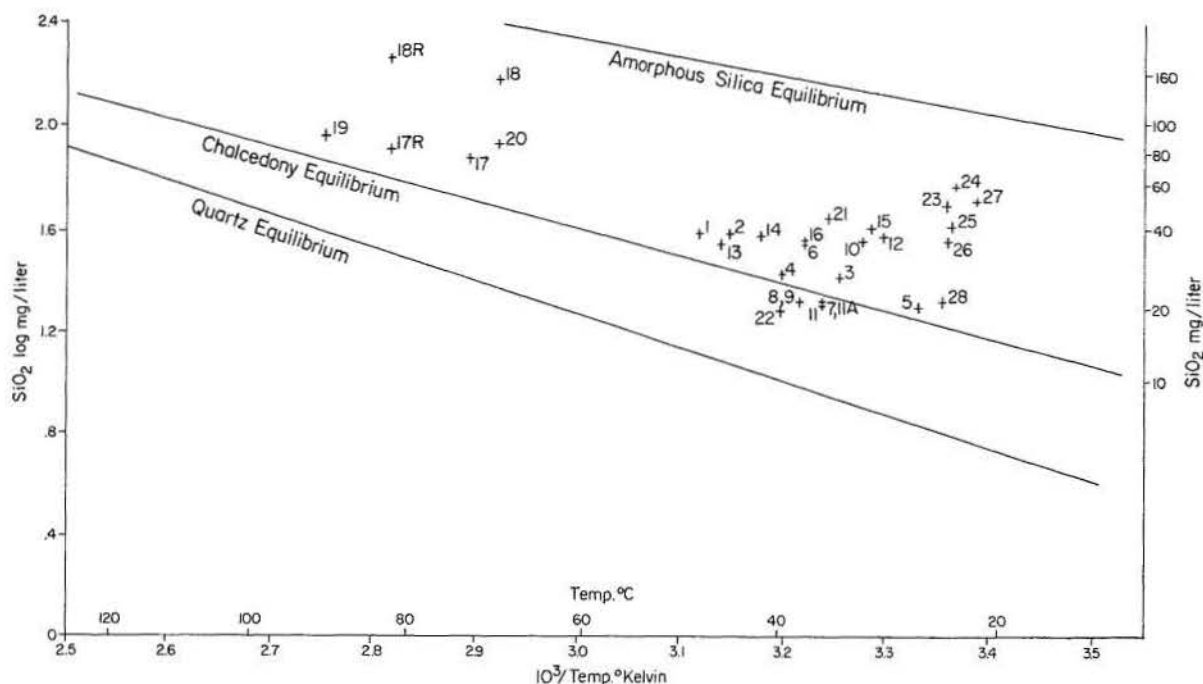


Figure 24. Measured temperature versus log silica concentration in thermal and nonthermal waters of this report. Silica phase equilibrium lines from Fournier (1976). The letter "R" after number indicates reported value from table 4.

First, the high-silica cold waters may actually be thermal waters which have been cooled before sampling either by heat conduction into wall rocks or by dilution with nonthermal ground waters. However, several of the cold water samples are from wells drilled near or into fault zones where thermal water could be upwelling (for example, Vizcaino, Mimbres, and Nuñez Wells). It seems unlikely that thermal waters in this setting could have cooled completely to ambient temperatures. Dilution with nonthermal water implies that the silica concentration of the supposed thermal water should be even higher than the measured silica concentration, implying still higher reservoir temperatures.

A second possibility seems more reasonable: that the waters are nonthermal and that their high silica concentration results from dissolution of amorphous silica. All the cold waters are oversaturated with respect to quartz and chalcedony but undersaturated with respect to amorphous silica (fig. 24). The waters are from springs or wells occurring either in bolson sediments composed of volcanic fragments or in siliceous volcanic rocks rich in volcanic glass and opal (Walton, 1975); both the volcanic glass and opal are highly soluble forms of silica. Hydrolysis of volcanic glass or silicate minerals releases silica into solution. If, as seems likely, the waters were never heated sufficiently to reach equilibrium with quartz, high silica concentrations could persist. The most logical explanation for the high-silica content of many of the waters in Trans-Pecos Texas is the dissolution of opal or volcanic glass from igneous rocks and the persistent nonequilibrium of the waters with quartz. This hypothesis is consistent with published analyses of cold ground water emanating from silicic igneous rocks for which silica concentrations ranged from 10 to 76 ppm (table 1 of White and others, 1963).

This conclusion creates problems in interpretation of thermal-spring geothermometry. Water with a silica content of 60 mg/l in equilibrium with quartz requires a reservoir temperature of 110°C. Yet the same water is undersaturated with respect to amorphous silica at 20°C. Assuming quartz equilibrium where it does not occur could be misleading in assessing the geothermal potential of an area. A common assumption is that diluting thermal waters with nonthermal waters will lower the silica concentration of the thermal waters. Actually, dilution with these nonthermal waters could, in fact, raise the silica content of most of the hot waters studied for this report.

