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

Final Report

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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^{o} and $72^{o}C$ for hot springs and wells in Texas and $90^{o}C$ 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 as the youngest observed igneous activity in West Texas is Miocene based on both paleontologic and isotopic evidence.

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 the rise of thermal water from below. Recharge for the thermal systems probably occurs in adjacent highlands. Hot springs are not restricted to faults of large displacement or to particular rock types. However, many faults 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 thermal circulation could occur along other faults but does not discharge to the surface. For example, several wells tap hot water at depths of about 20 to 1000 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_3, and SO_4. Waters from zeolitized, silicic volcanic rocks contain low to moderate total dissolved solids composed primarily of Na and HCO_3. Evaporite waters have high total dissolved solids, Na, Cl, SO_4, HCO_3, 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 negates use of the sodium-potassium-calcium method. Solution of amorphous silica and nonequilibrium with quartz complicates use of the silica method.

The best interpretation of geothermometry indicates that there is an intermediate temperature group of springs with maximum subsurface temperatures around 60°C approximately equal to their surface temperatures. Three thermal systems, the two Gulf wells in Texas and Ojos Calientes in Mexico, have higher temperatures, at least 100°C and probably up to 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 greatest 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. The 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 the Presidio and Hueco Bolsons. Although it has many hot springs, the Big Bend area has low subsurface temperatures and no evidence of recent faulting. Hot springs in this area 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, has recent fault scarps and high silica content in the ground water. However, there are no hot springs or wells, and the high silica content probably reflects shallow circulation through volcanic and volcaniclastic rocks containing amorphous silica. The lack of thermal water suggests normal heat flow.

The 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 sources of energy proposed to supplement our dwindling oil and gas supply. In recent years, geothermal exploration has been intense throughout much of the western United States. Trans-Pecos Texas is an area for potential development of geothermal energy although information necessary to evaluate its 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 favorable geothermal areas in the western United States.

With this sketchy but suggestive background, a preliminary evaluation of the geothermal potential in Trans-Pecos Texas was warranted. Within the Trans-Pecos region, the study focuses on the Rio Grande Valley, which has almost all of the known hot springs or wells and has the greatest geothermal potential according to this study. However, other parts of Trans-Pecos Texas are also evaluated with the available data.

This report summarizes existing information and presents the results of a 1-year intensive study of the area. The study proceeded through several overlapping phases: (1) compilation of existing geologic information, both regional studies of geology, structure and geophysics, and more detailed local studies of individual hot spring areas; (2) detailed geologic mapping of hot spring areas to understand the origin and geologic controls of hot 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 had been made directly dealing with geothermal energy. Nevertheless, basic geologic information is necessary for understanding the origin of thermal waters. Applicable studies are summarized in the sections on regional geologic setting, source of heat, and geologic setting of hot springs. The published work was supplemented by detailed examination and mapping of hot springs and significant structural areas and is 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 understand the geochemical history of the thermal waters. This information is discussed in the sections on geochemistry and geothermometry. All 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 them 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 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 5700 m (19,000 ft) of folded, regionally metamorphosed arkose, quartzite, schist, limestone, and rhyolite (King and Flawn, 1953). Deformation and metamorphism occurred about 1250 million years ago (Denison and Heatherington, 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 (2500 ft) of limestone and dolomite with minor volcanic rocks. The Hazel Formation consists of 1500 m (5000 ft) of conglom-

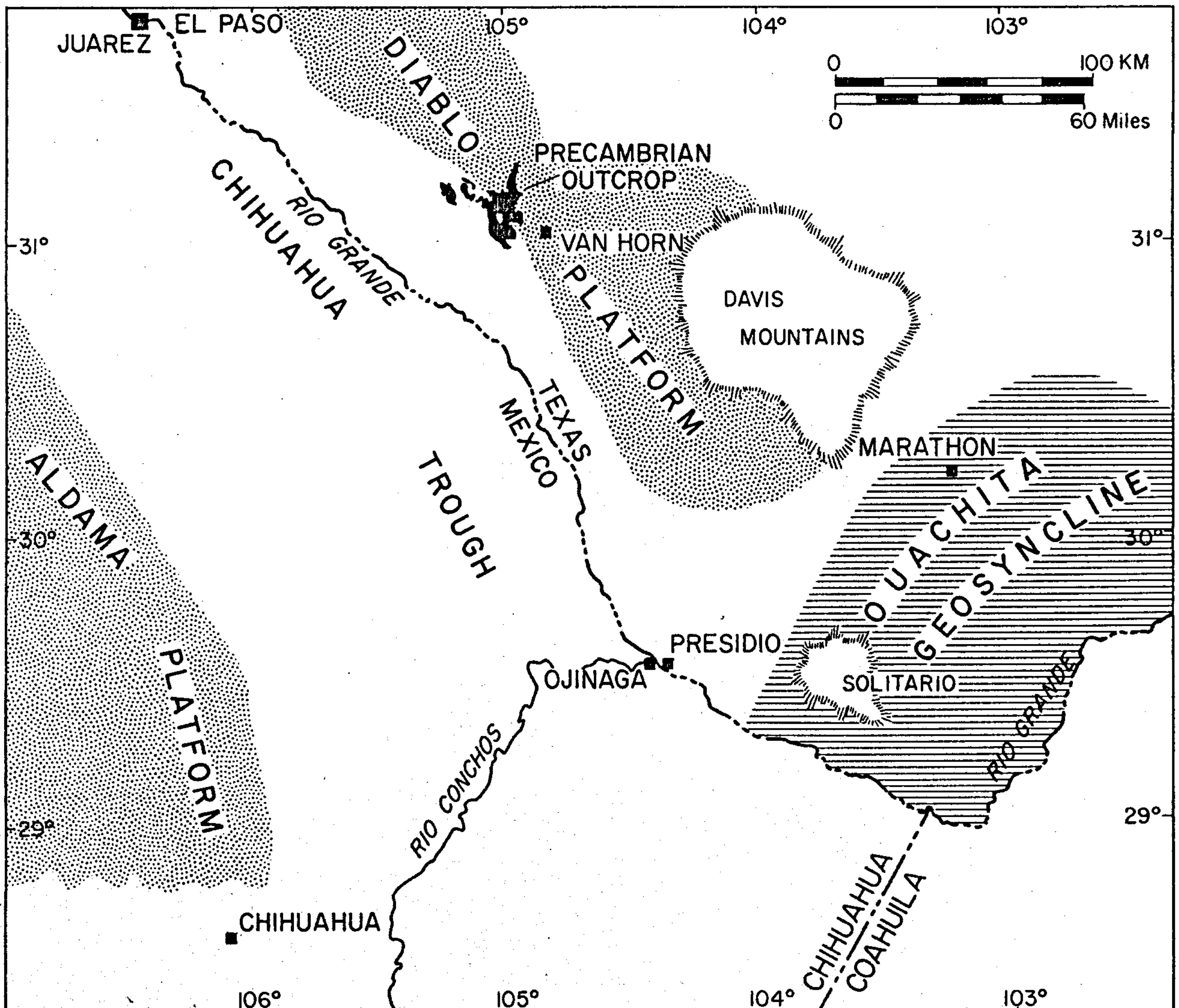

Figure 1. Major structural elements of Trans-Pecos Texas and adjacent Mexico.

erate and sandstone (King, 1965). Both were complexly deformed but only mildly metamorphosed about 1000 million years ago (Denison and Heatherington, 1969). All three formations are unconformably overlain by the Van Horn Sandstone 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 4500 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 and shales are sandstones grading upwards 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 that were the sources for coarser Upper Pennsylvanian strata, and ended in Late Pennsylvanian or Early Permian time (Flawn, 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, the Paleozoic rocks are completely covered beneath thick Cretaceous and Tertiary sequences southwest of the Solitario.

On the Diablo platform to the north (fig. 1), equivalent Paleozoic rocks are only 750 m (2500 ft) thick and are mostly shelf carbonates. The rocks were uplifted, gently folded, and faulted in Pennsylvanian time, probably contemporaneously with intense folding in the Marathon region.

Deposition over much of Trans-Pecos Texas resumed in 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, 1959; Deford, 1969; Wilson, 1971). 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 in the Cretaceous. Before that, possibly as early as Permian but probably Jurassic (Haenggi, 1966), evaporite deposits of halite and gypsum accumulated in much of the trough. Total thickness of evaporites is unknown as they are deformed everywhere observed. Following evaporitic deposition, up to 5500 m (18,000 ft) of Cretaceous limestone and shale was deposited in the trough. 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 4300 m (14,000 ft) of Lower Cretaceous rocks (Jones and Reaser, 1970). In the Eagle Mountains, Cretaceous sequences average about 2100 m (7200 ft) thick (Underwood, 1962), and on the Diablo platform there are only 600 m (2000 ft) of equivalent rocks. In all three areas Upper Cretaceous rocks have primarily been removed by erosion. Amsbury (1958) reported approximately 850 m (2800 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 3650 m (12,000 ft) thick (Gries and Haenggi, 1971). Individual formations within the Cretaceous show similar changes in thickness (Deford and Haenggi , 1971).

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 dominantly

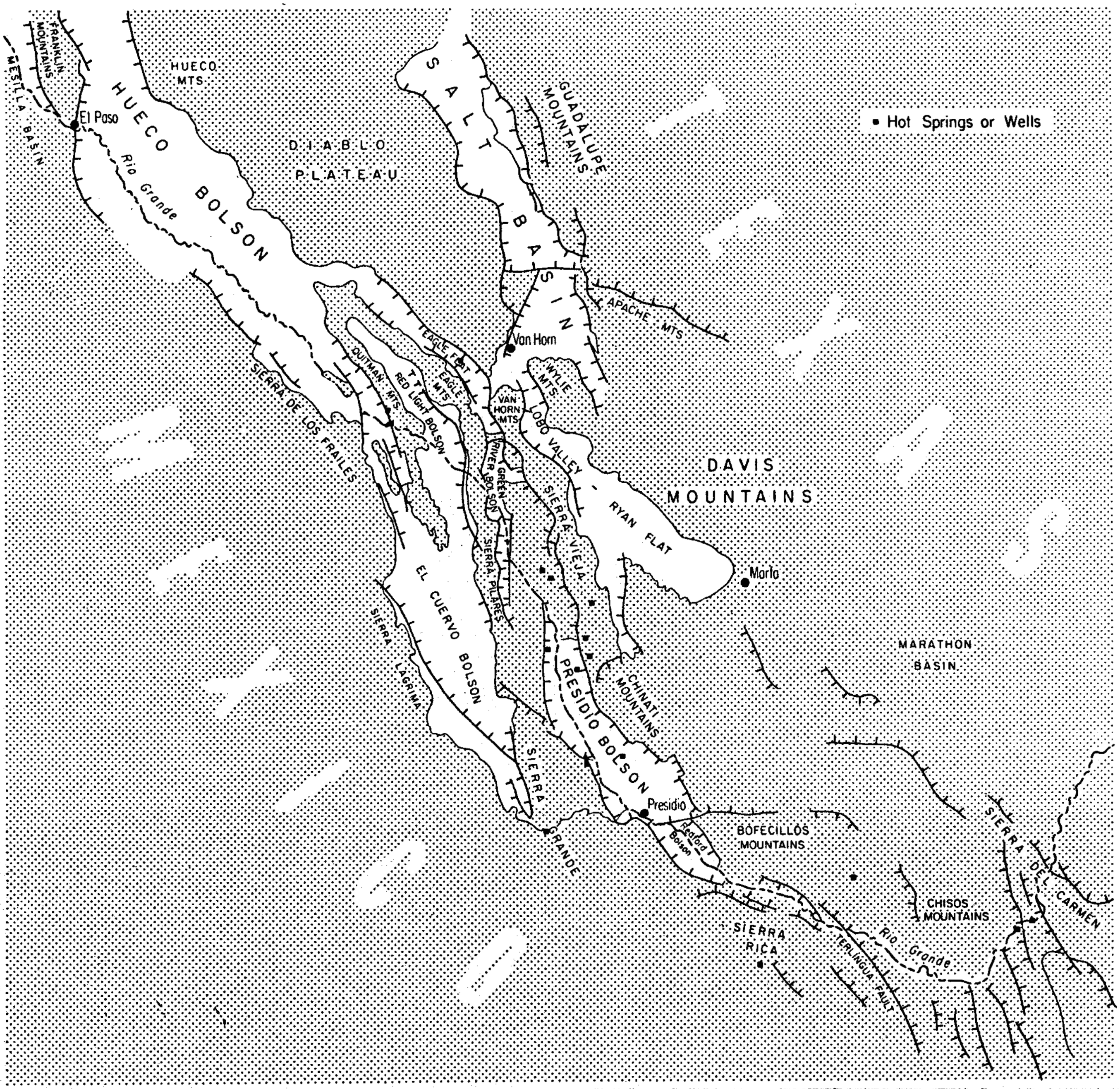

Figure 2. Basins and ranges and major normal faults, Trans-Pecos Texas and adjacent Mexico.

clastic, containing thin limestone, shale, clay, and sandstone beds 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 (1971) 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 was 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.