The silica content of the intermediate-temperature (30° to 47°C) warm springs varies only between 20 and 45 mg/l (fig. 24). Among these samples, most of which have a silica concentration of approximately 40 mg/l, there is no apparent increase in silica concentration with increase in temperature. This relationship implies two possibilities: (1) thermal waters for most of the samples come from the same reservoir water, which has a silica concentration of approximately 40 mg/l; or (2) the samples represent waters that have circulated to similar depths in areas of similar geothermal gradient. Both possibilities are probably important. All but two of the thermal springs or wells with a silica content of approximately 40 mg/l are located in the northern part of Presidio Bolson or the Indian Hot Springs area. For each area, the first possibility, a related reservoir, may be important. The second possibility, similar depth of circulation in a similar heat-flow province, is probable despite the considerable differences in water chemistry among springs in the two areas.

Are the waters in equilibrium? If they are, what are their reservoir temperatures? The cold waters in the northern Presidio Basin, which presumably recharge the thermal



waters, have similar if not greater concentrations of silica than the thermal waters; the silica concentrations in the thermal waters could simply be inherited from the recharge waters. Recharge waters for the Indian Hot Springs and for several springs from Big Bend National Park probably do not have high silica concentrations because they do not contact silicic volcanic rocks. Inheritance of high silica concentrations should not be a problem.

For all intermediate temperature waters the tight clustering of silica concentrations around 40 mg/l suggests that the waters are tending towards equilibrium. Many of the samples plot very near the chalcedony equilibrium curve. This clustering could be coincidental, and Fournier and others (1974) point out that an assumption of equilibrium at low temperatures is tenuous. However, chalcedony equilibrium at spring temperatures and at proposed reservoir temperatures is consistent with the observations of Arnorson (1975) and Fournier and Truesdell (1970).

Equilibrium with chalcedony indicates a reservoir temperature of no more than 60°C (well below the temperature necessary for power generation). The maximum temperature reported for any of these thermal systems is 60°C in a shallow well at Indian Hot Springs (Dorfman and Kehle, 1974) and is consistent with the proposed chalcedony equilibrium. Equilibrium with chalcedony is also consistent with observations of Fournier and Truesdell (1974) that warm springs with moderate to high rates of discharge have reservoir temperatures only slightly greater than their surface temperatures. Many of the springs have surface temperatures of approximately 40° to 47°C. Two of the springs with the lowest flow rates, Nixon and Las Cienegas, which presumably could have cooled the most from their reservoir temperatures, also have the lowest surface temperatures. Equilibrium with chalcedony is consistent with the mineralogy of the various reservoir rocks. Reservoir rocks at Indian Hot Springs and Big Bend National Park include Cretaceous limestones that contain chert. Bolson sediments commonly contain chalcedony nodules (Groat, 1972), and volcanoclastic sediments contain a variety of silica minerals (Walton, 1975).

The warm springs in Big Bend National Park and at Peguis on the Rio Conchos contain approximately 20 mg/l of silica and plot near the chalcedony curve (fig. 24). These springs are probably in contact exclusively with Cretaceous limestones containing chert. They have high flow rates (equal to or greater than 400 l/min), are likely to be in equilibrium with chalcedony, and have low calculated reservoir temperatures approximately equal to their surface temperatures (40°C).

Dilution of these waters is a possibility. Fournier and Truesdell (1974) indicate that mixing with shallow, cool ground water is the most likely failure in the five assumptions listed above. The narrow range in silica concentrations of the warm spring waters could indicate only minimal dilution of thermal water with a maximum silica concentration of 40 to 50 mg/l. Several of the springs (for example, Capote and Nixon Springs) emanate from fractures in bedrock which provide permeability in otherwise impermeable crystalline rocks. Near-surface mixing for

these springs is highly unlikely. Deeper mixing is harder to determine but presumably would involve only thermal waters. Mixing in some thermal systems would be difficult to detect because the thermal and nonthermal waters are chemically similar. In addition, many of the cooler waters have high silica concentrations. Mixing might not lower the silica concentrations of hot spring waters and could even raise them. Soda Spring of the Indian Hot Springs group is the only spring for which mixing with nonthermal water can be established conclusively. The temperature, silica concentrations, distinctive chemistry (table 4), and physical setting of Soda Spring all demonstrate mixing. Mixing probably has not significantly affected any of the intermediate-temperature waters.

The sodium-potassium-calcium method for calculating reservoir temperatures cannot be used to evaluate temperatures calculated by the silica method. The sodium-potassium-calcium method requires that waters be in equilibrium with feldspars (Fournier and Truesdell, 1973), but many of the spring waters sampled for this study have been in contact with evaporites or zeolitized volcanic rocks which contribute sodium, potassium, and calcium to the water. For this reason temperatures calculated by the sodium-potassium-calcium method are not meaningful.

The methods for calculating subsurface temperatures from sodium-potassium-calcium geothermometry allow a sufficient range in temperatures to overlap with most preconceived subsurface temperature ranges. For example, sodium-potassium-calcium temperatures of 44°C for Hot Springs in Big Bend National Park and of 48° to 50°C for the warm spring at Peguis correspond well with silica-chalcedony equilibrium temperatures and surface temperatures. The agreement, however, may simply be fortuitous because equilibrium at such low temperatures is unlikely, even discounting problems that arise from evaporite solution.