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 throughout the Oligocene and locally into the Miocene. Most of the volcanic rocks are silicic and alkalic (Barker, 1977) with peralkaline rocks dominating along an eastern trend from the Chisos Mountains through the Davis Mountains, and a metaluminous trend dominating from the Bofecillos Mountains through the Chinati Mountains to the Sierra Vieja and the Van Horn

Mountains (fig. 2). Most of the youngest volcanics are more basic, although still alkalic. The latest igneous activity is Miocene and 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 middle Tertiary, volcanism was rare in Mexico immediately west of Trans-Pecos Texas. South of the Bofecillos Mountains in the Sierra Rica (fig. 2) is a thick sequence of tuffs and lavas evidently derived from a major caldera center. 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, 1968; Gries, 1970).

Although faulting had occurred earlier, major Basin and Range style block faulting began in late Oligocene - early Miocene time (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 were accumulating in them at least by 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 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).

Older structures apparently controlled some of the faults. Besides the general parallelism of 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 1200 m (4000 ft) occurred on the Rim Rock, 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 the Hueco Bolson east of the Franklin Mountains, the 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 2750 m (9000 ft) of sedimentary fill in the Hueco Bolson east of El Paso (Gates, 1976), but Presidio Bolson and Lobo Valley probably do not contain more than 900 to 1370 m (3000 to 4500 ft). The Salt Basin (fig. 2) is relatively shallow with a maximum of 750 m (2500 ft) of fill, but generally 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 presently 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. Although it is structurally low, the Big Bend area has bedrock outcrops throughout most of it. 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 the 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 sides of the Salt Basin and Lobo Valley (Belcher and Goetz, 1977), along the west side of the Eagle Mountains (Underwood, 1962) and along both sides of Presidio Bolson (Gries, 1970 and this study). Scarps are not evident bordering the Quitman Mountains, but Chan and others (1977) identified recent epicenters in that 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.

Extension of the Rio Grande rift 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 may be differentiated by gravity and heat-flow data (Decker and Smithson, 1977). Bedrock relief across basin margins in the rift ranges up to 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 is probably not much more than 2000 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 none to "secondary faults parallel to but not necessarily directly overlying the basement faults" (Gries, 1977). Heat-flow evidence related to the question is discussed in the following section.

SOURCE OF HEAT TO THE THERMAL SYSTEM

Three sources of heat for hot springs which have been proposed for other thermal systems, are (1) shallow (less than 6 km), young (but not necessarily molten) magma chambers; (2) deep circulation of ground water in a convective system with heat supplied by an area's normal thermal gradient; and (3) blockage of normal heat flow by rock layers of low thermal conductivity.

Each of these sources of heat requires "deep" circulation. The major difference between them is in ultimate source of heat--magmatic, normal thermal gradient, or enhanced thermal gradient. The word "deep" in this study does not have a precise quantitative meaning. Maximum depths of circulation of ground water, even in a tectonically active area, are probably under 10 km. 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 on the order of several hundred meters to several kilometers. More precise estimates of depths of circulation are given in another section.

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, the only commercial geothermal-power-producing facility in the United States. Magmas are unlikely sources of heat for the Texas thermal systems, however, 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 Potrillos 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 previously listed are rhyolites or dacites. Finally, because the hot springs in Texas are widespread, each group would require a separate magma. More evidence should exist for magmas than just hot springs.

Deep ground-water circulation of approximately 1 km or more is a much better source of heat. Meteoric ground water has to circulate relatively deeply to be heated even with a magmatic source. Association of hot springs with normal faults shows that conduits do exist for deep circulation. What is not known is the depth necessary to produce either the measured surface temperatures or subsurface temperatures inferred from geothermometry. The setting of Trans-Pecos Texas in the Basin and Range province or possibly even the Rio Grande rift implies that heat flow and thermal gradients may be considerably greater than continental averages. However, an abnormal gradient is not required because sufficiently deep circulation in an area of normal heat flow can produce hot springs (for example, Hot Springs, Arkansas, Behringer 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. Theoretical 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 conductivity. Poorly consolidated, water-saturated sediments, such as those that fill the late Tertiary basins of West Texas, exhibit low conductivity. Some volcanic rocks with high porosities such as ash-flow tuffs also have low conductivity. In the Basin and Range 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 is similar. Basins in Texas could have similar properties and provide a source of heat for some of the hot springs.

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 total minus a component from radioactive heat generation in the upper crust. Unless otherwise noted, heat flow presented here is unreduced. Reduced, or mantle and lower crustal heat flow, is about 0.8 heat flow unit (HFU) (1 HFU = 1 x 10^6 cal/cm^2 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 (Ray 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.

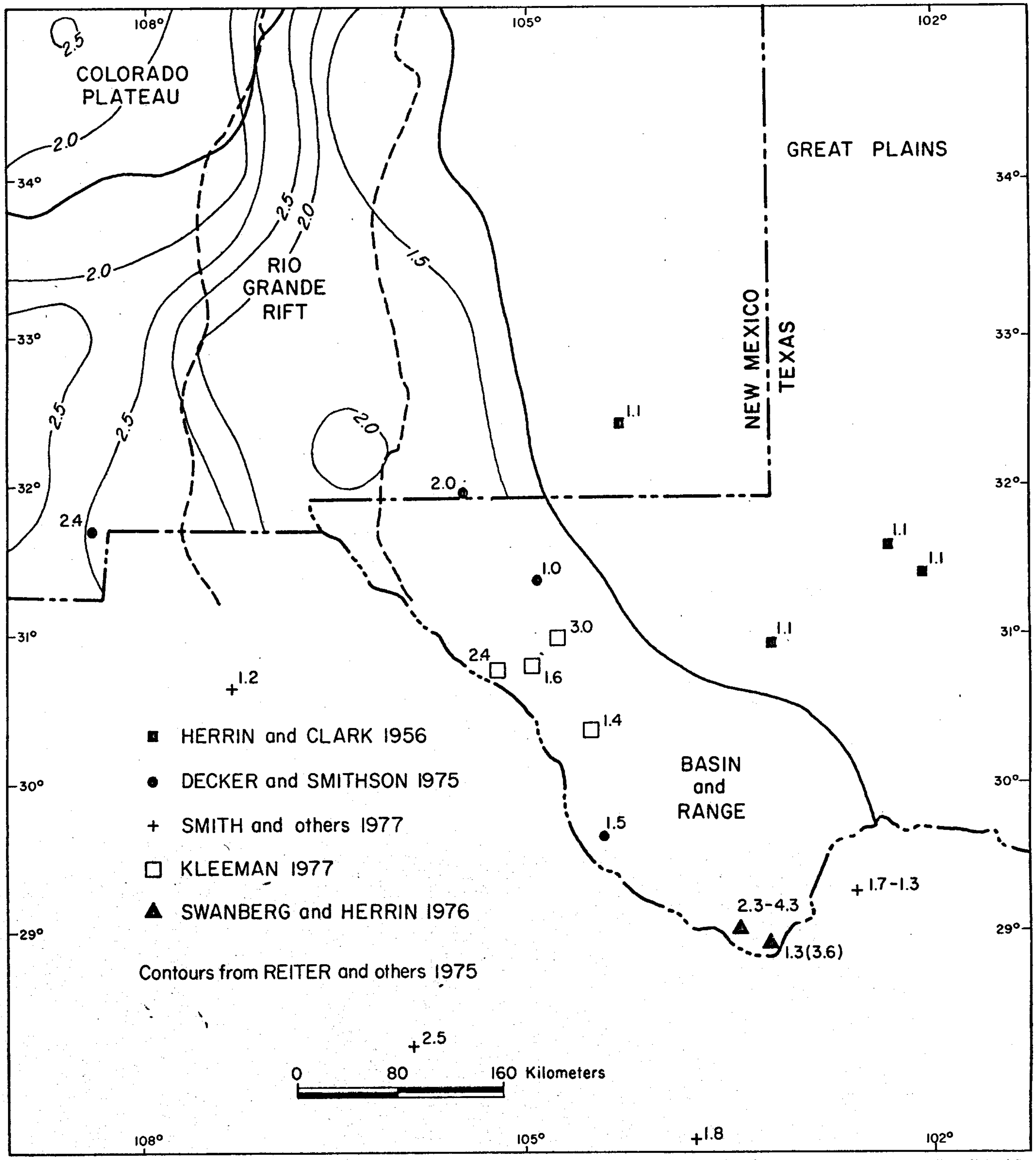

Figure 3. Heat-flow determinations in Trans-Pecos Texas and adjacent regions. Physiographic provinces from Fenneman (1946).

Smith (1977) presented several heat-flow values in Chihuahua and Coahuila, but these values unfortunately do not aid in clarifying 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 more southeastern trend from New Mexico at least parallel to (but not necessarily coincident with) the Rio Grande.

Decker and Smithson (1975) determined two values in Texas (fig. 3). One near Van Horn in the Basin and Range physiographic province is 1 HFU below the Basin and Range average; the well has a thermal gradient of only $14^{o}C/km$. A second location near Shafter just east of Presidio Bolson has a value of 1.5 HFU (1.2 HFU reduced). This value is much lower than the Rio Grande rift values are, is below even 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: with 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 $25^{o}C/km$.

Kleeman (1977) assumed conductivities for well data reported by Gates and White (1976) to give heat-flow values of 1.4 to 3.0. Thermal gradients for the wells range from 27^{o} to $50^{o}C/km$. All of these wells are water tests 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.

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 it 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, then 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 American Association of Petroleum Geologists (AAPG) thermal gradient map (AAPG, 1975) displays gradients in degrees Farenheit per 100 ft based on deep oil- and gas-well-temperature measurements. Unfortunately, there are few oil tests in Trans-Pecos Texas and almost none with thermal-gradient measurements in the Rio Grande valley. What the map does show is a general increase from the midcontinent part of West Texas with values around 15-18^{o}C/km (0.8-1^{o}F/100 ft) to Trans-Pecos Texas with values around 18-26^{o}C/km (1-1.4^{o}F/100 ft). Collins (1925) listed a normal thermal gradient for the Trans-Pecos area at about 29^{o}C/km (1.6^{o}F/100 ft). None of these gradients is particularly promising for geothermal energy. They do at least suggest that the Trans-Pecos region is distinct from the Great Plains and that thermal gradients increase towards 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 (for Basin and Range) in central Nevada has fewer and cooler hot springs than surrounding areas. With the exception of the springs in and around Big Bend Park, all the hot springs in the area are along the nearly linear southeast-trending section of the Rio Grande from El Paso to Presidio. Presence of the hot springs there and the high temperatures for some of them indicate that it is an area of high heat flow--even higher than that of adjacent areas to the east.

These lines of evidence indicate that Trans-Pecos Texas is a region of transition 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. The Rio Grande region of Texas probably has Basin and Range heat flow. However, the presence in Texas of extremely high heat flow, such 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.

The source of a thermal anomaly expressed by high heat flow in the Basin and Range and Rio Grande rift is generally attributed to 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 in the Basin and Range are around 20 to 30 km (Healy and Warren, 1969), similar for the rift, and about 50 km under the Great Plains (Pakiser, 1963).