Many estimated temperatures for sodium-bicarbonate waters determined by the  $\beta$ -equals-4/3 method correspond well with chalcedony temperatures for the same waters; however, many other estimated temperatures do not. Attempts to explain the results of the sodium-potassium-calcium method can lead to circular reasoning. It is better to recognize that the method is ineffective for these thermal systems.

Stable isotope ratios for most of the intermediate-temperature waters plot on the meteoric water line (fig. 22), indicating maximum temperatures less than approximately 100°C or a short residence time. The cause for the slight isotopic shift for Indian Hot Springs and Rancho Cipres waters is uncertain but is probably not high-temperature exchange.

In this study, the high silica concentration in many cold ground waters, which is equal to or greater than the silica content in all but three of the thermal waters in the area, is due to solution of amorphous silica. This finding greatly complicates interpretation of the thermal-water geothermometry. Nevertheless, the intermediate-temperature waters apparently are approximately in equilibrium with chalcedony and have only modest reservoir temperatures of about 60°C, which is consistent with the isotopic results.

Even if equilibrium were with quartz, reservoir temperatures would be only about 90°C.

Thermal waters in the areas of Presidio and Hueco Bolsons have higher silica content and, thus, higher reservoir temperatures than thermal waters in the Big Bend area. Although this relationship may be caused by solution of amorphous silica, it is consistent with the geologic setting of the area. Presidio Bolson and Hueco Bolson are in areas of recent faulting and relatively high thermal gradient and have the only known high-temperature thermal systems. There is no evidence of recent faulting in the Big Bend area, and heat flow and thermal gradients are apparently lower than in the Presidio area.

### **GEOOTHERMOMETRY OF GULF WELLS AND OJOS CALIENTES**

The geothermometry of Gulf Wells and Ojos Calientes needs to be discussed separately because they represent the only known thermal systems that show potential for power generation. Estimated silica temperatures are 112° and 94°C for Gulf-Presidio, 159° and 135°C for Gulf-Swofford, and 134° and 107°C for Ojos Calientes for quartz and chalcedony equilibrium, respectively (table 5). Published silica concentrations of water from Gulf-Swofford (White and others, 1977) give temperatures of 176° and 154°C for quartz and chalcedony equilibrium. At these higher temperatures, equilibrium is more likely than at the lower temperatures of the intermediate-temperature springs. Although most geothermometry studies, including some in the Rio Grande Rift area, have assumed quartz equilibrium (Miller and others, 1975; Pearl and Barrett, 1976), Arnorson (1975) states that the temperature range from 100° to 180°C is transitional between chalcedony and quartz equilibrium. Interpretation of the geothermometry is arguable. Plots of silica concentration and surface-temperature of Ojos Calientes and Gulf-Presidio (fig. 24), like those of the intermediate-temperature springs, are near the chalcedony-equilibrium curve, which suggests that maximum temperatures for the waters are only slightly higher than their surface temperatures.

The results of other geothermometers are also equivocal. Temperatures derived by the sodium-potassium-calcium method are all approximately 200°C. However, the sodium-potassium-calcium content of all three waters is controlled by evaporite solution and is not a reliable indicator of subsurface temperatures. Isotopic ratios suggest high temperatures (greater than 100°C) for Ojos Calientes waters, which show an isotopic shift; isotopic ratios possibly suggest high temperatures for Gulf-Swofford water, which shows a slight oxygen shift. However, these ratios do not suggest high temperatures for Gulf-Presidio waters, which are isotopically similar to meteoric water.

All three waters deposit travertine. Calcite solubility decreases with increasing temperature so that subsurface temperatures may not be much above the surface-discharge temperatures of the waters. However, the waters could have re-equilibrated with limestone during cooling from their maximum temperatures. Re-equilibration should be expected

because the producing reservoir for both Gulf Wells is Cretaceous limestone and because Ojos Calientes rises along a fault cutting Cretaceous limestone. Travertine deposition simply indicates that the waters last equilibrated with calcite at near-surface temperatures. For the same reason, magnesium concentrations and magnesium-calcium ratios cannot be used as qualitative indicators.

Petrographic study of the Ojos Calientes travertine deposits does not reveal any siliceous deposits. Sinter is commonly deposited by springs that discharge water with greater than 180 mg/l silica (White, 1970). By this criterion, none of the thermal springs or wells should or does deposit siliceous sinter.

Water from Gulf-Swofford is far from being in equilibrium with either quartz or chalcedony. Its high silica concentration and implied high equilibrium temperatures are surprising in two respects. First, the well produces water at only 80°C. Wells drilled into high-temperature reservoirs generally produce water with approximately the same temperature as the reservoir. The temperature log of Gulf-Swofford indicates that the hot-water-producing zone must tap convective water circulating from depths below the reservoir. It is possible that the sampled water was cooled by conduction or by mixing with cooler ground water at the higher level, but such mixing would require still higher reservoir silica concentration and temperature.