An alternative explanation of the high heat flow calls for 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 alternating areas of high and low heat flow. Present data suggest a more continuous source, but refinement of the data might reveal a more complicated picture.

GEOLOGIC AND HYDROLOGIC SETTING OF HOT SPRINGS AND WELLS

Introduction

Chemical (White, 1957) and stable-isotope studies (Craig, 1963) have provided conclusive evidence that hot spring waters are derived almost entirely from local meteoric water. Preliminary stable-isotope analyses of thermal and nonthermal waters in this study demonstrate the meteoric origin of hot springs waters in Trans-Pecos Texas. The previous chapter demonstrated that the source of heat for those thermal waters is deep circulation 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. The geologic setting of hot springs in Texas shows how this occurs.

The summary by White (1968) of the geometry and hydrodynamics of hot-spring systems is useful in understanding the origin and flow paths of springs in Trans-Pecos Texas. Hot springs have two driving forces: one is the difference in elevation between areas of recharge and areas of discharge and acts on all springs; the second is the difference in density between cold recharge water and hot discharge water. As pointed out 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 hot spring. 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), 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 (85°F). This definition excludes springs which discharge waters of thermal origin but which have cooled below this temperature because of mixing with nonthermal ground water or to conductive cooling during slow discharge. Such springs could be recognized by other criteria, for example, 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 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 unique 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 this region: the southern Hueco Bolson adjacent to the Quitman Mountains; the 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 occur in all three areas and on Eagle Flat (fig. 2).

Southern Hueco Bolson

Indian Hot Springs

Hot springs in the 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 Reaser (1974). There are at least seven springs ranging in temperature from 27^{o}C (81^{o}F) at Soda Spring to 47^{o}C (117^{o}F) at Stump Spring when measured in September 1976. In 1968, Jones reported temperatures that range up to 52^{o}C (126^{o}F). Dorfman and Kehle (1974) reported an unconfirmed temperature of greater than 60^{o}C (140^{o}F) for a shallow well. When visited in 1976, Dynamite Spring and Masins Spring were not active. Dynamite Spring at present is only a shallow well tapping Rio Grande alluvial water with 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 about 0.5 m above the plateau (fig. 5). Chief and Squaw have rock and wood "bathhouses" built around them, whereas the others have rock "tubs." Only Stump discharges to the surface; the others discharge through permeable travertine or alluvium below the travertine to flow to the Rio Grande. Total discharge is at least 400 liters per minute (l/min) and probably considerably greater but difficult to estimate precisely.

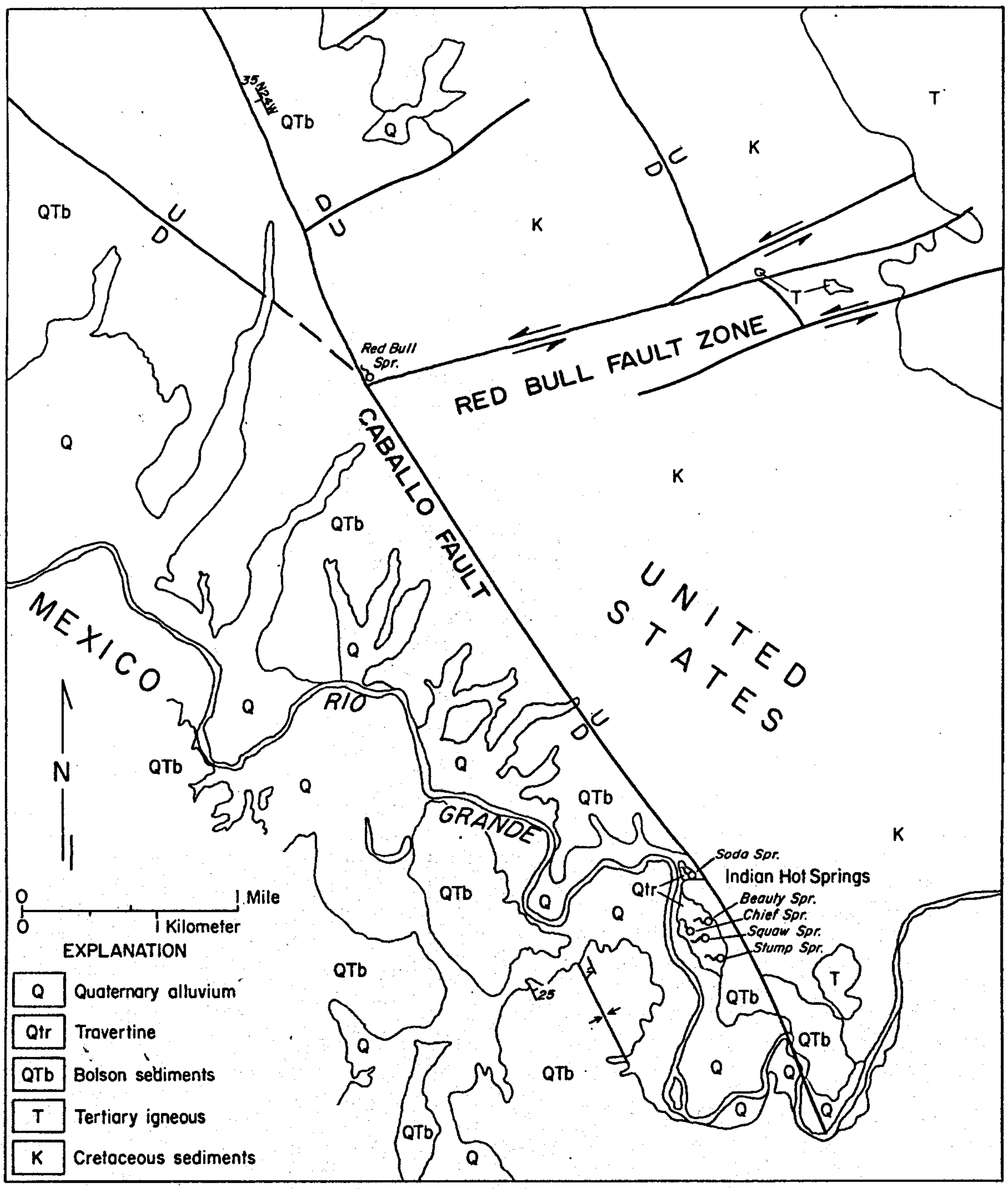

Figure 4. Geologic map of Indian Hot Springs and Red Bull Springs area, southern Hueco Bolson, Texas. Geology from Jones and Reaser (1970) and this report.

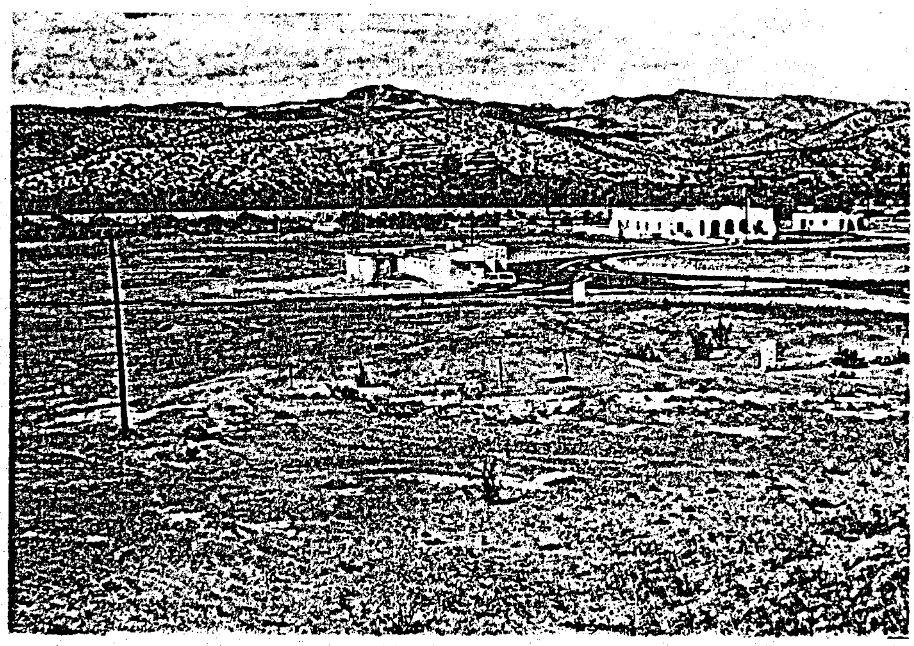

Figure 5. View to southwest of Indian Hot Springs. Flat area in middle distance is underlain by travertine deposited by spring waters. Chief Spring (center, inside bathhouse on left) has built up a low travertine mound above plateau. Tilted bolson sediments form cliff face on far side of Rio Grande in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

Bell (1963) and Jones (1968) suggested that most of the springs are really 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^{o}C (81^{o}F) below the required temperature for a thermal spring. Soda Spring has built up an extensive travertine deposit, 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 occurs just basinward of the Caballo fault, a major northwest-trending normal fault dividing the Quitman Mountains and the Hueco Bolson. The fault separates Cretaceous sediments on the northeast from down-dropped bolson fill on the southwest (figs. 4, 6). The fault trace follows a narrow bench along the range front 6 m (20 ft) above the travertine plateau (fig. 6). Road-cut exposures of bolson sediments along this bench are calcite cemented. A cap of travertine that overlies the sediments may be from earlier spring activity at the higher elevation of the bench. Subsequently, down-cutting of the Rio Grande lowered the hydrologic base level; the thermal waters now emanate from a lower level. Continuation of spring activity during down-cutting of the Rio Grande suggests that the springs and thermal circulation are as old as some of the terraces.

Movement on the Caballo fault started in early Miocene (Jones and Reaser, 1970); it may still be active. Bolson sediments are brecciated and strongly tilted from drag along the trace of the fault. Jones and Reaser (1970) show

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.

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. Dorman (1977) reported a magnitude 2 to 3 earthquake epicenter within about 20 km of Indian Hot Springs. Uncertainty in the precise epicenter location does not permit correlation with any known fault, but the earthquake demonstrates that faults in the area are still active.

Red Bull Spring

Red Bull Spring (fig. 4) discharges 37°C water at 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 the 1931 Valentine earthquake the discharge of Red Bull Spring was considerably greater than it is at present.

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. 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. The fault is not as evident on the ground. An indistinct scarp with the southwest side downdropped 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.

Presidio Graben

Much of the present thermal activity in Trans-Pecos Texas and Mexico occurs within the structural low that includes the 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 the Presidio Bolson (fig. 7,8) in an area mapped by Buongiorno and others (1955) and Twiss (unpublished). The springs occur 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 37°C (98°F) from numerous fractures in a rhyolite lava flow on the down-dropped side of the fault (fig. 9). Total discharge is approximately 400 l/min according to estimation of stream flow below the springs. There are no spring deposits.

The fault is one of several normal faults which trend approximately north and dip steeply to the west. Displacement is about 50 m down to the west.

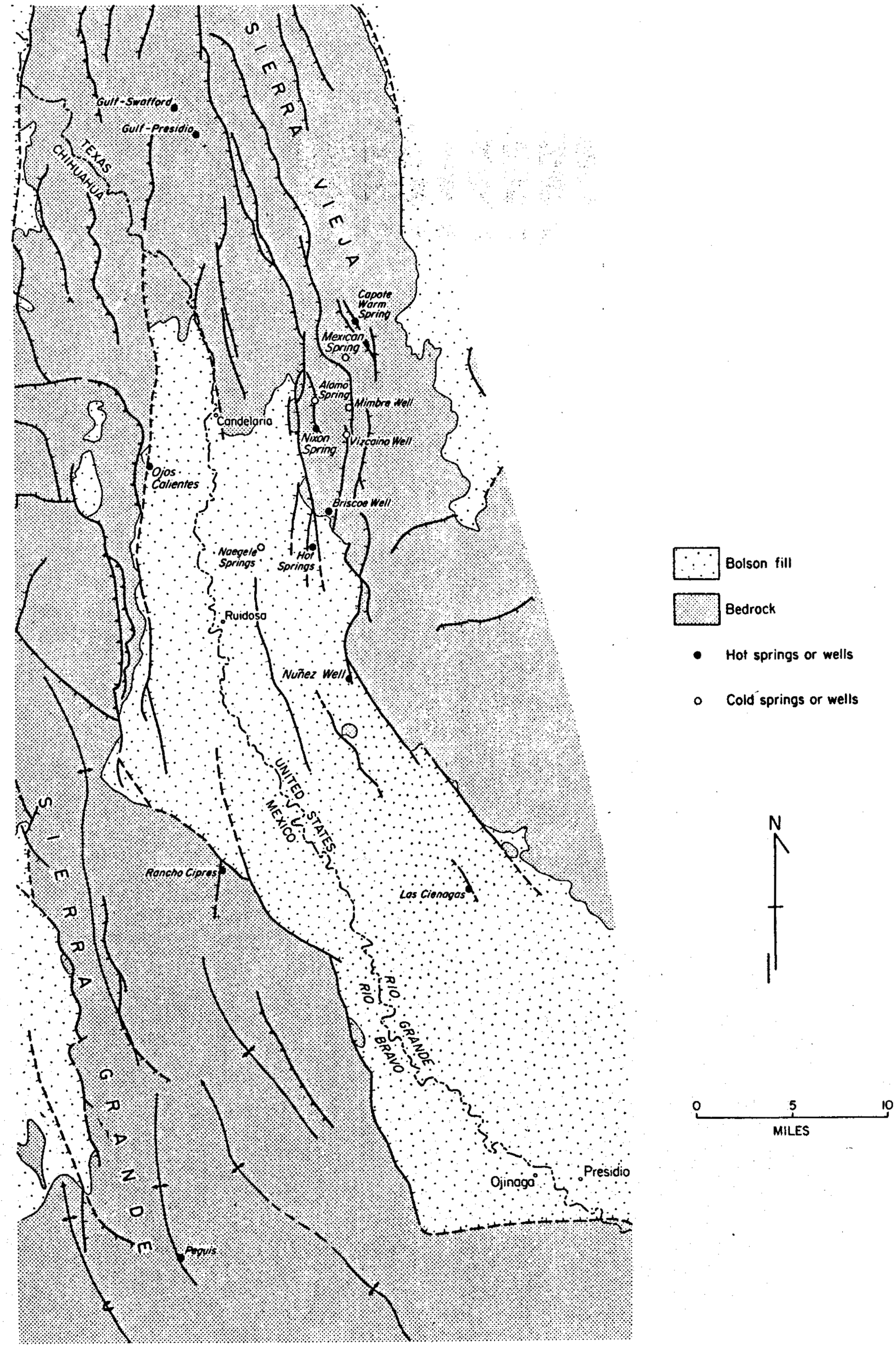

Figure 7. Presidio Bolson and location of sampled hot and cold springs and wells.

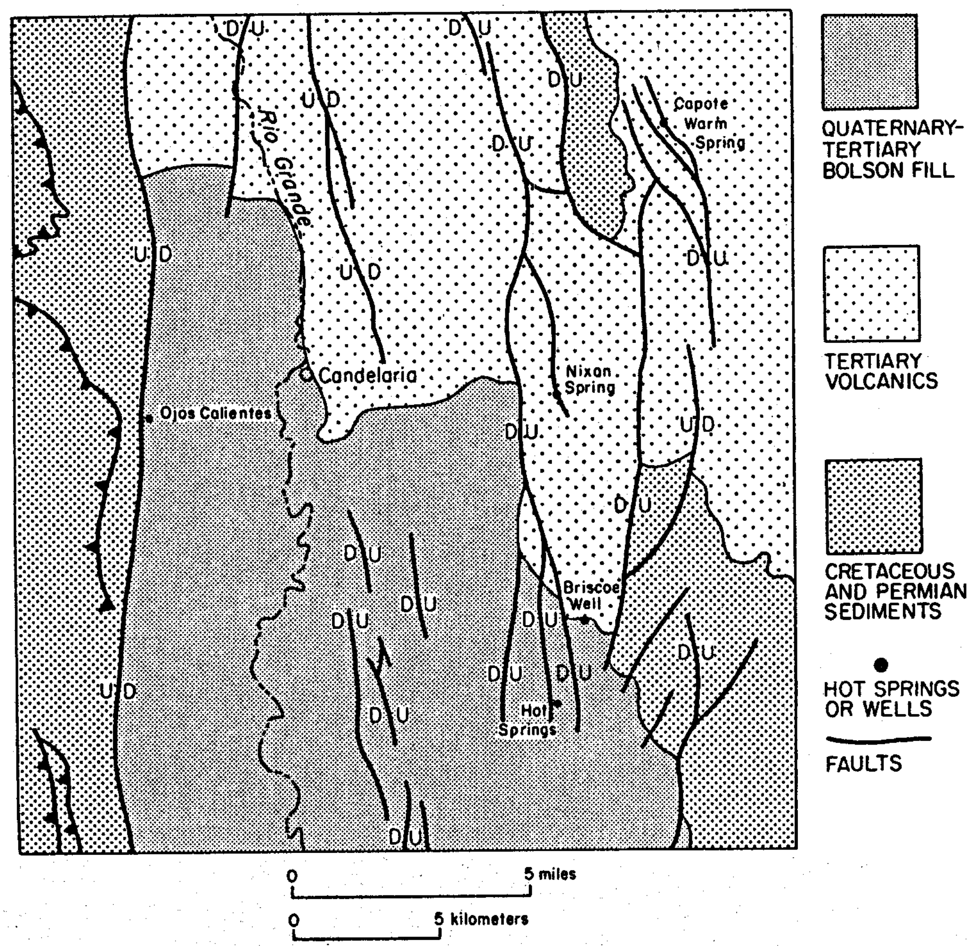

Figure 8. Geologic map of northern Presidio Bolson showing thermal springs and wells.

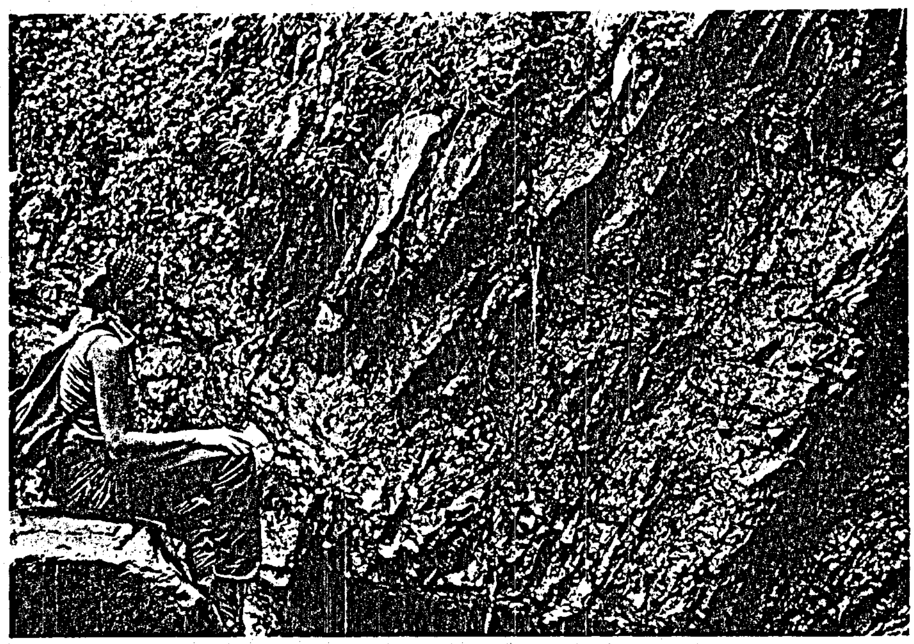

Figure 9. A part of Capote Springs emerging from fractures in rhyolite lava flow along a minor normal fault.

Movement on this fault system started in late Oligocene or early Miocene (Dasch and others, 1969). Evidence of more recent movement is not available as the fault is entirely within Tertiary volcanic rocks.

Nixon Spring

Nixon Spring occurs on a minor branch of the Candelaria fault in an area mapped by Buongiorno and others (1955) and Twiss (unpublished) approximately 8 km southwest of Capote Springs (figs. 7,8). 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 occur 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 the Presidio Bolson (figs. 7,8). 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 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.

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 on the ground or on aerial photographs south of Hot Springs Creek. 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 other 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 the 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 1000 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 an approximation. 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 which 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). 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 fault. 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 the sandstone facies of Groat (1972). Sediments at and below the springs are Groat's mudstone facies. Presence

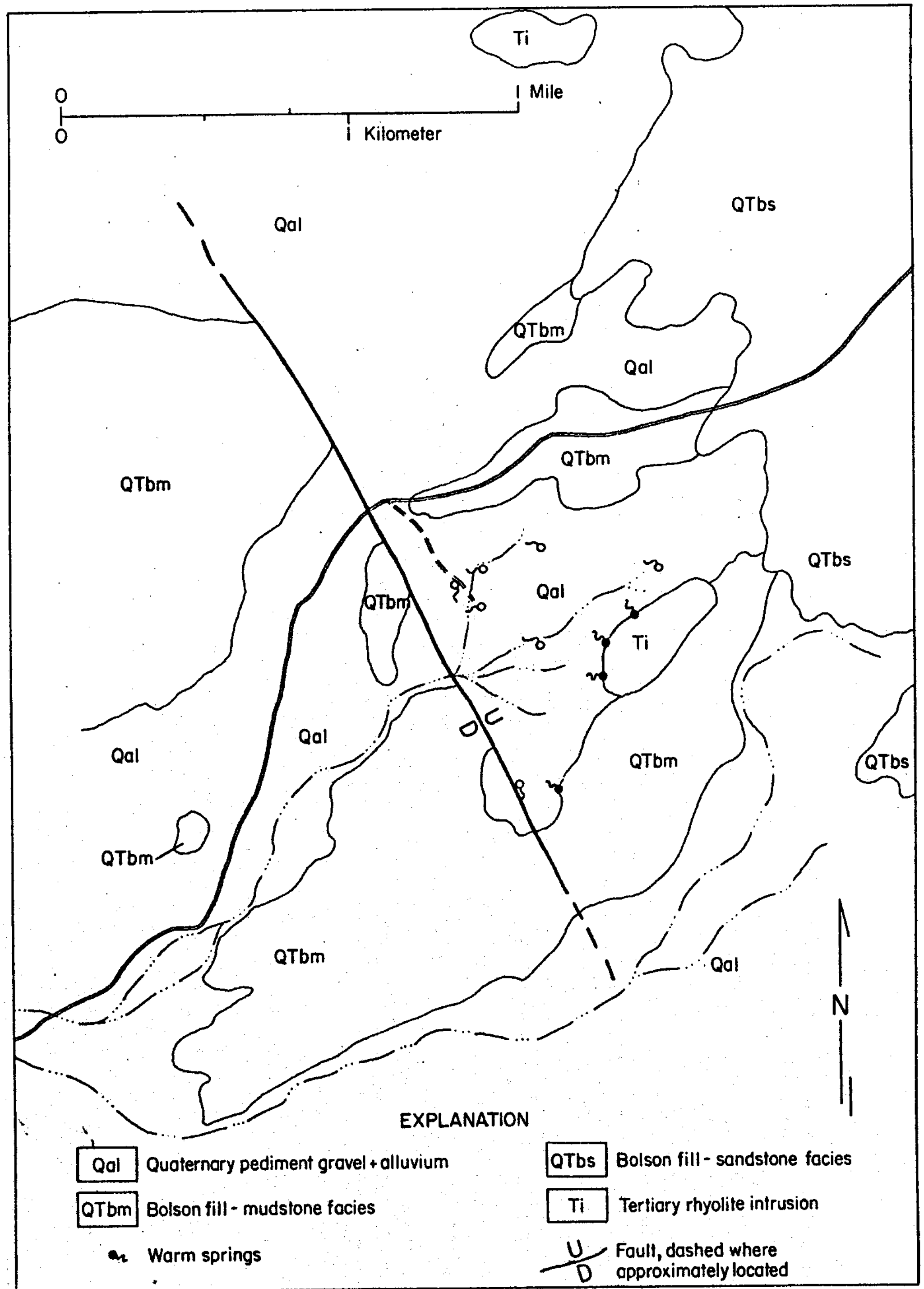

Figure 10. Geologic map of Las Cienagas area, Presidio Bolson, Texas. Geology from Groat (1972) and this report.

of impermeable clay accounts for the surface 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 only be identified as postbolson 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 cut youngest Quaternary terrace gravels with as much as 5 m of displacement down to the west.