The other surprising aspect of the water's high silica content is that, although the chemistry and geologic setting of both Gulf Wells are similar, implying similar histories, their silica contents differ by a factor of almost 2. Gulf-Presidio water is somewhat more dilute than Gulf-Swofford water but only by about 25 percent for most constituents other than silica. It is conceivable that the diluting water is low in silica but that the concentrations of other dissolved constituents are only a little lower than their concentrations in the thermal water. Another explanation is that Gulf-Presidio is farther from the source of thermal water than is Gulf-Swofford; silica therefore has precipitated during cooling, yielding water with lower silica concentration.

No clear conclusions about the high silica concentrations and the high implied equilibrium temperature can be made from the available information. Ojos Calientes and Gulf-Presidio may be in equilibrium with chalcedony or with quartz at maximum subsurface temperatures around 100° or 130°C, respectively. Dilution of the water, implying still higher silica concentrations and subsurface temperatures, is also possible. On the basis of its geologic setting and water chemistry, Gulf-Swofford ought to have reservoir temperatures similar to those of Gulf-Presidio. That it apparently does not may mean that a chemical record of higher temperatures has not been well preserved in Gulf-Presidio, or that the high silica concentrations in Gulf-Swofford have some source other than high reservoir temperatures. Because of the potential for geothermal power generation indicated by these three thermal systems, they deserve more thorough evaluation than can be obtained from surface studies.

Many geochemical studies (for example, Pearl and

Barrett, 1976; Truesdell and Fournier, 1976) now rely on mixing models like that developed by Fournier and Truesdell (1974) to determine the temperature and fraction of a hot-water component in waters that have undergone dilution with cold water. As indicated by Fournier and Truesdell, thermal waters with high discharge rates can originate either by deeply circulating meteoric water which is heated only to its approximate discharge temperature, or by mixing a much hotter water with cold, shallow ground water.

Whether or not mixing has occurred is generally determined by the sodium-potassium-calcium geothermometer. Water with a sodium-potassium-calcium temperature within 25°C of the discharge temperature is in equilibrium and has probably not been diluted, whereas water very much out of equilibrium probably has been diluted. Because of the problems of applying the sodium-potassium-calcium method, this test cannot be used for either Ojos Calientes or the Gulf Wells.

More qualitative tests on the available data, such as analyzing variations among nonreactive constituents (for example, chloride variation with temperature) in related waters, suggest that water from Ojos Calientes has not been diluted but that waters from the Gulf Wells have been mixed. Waters that have been mixed ought to show a regular variation of chloride content and temperature. Chloride content from the two sampled springs at Ojos Calientes is very similar, although the temperatures of the two springs differ appreciably. Also, the lower pH and higher bicarbonate content and partial pressure of carbon dioxide of Ojos Calientes number 4, the cooler of the two, suggest that this spring should have experienced less dilution than did Ojos Calientes number 3. There are numerous springs at Ojos Calientes, and analysis of only two is not conclusive. Additional sampling to evaluate mixing would be relatively simple and inexpensive and should be done.

The chemical compositions of water from the two Gulf Wells are similar, and the wells tap the same reservoir, the Loma Plata Limestone. However, as the wells are physically separated, the waters may not be genetically related, in which case mixing models using the two may not be valid. Nevertheless, their water chemistry does suggest that water from Gulf-Presidio is a more dilute version of water from Gulf-Swafford. Dilution would have had to have taken place at a depth of at least 900 m, the producing level of the wells, and could have occurred as the rising thermal water displaced cooler water within the aquifer. If so, the diluting water is likely to be a calcium-magnesium-bicarbonate water, but it cannot be sampled. Mixing at depth is particularly difficult to evaluate. Because of its implications in geothermometry, the possibility of mixing should be considered more thoroughly.

### ***Comparison with Geothermal Indicators in the Rio Grande Rift***

The Rio Grande Rift in Colorado and New Mexico is considered an area with potential for geothermal

development because of its extremely high heat flow (Reiter and others, 1975). The Rio Grande area of Trans-Pecos Texas has a similar geologic setting, and existing geothermal indicators in the two areas are comparable. With the exception of springs in the vicinity of the Valles Caldera in northern New Mexico, no hot springs in the Rio Grande Rift have silica concentrations above approximately 100 mg/l (Summers, 1976; Pearl and Barrett, 1976). The thermal activity at Valles Caldera is probably related to young silicic igneous activity (Ross and others, 1961; Smith and others, 1970). Other thermal systems in the Rio Grande Rift probably result from deep circulation of ground water in a region of high heat flow, similar to the origin of hot springs in Trans-Pecos Texas. Published temperatures and silica concentrations of thermal waters near Socorro and Truth or Consequences, New Mexico, are no more than 45°C and 45 mg/l (Summers, 1976), comparable to the intermediate-temperature group of springs in Texas. Radium Hot Springs has maximum temperatures of 60°C and silica concentrations of up to 75 mg/l (Summers, 1976). Southern New Mexico is an area of demonstrably high heat flow (Reiter and others, 1975; Decker and Smithson, 1975). The similarity in thermal water geothermometry suggests that the Rio Grande area of Texas and Mexico is also an area of high heat flow.