Ojos Calientes

Ojos Calientes, the hottest thermal activity observed during this study, is located on the Mexican side of the Presidio Bolson approximately 7 km west southwest of Candelaria (figs. 7,8). 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 occur 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 temperatures range from approximately 60°C (140°F) to a maximum of 90°C (194°F). The lower temperatures may result from mixing with nonthermal ground water or shallow cooling of low-discharge springs. Several springs continuously spurt approximately 0.5 m into the air like miniature geysers. The driving force may be degassing of CO_2 dissolved in the thermal water rather than steam pressure. Total discharge of all the springs is estimated from flow of the arroyo below the springs to be more than 1000 l/min.

The springs have built extensive travertine deposits including many small knobs and several broad mounds as high as 3 m above the arroyo bottom (figs. 11, 12). Discharge occurs from the tops and sides of the mounds, and changed significantly in the period of a few months. During this same period, 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^{o}E., dips 55^{o} 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^{o} to the east. Within 100 m bolson beds return to the normal 3^{o} western dip. The fault zone is filled with fine-grained, massive calcite which could have been deposited by thermal water.

Haenggi (1966) estimated 900 m of displacement on the Palo Pegado fault in an area approximately 10 km north of the springs where Cretaceous sediments occur on both sides of the fault. Displacement increases towards the springs, thus 900 m is a minimum estimate of displacement in the springs vicinity. No other faults occur 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. However, to the south in an area mapped by Gries (1970), a similar boundary fault cuts Quaternary deposits.

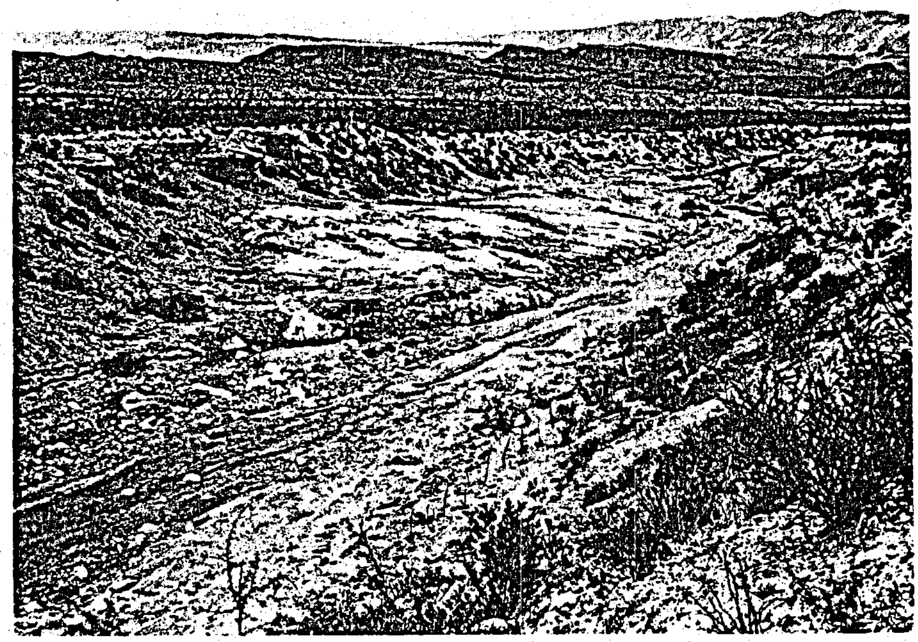

Figure 11. View to northeast across travertine mound deposited by spring waters at Ojos Calientes. Sediments in side of arroyo are Quaternary terrace gravels. In background are ridges of dark volcanic rocks and white volcaniclastic sediments of the Vieja Group in Texas.

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

Spring at Rancho Cipres

Another hot spring in Presidio Bolson in Mexico occurs near Rancho Cipres (fig. 7), approximately 40 km northwest of Presidio and 20 km south of Ruidosa in an area mapped by Heiken (1965) and Gries (1970). A single spring occurs near the top of a large, partially dissected travertine mound overlying Tertiary volcanic rock and Cretaceous sandstone and shale. The spring consists of 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), and there is a strong odor of H_2S. The pool bubbles constantly, probably from degassing of CO_2 and H_2S dissolved in the thermal water. Water does not flow out of the pool but percolates through permeable travertine and discharges around the edge of the mound as cool seeps. From the size of the travertine mound, approximately 700 m by 200 m in area and as thick as 10 m, 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

Figure 13. Spring at Rancho Cipres surrounded by travertine deposited by spring waters. Spring surface is coated with scum except where bubbling of CO_2 and H_2S has disrupted the surface.

hot water at a depth of 27 m (figs. 7,8). When the well was sampled in 1976, the temperature was 42°C (108°F); 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 close to the hot well produces only cold water. If the two water sources were hydraulically connected, the hot water should 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 towards the well location, but their relation to the well and thermal water are uncertain.

One other well approximately 4 km southeast of Hot Springs is reported to produce water at about 34°C (93°F) by 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; wells nearby are windmill driven and pump too slowly to measure a meaningful temperature.

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 the 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 only sampling conditions because the water cannot be sampled directly at the well head and air temperature at the time of this study was about 7°C (45°F). Temperature logs for Gulf-Swafford show a temperature inversion below the hot-water-producing horizon. Below that level, the temperature drops and reaches 80°C at a depth of about 2500 m (8200 ft). Thus, the high-temperature water of the shallow producing horizon must come from greater depths by thermal convection.

Both wells would flow to the surface if permeable channelways were available. That they do not reach the surface is somewhat surprising because the wells are in highly faulted terrain. Cretaceous formations above the producing horizon are largely impermeable shales which may prevent any flow.

Hot Springs at Peguis

Two hot springs 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 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 1000 l/min, and much additional water may flow into the river unobserved. 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

Table 1. Gulf wells: depth, temperature, and hydrologic data.[a]

	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)

[a]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

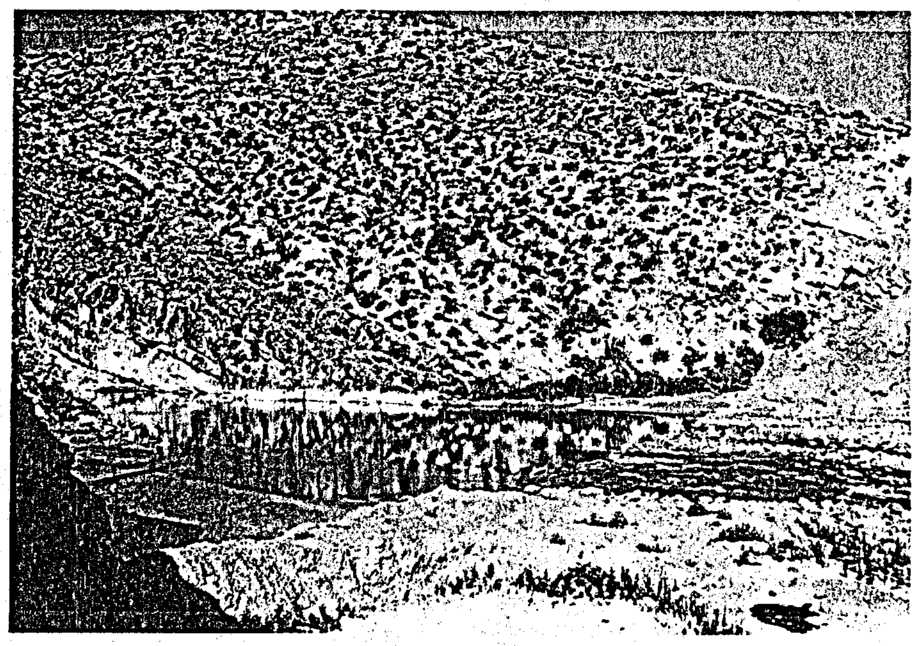

Figure 14. Hot springs at Peguis on east limb of anticline at Buda Limestone (massive limestone at left)--Ojinaga Formation contact (thin-bedded limestone and shale overlying Buda).

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, which is 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 in Mexico on the south side of the river (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 are probably representative of the others. The three sampled are Hot Springs (40°C, 104°F), a spring (41°C, 106°F) in cane approximately 300 m downstream from Hot Springs, and a spring (36°C, 97°F) used for water supply at Rio Grande Village (fig. 15). There are no spring deposits.

The spring at the village and all of the 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 through massive, cavernous

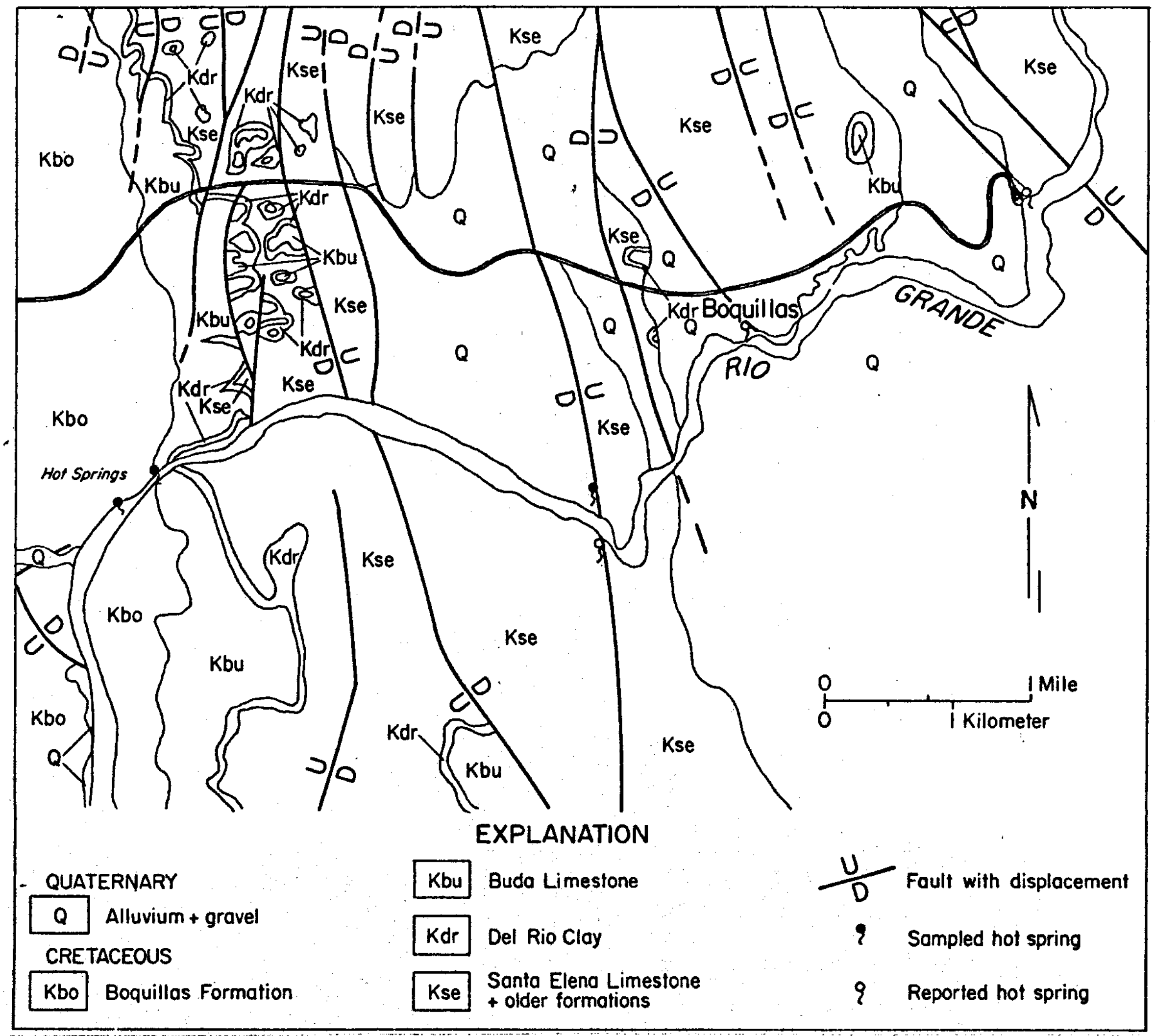

Figure 15. Geologic map of hot-spring area of Big Bend National Park and adjacent Mexico. Geology in Texas from Maxwell and others (1967) and in Mexico from Smith (1970).

(therefore permeable) Buda and Santa Elena Limestone but not through the thin-bedded Boquillas Limestone where it is not fractured. Thus, hot water discharges either at the contact or through fractures only slightly above the contact in a similar setting to the springs at Peguis. The Del Rio Clay, which should be impermeable, is very thin in the area 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 examined the whole park area more thoroughly.

Several warm springs have been reported along the Rio Grande downstream from the park. These springs are accessible only by a multiple-day boat trip and were not visited. 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

Abandoned mercury mines in the Terlingua mining district were reported by local inhabitants to be flooded with 45°C (113°F) water at a depth of approximately 275 m (900 ft). In addition, there is at least one deep (265 m, 875 ft) well that taps hot water (43°C, 110°F) at a similar depth. These reports could not be verified because the well presently has no pump and the mines are collapsing and 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 air is forming a convection system 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 the 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. The springs are all in the upper plate of a high-angle reverse fault, which thrusts Lower Cretaceous Limestone (table 2), over Upper Cretaceous rocks (figs. 17, 18). The Loma Plata is massive, cavernous, and highly fractured. Although fractured near faults, shales in the Ojinaga Formation are otherwise impermeable. Thus, the springs discharge at the contact of permeable and impermeable units. Cretaceous strata are complexly folded, similar in many respects to Laramide folds to the northwest. Laramide folding was undoubtedly a factor, but much of the deformation evidently occurred contemporaneously with caldera development, possibly during resurgence. Volcanic units nearby 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

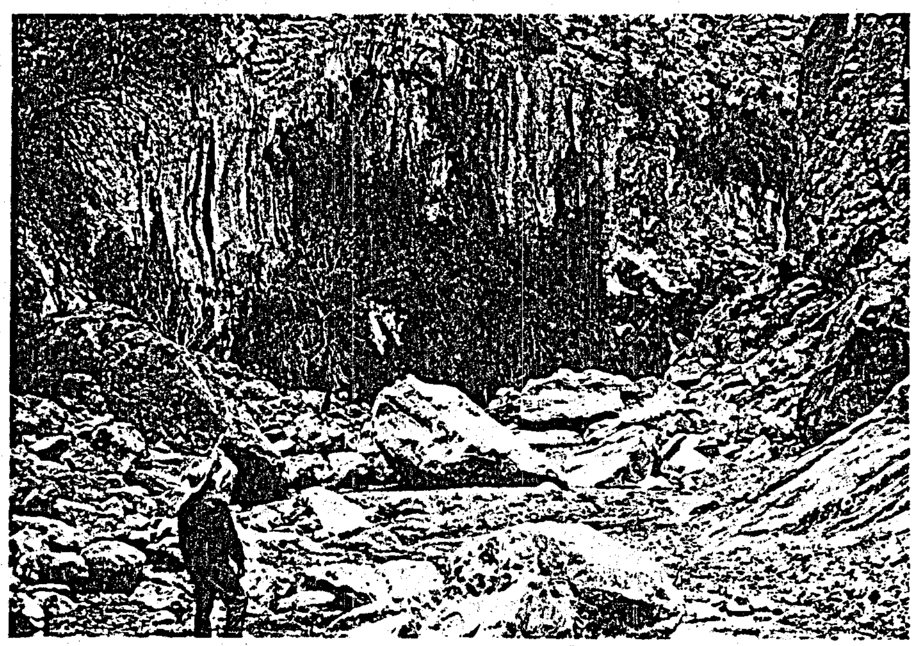

Figure 16. Travertine deposited by warm spring waters coats Loma Plata Limestone on far wall of canyon of Rio San Carlos.

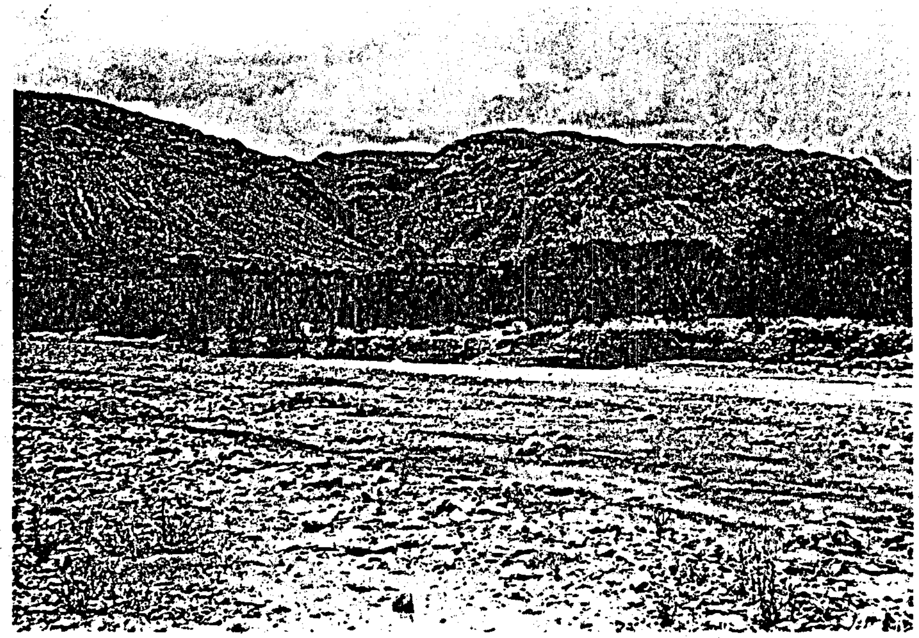

Figure 17. View to west of high-angle reverse fault with overturned Loma Plata Limestone thrust over Upper Cretaceous shales which crop out in Rio San Carlos at mouth of canyon. Warm springs discharge into canyon upstream from fault.

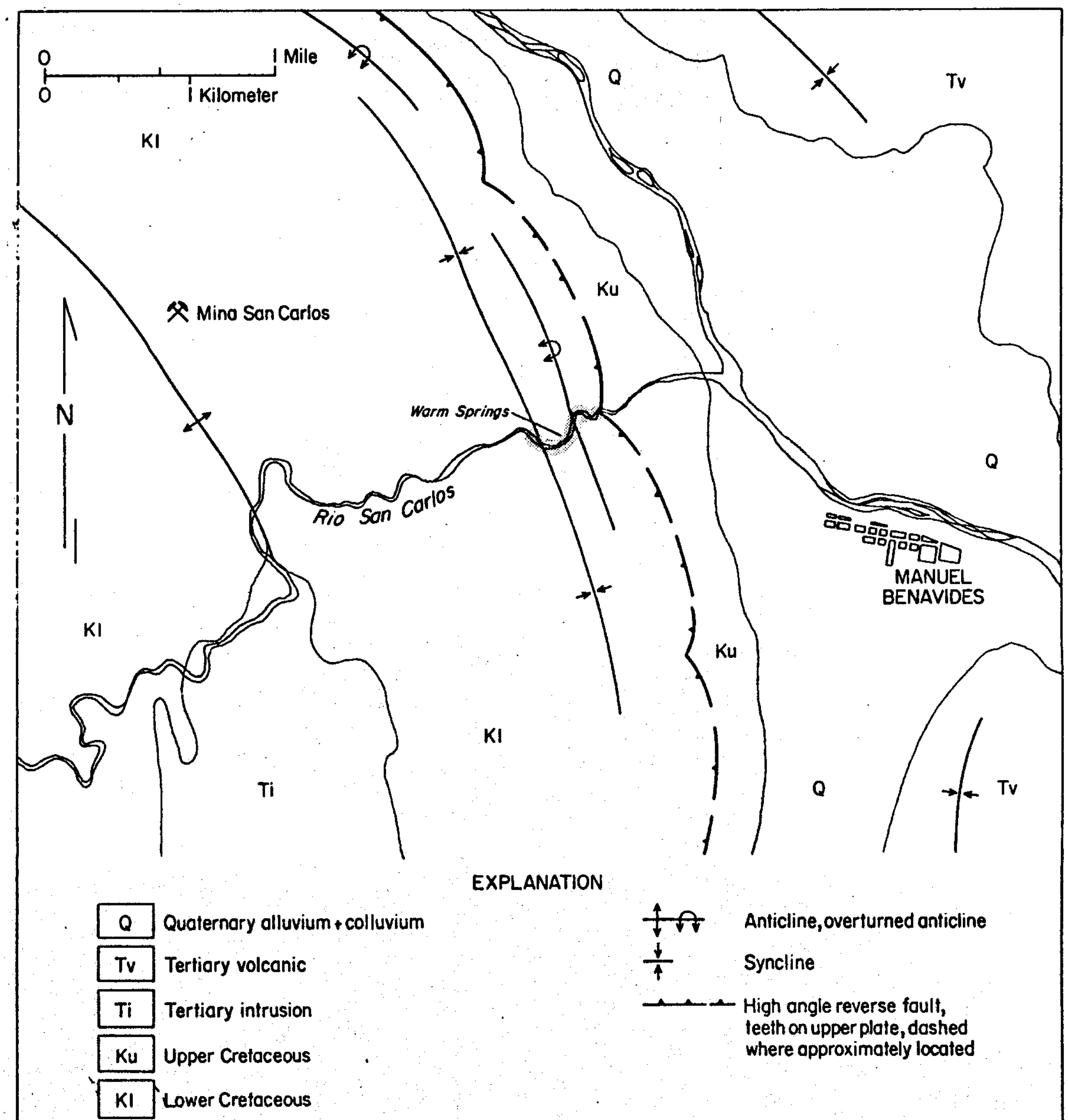

Figure 18. Geologic map of San Carlos warm spring area, Chihuahua, Mexico. Geology from Chacon (1972), Hernandez-Rios (1974), and this report.

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 circulation.

Sotano de Sauz

An unusual hot cave, Sotano de Sauz occurs 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 the description of Sprouse and knowledge of the geology of the area, the cave must be in the Loma Plata Limestone and possibly 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^{o}C (106^{o}F) at a depth of 220 m, maximum depth of the cave. The thermal gradient of about 100^{o}C/km could result from hydrothermal convection to the water table below the cave. The cave is apparently a point of recharge, however, so the high 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 (1000 ft); water level is 206 m (675 ft) below the surface. The maximum surface temperature of the water was 40^{o}C (104^{o}F) after approximately 1 hour of pumping with a submersible pump, and the water probably had not cooled significantly from the temperature at depth.

With an average annual temperature of 17^{o}C and the maximum depth of 305 m, the thermal gradient in the well is at least 75^{o}C/km. Such a high gradient without convection is unlikely, thus, the water probably rose from greater depths by hydrothermal convection.

Hot Wells is near the extension of the Rim Rock fault as indicated by the gravity data of Wiley (1970), 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 (2013 ft) 8 km to the southeast of Hot Wells and near the trace of the Rim Rock fault, encountered a more normal thermal gradient of 32^{o}C/km and a maximum temperature of 38^{o}C (100^{o}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 towards basin centers, towards 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 occur near the basin centers (Groat, 1972). Coarse-grained facies of the late Tertiary basin fill are generally permeable to horizontal flow, but interlayered fine-grained facies probably restrict vertical permeability. Fine-grained basin-center facies are generally impermeable. Cold springs occur 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 a similar 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-center 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 shows that there are permeable channelways which allow circulation of ground water to a source of heat at depth, where the water is heated and returns to the surface. The setting of hot springs in this study area shows that most are on or near the normal fault system created during late Tertiary extension. A second, smaller group of hot springs occurs at the contact of permeable and impermeable rocks. One group of springs is adjacent to a reverse fault.