Other than in the Valles area, the hottest springs and highest silica concentrations occur in several thermal areas of the rift in Colorado. Mount Princeton Hot Springs, Poncha Hot Springs, and hot springs in the upper San Luis Valley have maximum surface temperatures ranging from 60° to 82°C and maximum quartz equilibrium temperatures ranging from 100° to 125°C (Pearl and Barrett, 1976). Pearl and Barrett did not provide data concerning silica concentrations, but these temperatures indicate concentrations of approximately 60 to 80 mg/l of silica. Olson and Dellechiaie (1976) gave maximum values of 85°C and 85 mg/l (130°C quartz equilibrium) for springs in the Mount Princeton group. All these values are comparable to temperatures and silica concentrations of Ojos Calientes and Gulf-Presidio. The silica concentrations of thermal water from Gulf-Swafford is considerably greater than any of the thermal systems in the identified Rio Grande Rift. By this comparison the Trans-Pecos Texas region has at least the geothermal potential of the Rio Grande Rift.

Pearl and Barrett (1976) assumed quartz equilibrium for all springs. Chalcedony equilibrium is more likely for some springs in Texas, and quartz equilibrium is possible for even the highest temperature thermal waters. Pearl and Barrett (1976) interpreted sodium-potassium-calcium and mixing-model geothermometers to indicate reservoir temperatures up to 200°C. These methods cannot yet be used to evaluate the thermal waters of this study. Nevertheless, available evidence shows that the Rio Grande area of Texas and Mexico has as much geothermal potential as does any part of the Rio Grande Rift.



## GEOTHERMAL MODEL AND AREA EVALUATION

### Geothermal Model

For Trans-Pecos Texas, a model of ground-water flow, circulation depth, source and magnitude of heat, and geothermal potential can be constructed using available evidence. Local meteoric water circulates downward through an intersecting net of permeable fractures created by late Tertiary to, in some areas, Recent crustal extension. Although most rock types are impermeable, some rock types, such as massive, cavernous limestones, are permeable yet unfractured. The water reaches depths where it is heated by the normal thermal gradient. The heated water, being less dense, is forced upward along fractures by descending cold water. Thermal water discharges to the surface as hot springs in areas where the water table intersects the surface; a large part of the hot water probably leaks into permeable rocks where the water table is below the surface.

Recent fault movement may be important in keeping fracture systems open to circulation. Many hot springs are in areas with recent fault scarps or seismic activity, such as Presidio Bolson and Hueco Bolson. However, some hot springs are located in areas with no evidence of recent faulting. In addition, Ojos Calientes, the spring with the hottest thermal activity in the area, is located along a major normal fault which shows no evidence of recent movement. Cooler hot springs on the side of the basin opposite Ojos Calientes are evidently supplied from a shallower depth, even though the springs are near several recent fault scarps.

Thermal water chemistry is controlled by reaction of the waters with surrounding rocks and is similar to the chemistry of cold ground water in the same area. No additional source of dissolved solids is required. The chemistry of the springs indicates numerous separate thermal convection systems.

The source of heat for the water is the earth's normal thermal gradient, which increases from Great Plains values in the eastern part of Trans-Pecos Texas to higher values—at least Basin and Range values—along the Rio Grande. Still higher heat flow, such as that along the Rio Grande Rift in New Mexico, is possible and even probable according to available data. Thermal gradients are at least  $30^{\circ}\text{C}/\text{km}$  and up to  $40^{\circ}\text{C}/\text{km}$  along the Rio Grande. The increase in thermal gradient and heat flow may be due to progressive thinning of the crust beneath Trans-Pecos Texas across the Great Plains Basin and Range structural province boundary. An additional source of heat may be derived from a thermal gradient increased by blockage of normal heat flow by low-conductivity rocks, particularly the thick, water-saturated basin-fill sediments.

### Presidio and Hueco Bolsons

According to available evidence, Presidio Bolson, its structural continuation to the north, and Hueco Bolson are the areas with the best potential for geothermal development in Trans-Pecos Texas. Presidio and Hueco Bolsons represent the most likely extension into Texas of the Rio Grande Rift and its associated high heat flow. Deep oil tests show thermal gradients as high as  $40^{\circ}\text{C}/\text{km}$  with maximum

temperatures of  $170^{\circ}\text{C}$  at a depth of 3.5 km. Presidio Bolson is a deep basin with more than 1 km of fill and possibly twice that much displacement along boundary faults. Recent scarps in several parts of the basin show that it is still subsiding.

Two groups of thermal waters in Presidio Bolson are distinguished by their estimated reservoir temperatures. Most waters belong to a lower temperature group with estimated subsurface temperatures of  $60^{\circ}\text{C}$ , assuming chalcedony equilibrium. Sixty degrees centigrade is well below the temperature required for any presently feasible energy production but is possibly sufficient for space and process heating. The second group, including Gulf Wells and Ojos Calientes, has distinctly higher reservoir temperatures: a minimum of  $100^{\circ}\text{C}$  and possibly a maximum of  $180^{\circ}\text{C}$ .

Given a reservoir temperature of  $60^{\circ}\text{C}$ , an average annual surface temperature of  $20^{\circ}\text{C}$ , and an estimated thermal gradient between  $30^{\circ}$  and  $40^{\circ}\text{C}/\text{km}$ , the depth of circulation of the lower temperature spring group can be calculated as between 1,000 and 1,300 m (3,300 and 4,300 ft). Low-conductivity water-saturated basin-fill sediments may prevent normal heat flow and raise the thermal gradients. If so, shallower circulation could produce the estimated high temperatures. Because thick Cretaceous shales may also have low conductivity, areas outside the sediment-filled portions of the basin might also have anomalously high thermal gradients.

From the surface temperature and estimated reservoir temperature and the water chemistry of Capote Springs, a test of this estimate can be made. The chemistry of Capote Springs indicates that its water circulates entirely within volcanic rocks. Cretaceous shales should occur at a depth of no more than 400 m (1,300 ft); water in contact with the shales should have markedly different chemistry from the observed chemistry of Capote Springs. Circulation to a depth of 400 m in an area with a thermal gradient of  $40^{\circ}\text{C}/\text{km}$  would raise water temperature only about  $16^{\circ}\text{C}$ , to about  $36^{\circ}\text{C}$ , almost identical to the discharge temperature of the springs but below the estimated reservoir temperature of  $60^{\circ}\text{C}$ . Either the estimated temperature is too high, circulation is to a greater depth, or the thermal gradient is higher than  $40^{\circ}\text{C}/\text{km}$ , possibly because of blockage of the normal thermal gradient.

Estimated depth of circulation for the higher temperature waters of Ojos Calientes and Gulf Wells varies considerably with assumed subsurface temperatures and thermal gradients. For a subsurface temperature of  $100^{\circ}\text{C}$  and a thermal gradient between  $30^{\circ}$  and  $40^{\circ}\text{C}/\text{km}$ , the required depth of circulation ranges from 2,000 to 2,700 m (6,550 to 8,750 ft). If subsurface temperatures are as high as  $180^{\circ}\text{C}$ , depth of circulation must be between 4,000 and 5,000 m (13,000 and 17,500 ft). Ground-water circulation to such great depths is more difficult to imagine, although Sammel (1976) estimated depths of circulation as great as 4,300 m (14,000 ft) for thermal waters near Klamath Falls, Oregon, which is also in the Basin and Range province. In an area of continued intensive brecciation caused by recent

fault movement, permeable fracture systems could stay open to these depths.

Hueco Bolson is one of the deepest basins in Texas; it has 2,740 m (9,000 ft) of fill as indicated by gravity and seismic data taken along the east side of the Franklin Mountains just east of El Paso (Mattick, 1967). A well drilled by El Paso Water Utilities penetrated 1,330 m (4,363 ft) of bolson fill (Davis and Leggat, 1967). Total bedrock relief may greatly exceed either figure. Gravity profiles by Decker and Smithson (1975) indicate Hueco Bolson in the vicinity of El Paso has a depth of 4 km. Earth-resistivity and aeromagnetic data of Gates and Stanley (1976) demonstrate that the basin becomes shallower to the south and dies out around the southern Quitman Mountains, which are the site of the only hot springs in the area. The basin is presumably still subsiding, although recent fault scarps have not been observed. Chan and others (1977), however, did locate several epicenters at the southern end of the basin west of the Quitman Mountains.

Chapin (1971) included the Tularosa-Hueco basin as a part of the Rio Grande Rift. Reiter and others (1975) and Decker and Smithson (1975) measured high heat flow in the New Mexico portion, and their data imply that high heat flow extends into Texas. Reiter and others (1975) report values around 1.5 to 2 HFU—possibly up to 2.5 HFU—in Hueco Bolson north of El Paso, and still higher values to the west. Decker and Smithson (1975) measured heat flow values of from 2.5 to 3.1 at Orogrande in New Mexico, which is at the northern end of Hueco Bolson. Heat flow in the Texas part of Hueco Bolson may be similar, although no evaluations have yet been made.

Hot springs occur only at the southernmost end of Hueco Bolson, at Indian Hot Springs and Red Bull Spring, in what is probably the shallowest part of the basin. Subsurface temperature indicators suggest reservoirs of only 60°C. The depth of circulation required to reach this temperature should be in the same range as that in Presidio Bolson—about 1,000 to 1,300 m (3,300 to 4,300 ft). If deeper circulation systems exist, they do not discharge to the surface. More sophisticated exploration techniques will be necessary to discover the thermal water in this area.

The higher temperature thermal systems examined for this study are at least comparable to and possibly higher in reservoir temperature than any of the purely convective thermal systems in the Rio Grande Rift. This observation implies two important points:

1. Heat flow along the Rio Grande in Trans-Pecos Texas and Mexico may be similar to heat flow in the Rio Grande Rift. A definition of the boundary of the Rio Grande Rift should probably include this region. Geophysical studies to determine deep crustal structure are necessary to validate such a conclusion.
2. Thermal systems in Texas are worthy of continued investigation as possible sources of geothermal energy.

### Other Areas

No other areas in Texas have as much potential for geothermal development as Presidio and Hueco Bolsons. The only hot springs in Texas not in either Presidio or Hueco Bolsons are along the Rio Grande in Big Bend National Park and south of the park in Mexico. Although

geothermal energy development will not take place in the park, information on the hot springs there helps determine the potential for geothermal energy in the general region.