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 their 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 still the major conduits for hot water, but at Indian Hot Springs and Ojos Calientes the water is probably diverting 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 to the surface 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, occur 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 to 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 be shallow at Hot Springs-Ruidosa, where Tertiary volcanic rocks crop out and plunge gradually towards the spring.

No hot springs occur near basin centers. In the basin centers, faults are not common, and the unfaulted fine-grained basin fill has low permeability, which does not provide channelways for the rise of hot water. Even where faults occur, there are no hot springs. Bedrock is at considerable depth, and 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 at all. Thermal water in the Briscoe Well was discovered only by chance. Thus, there are probably many still 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 even within the dissected basins (Presidio and Hueco Bolsons) they are rare.

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 by thermal water might seal fractures. Recent fault movement must not be a prerequisite for deep circulation of ground water, however, because several hot springs occur in areas without evidence of recent fault activity.

The springs at San Carlos are also fault controlled; the only difference is in the 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 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 it to the water table (fig. 19). Where the water table intersects the surface, the hot water discharges as springs that are analogous to cold springs, which commonly occur at the contact of permeable and impermeable rock types. Recharge to a cold spring flows from a higher elevation through confined permeable rock to emerge where the contact intersects land surface. The driving force is gravity. The additional driving force caused by the different density of thermal water allows it to discharge above its recharge area.

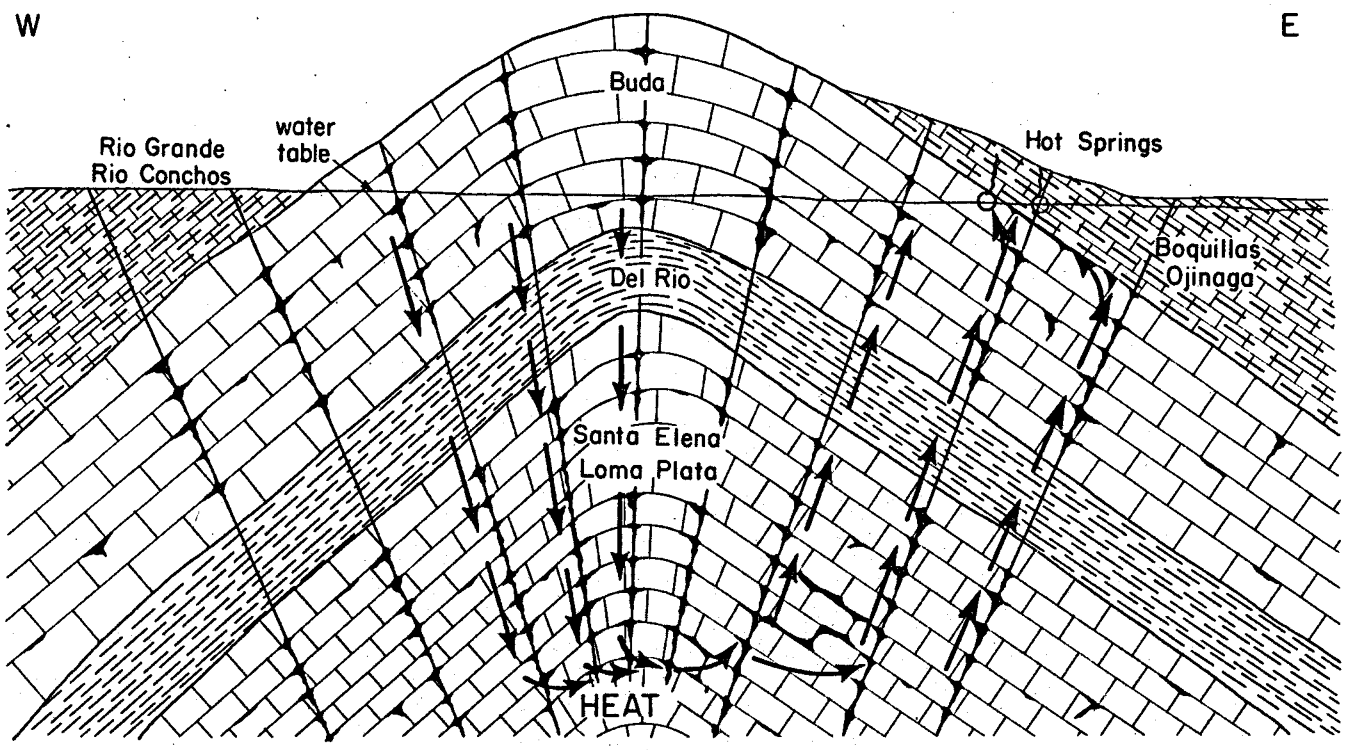

Figure 19. Postulated flow paths of hot spring waters at Peguis and Big Bend National Park.

Recharge for all the hot springs probably occurs mostly in the adjacent highlands and is spread over a broad area, whereas discharge is concentrated. Faults in the otherwise impermeable igneous rocks and faults and solution permeability 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, at least near hot springs. Downward-moving cold water would mix with upward-welling hot water and the hot spring would not occur. Thus, hot springs probably occur 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 occur 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, and Hot Wells and the Terlingua mines and well tap hot water that does not reach the surface. For all but the Gulf wells, the reason is simply that the water table is below the land surface. Whether or not it is rising from deeper convection, the hot water remains at the top of the water table far below the surface. Hot water in the Gulf wells occurs far below the water table. The temperature reversal in the wells proves that the water is rising from still greater depths--at least 2500 m according to the temperature log of the wells. Water must be rising to the bottom of an impermeable layer, possibly shales of the San Carlos or Ojinaga Formations, which 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 paper. Intersecting fractures and other permeable zones probably allow considerable mixing of cold and hot convecting water. A hot spring occurs where a somewhat fortuitous combination of circumstances 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 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 as near as possible to the discharge point. In practice this means that 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, especially the slowest discharging waters, could have undergone considerable chemical change before sampling, particularly regarding 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 1l was left unacidified, and 1l, acidified to a pH of 2 with distilled 70-percent 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 HNO_3, then with 10-percent HCl, then rinsed with distilled, deionized water. A third sample was collected for SiO_2 determination during the initial phase of sampling and for 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 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 sampling, only a few springs and wells which were suspected to have high silica concentration were diluted during sampling.

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

Analyzed Constituents[a] *General Analytical Methods*	1	2	3	4	5	6	7	8	9	10	11	11A	12	13	14
	INDIAN HOT SPRINGS						BIG BEND NATL. PARK				PEGUIS				
	Stump	Chief	Squaw	Beauty	Soda	Red Bull Spring	Spring at Rio Grande Village	Hot Springs	Big Bend #2	San Carlos Springs	Peguis	Peguis #2	Las Cienagas	Hot Springs—Ruidosa	Briscoe Well
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_3) *Electrometric titration at spring*	906	902	962	1034	623	474	280	273	266	207	243	244	415	290	281
Sulfate (SO_4) *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_3) *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_2) *Molybdate blue colorimetric*	40	40	35	35	20	36	21	22	22	37	22	21	39	35	39
pCO_2 *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_4	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_3	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

[a]Concentrations in milligrams per liter.

Table 3.—Concluded.

Analyzed Constituents[a] *General Analytical Methods*	15 Nixon Springs	16 Capote Warm Spring	17 Gulf–Presidio	18 Gulf–Swafford	19 OJOS CALIENTES #3	20 OJOS CALIENTES #4	21 Rancho Cipres	22 Hot Wells	23 Mimbre Well	24 Vizcaino Well	25 Naegele Springs	26 Mexican Springs	27 Nuñez Well	28 Alamo Springs
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_3) *Electrometric titration at spring*	220	226	497	563	756	783	1048	293	247	477	390	462	274	237
Sulfate (SO_4) *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_3) *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_2) *Molybdate blue colorimetric*	43	37	76	144[b]	95	87	46	20	51	60	43	39	52	22
pCO_2 *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	0.06	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[b]	69[c]	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_4	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_3	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

[a]Concentrations in milligrams per liter.
[b]187 reported.
[c]82°C reported.

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 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 presented graphically in fig. 20. Although there is wide variation in 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) calcium, magnesium, bicarbonate, sulfate waters with moderate dissolved solids; (2) low to moderate total dissolved solid waters 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 lowest calcium-magnesium ratios

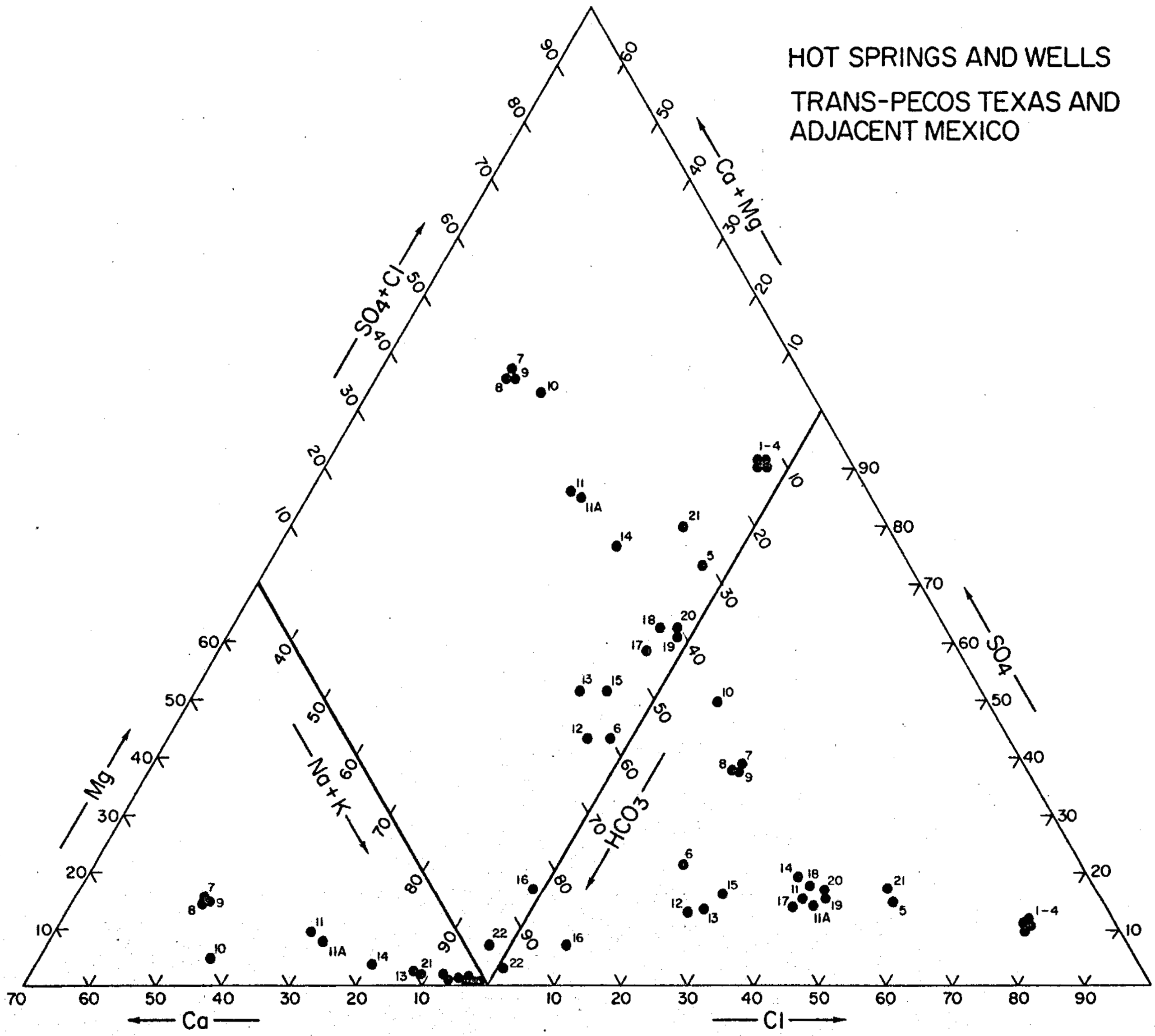

Figure 20a. Trilinear diagram of thermal and nonthermal waters discussed in text. Proportions are in moles per liter.