The Big Bend area is part of the physiographic and structural Basin and Range province (Fenneman, 1946), but heat flow is apparently similar to that in the Great Plains (Swanberg and Herrin, 1976). Geothermometry of the hot springs indicates low reservoir temperatures. Silica and sodium-potassium-calcium subsurface temperatures apparently correspond and are close to measured discharge temperatures (approximately 40°C). Evidently the spring waters do not cool significantly during ascent, which is consistent with their large discharge.

Hot springs in the Big Bend region discharge water that is apparently heated by relatively shallow circulation in an area of normal heat flow similar to that of the Great Plains. At a thermal gradient of approximately 18° to 29°C/km and an average annual surface temperature of 20°C (70°F), circulation to a depth of from 0.7 to 1 km is sufficient for the water to reach 40°C. Temperatures high enough to exhibit potential for geothermal energy production probably occur only at relatively inaccessible depths. The hot springs in the park area are best used for the enjoyment of tourists.

The area from Lobo Valley to Marfa, although it has neither hot springs nor wells, is considered a potential geothermal area because of the high silica concentration of its cold, shallow ground water (Hoffer, 1977). Many irrigation wells in the Lobo Valley agricultural district and domestic wells around Marfa have high silica concentrations—as much as 80 mg/l. If these waters are in equilibrium with quartz, such values would indicate subsurface temperatures around 125°C. However, as discussed in the section on geochemistry of thermal waters the high silica concentration probably results from solution of amorphous silica in volcanic rocks (tuffs and tuffaceous sediments) or in basin fill derived from volcanic rocks. There is no evidence that the well waters were ever hot. Measured temperatures are probably maximum temperatures.

The geologic setting of the Lobo Valley area suggests some geothermal potential, however. The Lobo Valley is part of a series of en echelon grabens which includes Presidio Bolson. The valley is a deep basin with recently active normal faults (Belcher and Goetz, 1977) and at least 1,000 m of displacement along the west side. The basin becomes shallower to the south. A water test well 40 km northwest of Marfa penetrated no more than 380 m (1,250 ft) of fill (Gates and White, 1976), and near Marfa volcanic rocks crop out.

Deep circulation should be expected along faults in the Lobo Valley area, yet absence of high-temperature springs or wells suggests that such circulation is not taking place. However, the basin is undissected; springs of any kind are rare because heavy ground-water withdrawal for agricultural land has lowered the water table well below the surface. It is conceivable that hot water rises along some of the basin-edge normal faults but discharges into basin fill at the water table. By the time thermal water reaches an existing well it has cooled by mixing or conduction. Nevertheless, if thermal water does exist in the Lobo Valley area, it is surprising that no wells have tapped any.

Most heat-flow and thermal-gradient studies indicate values similar to those characteristic of the Great Plains

(Decker and Smithson, 1975). Kleeman (1977), however, did show one higher heat-flow value near Van Horn (2.4 or 3.0). The slight ambiguity in heat flow and the similarity in setting and proximity to the Presidio Graben warrant further investigation, although, according to available evidence, the area has little geothermal potential.

Recent fault scarps also appear in Salt Basin, the northern extension of the Lobo Valley - Presidio series of grabens (Belcher and Goetz, 1977). Geophysical measurements indicate a maximum of 760 m (2,500 ft) of fill (White and others, 1977). There are no hot springs or wells. Silica concentrations of ground water range from 10 to 20 mg/l. Adjacent highlands are composed of carbonate sediments; silicic volcanic rocks do not occur in the area so the water does not have a high silica content. For the same

reasons as those given for the Lobo Valley region, Salt Basin is not a potential area for geothermal development.

A heat-flow value of 2.0 measured by Decker and Smithson (1975) at Cornudas, New Mexico, just across the state line in the highlands bordering Salt Basin (fig. 3) seems to contradict this conclusion about Salt Basin. If this value is indicative of heat flow in the area, Salt Basin could be expected to have a high thermal gradient. Decker and Smithson (1975), however, also determined a heat-flow value of 1.0 north of Van Horn at the south end of Salt Basin. Thermal gradients are low in oil tests in the area (AAPG, 1975). There is no apparent explanation for the difference unless great changes in heat flow are possible over short distances. Although Decker and Smithson (1975) did not consider either value anomalous, ground-water flow might have disturbed the measured heat flow.

## ADDITIONAL WORK

Sufficient evidence for potential geothermal development exists in Trans-Pecos Texas to warrant additional investigations. More evaluation is needed rather than a development program. Most of the known hot springs or wells do not have much production potential. Only data from Ojos Calientes and Gulf Wells suggest high temperatures at depth, and even in these waters there is as yet no evidence of sufficiently high temperatures at accessible depths. Geothermal exploration, in general, is not well developed, and in Trans-Pecos Texas such exploration is still in the reconnaissance stage.

Knowledge of the geologic setting and most of the geochemical data necessary for exploration exist. Trans-Pecos Texas is exceptionally well mapped, and much geologic and geochemical data have been gathered for this report. However, geochemical data to evaluate the effects of dilution on the highest temperature thermal waters as well as geophysical information to define heat flow and crustal structure are lacking. Geochemical investigations of mixing thermal and nonthermal waters would be useful, as discussed in the section on geothermometry.

Important new insights can be gained from geophysical studies. However, geophysical methods designed explicitly for geothermal exploration are not well developed, and methods developed to discover buried magma chambers (Eaton, 1975) are not applicable in Texas. Nevertheless, much useful information could be gained from additional geophysical research in the area. A direct method of geophysical exploration might be to find additional areas of rising hot water by measuring thermal gradients in existing wells or in shallow wells drilled specifically for geothermal exploration in and near basin-margin fault zones. These fault zones are the conduits for most of the known thermal waters. Anomalously high thermal gradients in the shallow wells, even those which do not reach the water table, would indicate upwelling thermal waters. Some thermal waters which do not reach the surface have been discovered accidentally; there should thus be many as yet undiscovered thermal convection systems, particularly in Presidio and

Hueco Bolsons. Earth resistivity or self-potential measurements could be used to locate upwelling thermal waters, but interpretation of this data would be complicated by the heterogeneity of the rocks and ground water in the study areas (Gates and Stanley, 1976).

Although more thermal water should be discovered by these methods, it is possible that there are no additional hidden high-temperature systems. Private exploration companies have measured gradients in many existing wells and in wells they drilled. Information from these studies is, of course, confidential, but the absence of continued interest by the companies suggests that results were not encouraging.

Thermal-gradient and shallow-heat-flow studies and geochemical analyses are undertaken to examine existing thermal-convection systems. However, hot springs and wells are not ideal exploration subjects. Most hot springs of the Rio Grande area discharge meteoric water that circulates to moderate depths and is heated to moderate temperatures. Depths of circulation have been estimated, but data that could accurately define those depths are unavailable. Deep circulation of water in low-thermal-gradient areas of the world produces hot springs, although in those areas temperatures hot enough for power production exist only at economically inaccessible depths. On the other hand, adequate temperatures could possibly exist at accessible depths in Trans-Pecos Texas even though the present hydrothermal circulation either does not reach that depth, or waters from that depth do not reach the surface. Knowledge of the thermal gradient and thermal structures at depth is necessary to determine whether sufficiently high temperatures do exist and in what settings they are closest to the surface. Such information can be obtained in two ways: indirectly by geophysics or directly by drilling. Geophysical studies ought to precede expensive drilling.

Applicable methods of investigation include gravity, magnetic, and seismic studies to delineate subsurface structure and heat-flow studies to determine thermal gradients. For the same reasons that knowledge of thermal



gradients in shallow wells could be useful in detecting hidden thermal convection, such knowledge is a poor indicator of regional heat flow or thermal gradient. Convection of meteoric water is occurring over much of the area. Cold recharge water moving downward would depress the thermal gradient, whereas upwelling thermal water would increase it. As an example, the thermal gradient in a 100-m-deep well above thermal water at 60°C at a depth of 200 m with a surface temperature of 20°C would be 40°C per 200 m, or 200°C/km. This gradient is impressive—until one realizes that the thermal gradient below the thermal water would decrease and probably even reverse as cold ground water is re-encountered. Such a setting could and probably does exist in many of the areas of hot springs in Trans-Pecos Texas. The thermal gradient reverses in Gulf-Swafford, where temperatures drop below the thermal-producing horizon and do not rise again except near the bottom of the well. Swanberg and Herrin (1976) illustrated a similar problem and its effect on heat flow measurements in the Big Bend National Park area. True thermal gradients can be determined only where convection does not occur.

## ACKNOWLEDGMENTS

The assistance of all who helped in this study is gratefully acknowledged. Many landowners generously allowed access to their property for visits to hot springs or for examination of important structural relationships. Discussions of geology, geochemistry, and hydrology with C. G. Groat, C. W. Kreitler, and A. W. Walton have been highly informative and have helped shape many ideas presented in this study. Chemical analyses were performed by L. C. McGonagle and K. C. Street under the supervision of D. A. Schofield at the Mineral Studies Laboratory of the Bureau of Economic Geology.

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Generally preferred sites for thermal-gradient and heat-flow measurements are in unfractured, therefore impermeable, crystalline rocks. Faulting and associated fracturing are extensive in Trans-Pecos Texas, but adequate sites should exist in mountain blocks bordering the Rio Grande. Some of the best possible sites for measuring heat flow and also for discovering areas of anomalously high but deep thermal gradients are basin centers. The unfaulted, fine-grained sediments should be highly impermeable and preclude convection. Blockage of heat flow by low-conductivity sediments could produce abnormal thermal gradients and high temperatures at relatively shallow depths below the sediments (Diment and others, 1975; Hose and Taylor, 1974). Distinction between this kind of high thermal gradient and that caused by convection may be difficult, yet examination of the permeability of the material penetrated by a well should provide some differentiation. In general, heat-flow studies in both the basins and crystalline rocks in the adjacent highlands should determine regional heat flow and provide some evidence concerning the area's true geothermal potential.

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