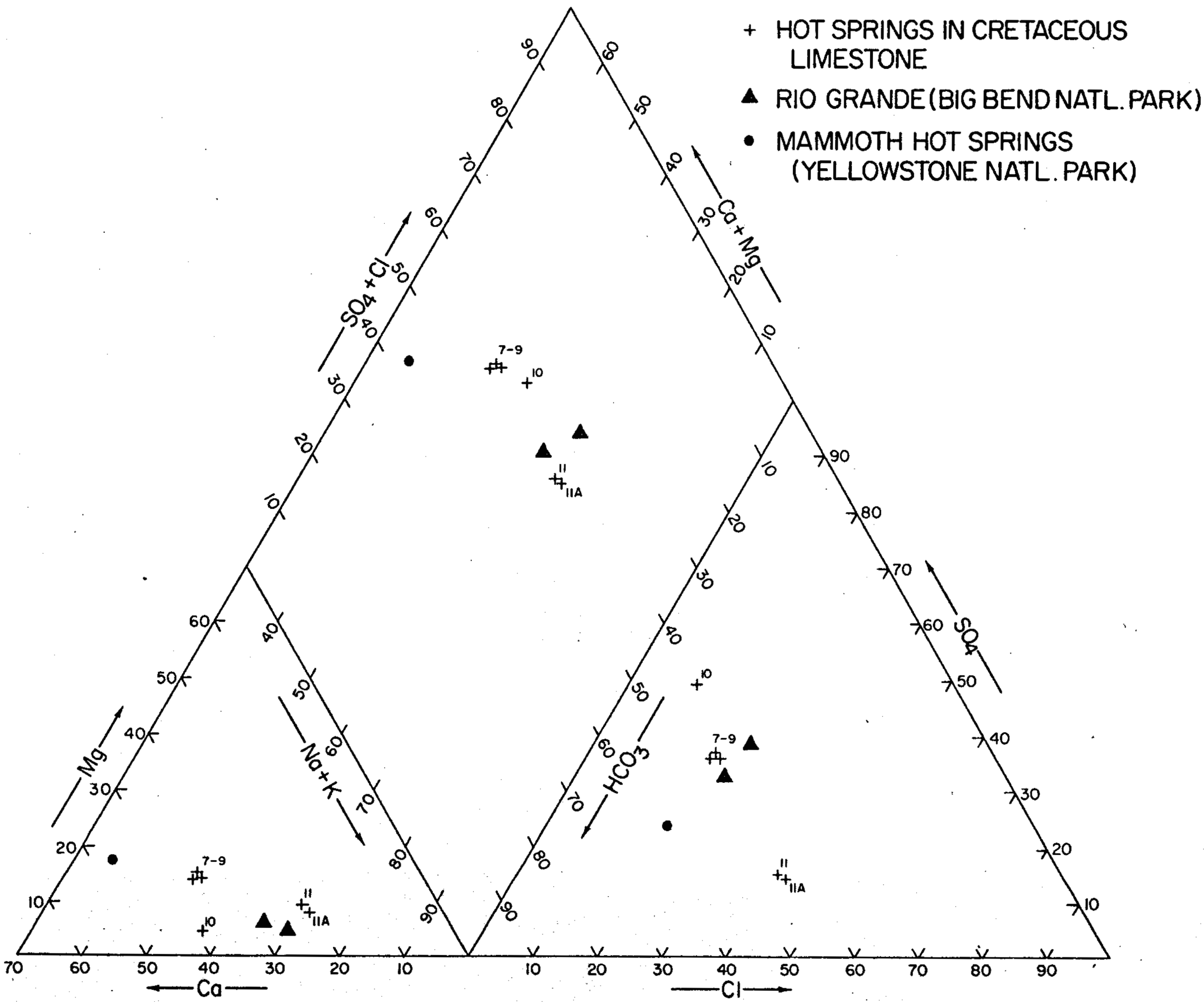

Figure 20b. Trilinear diagram of thermal and nonthermal waters discussed in text. Proportions are in moles per liter.

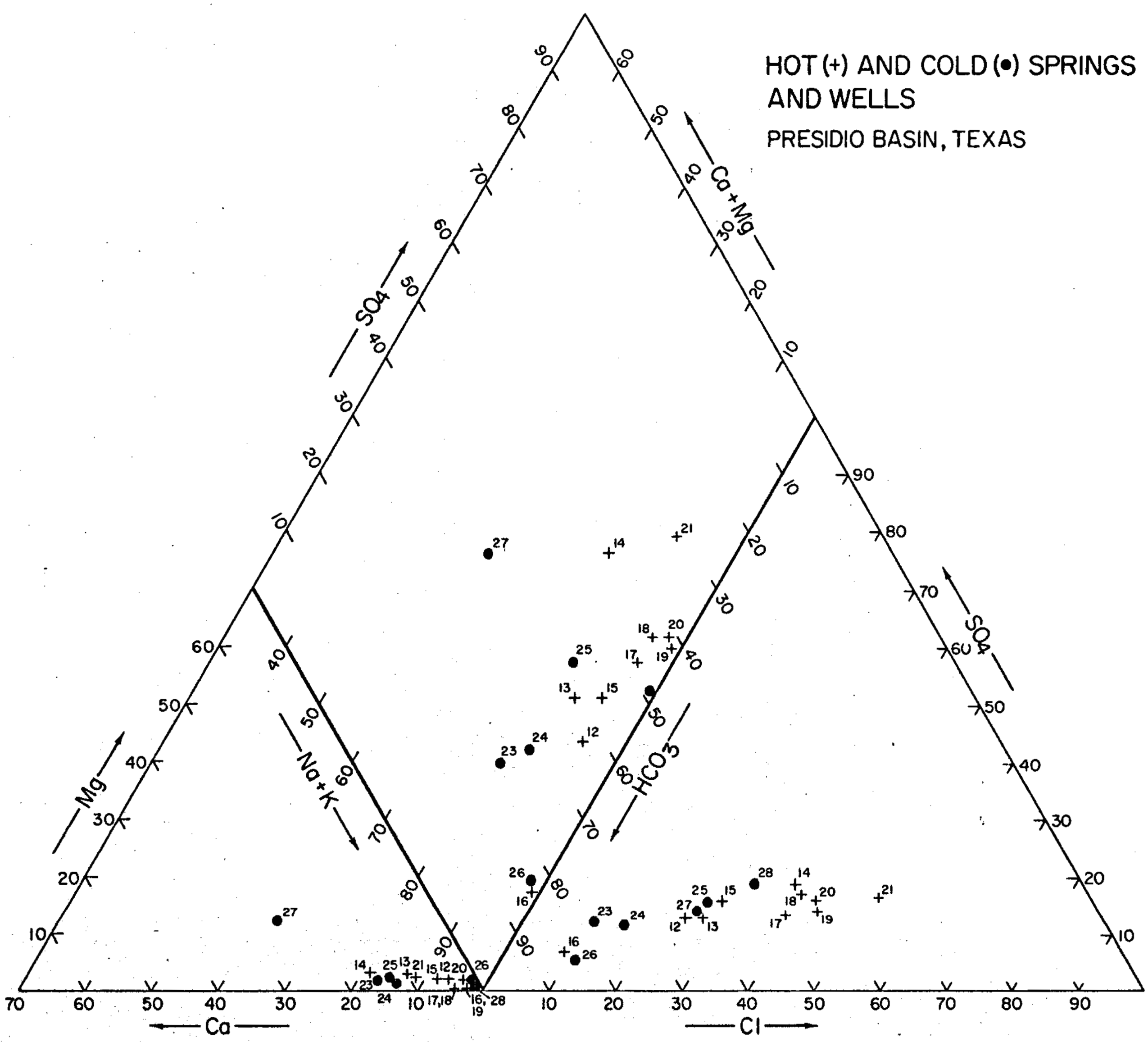

Figure 20c. Trilinear diagram of thermal and nonthermal waters discussed in text. Proportions are in moles per liter.

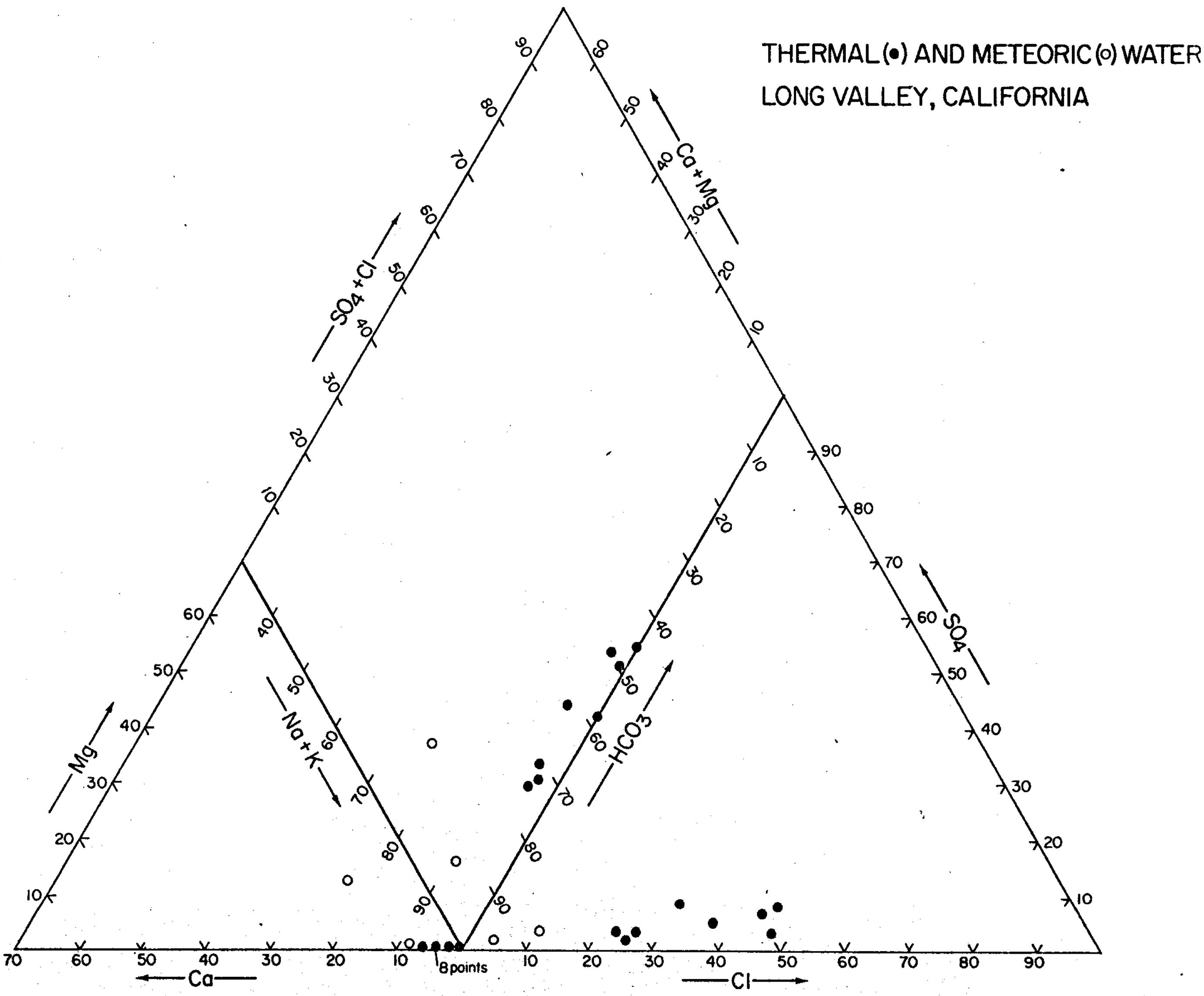

Figure 20d. Trilinear diagram of thermal and nonthermal waters discussed in text. Proportions are in moles per liter.

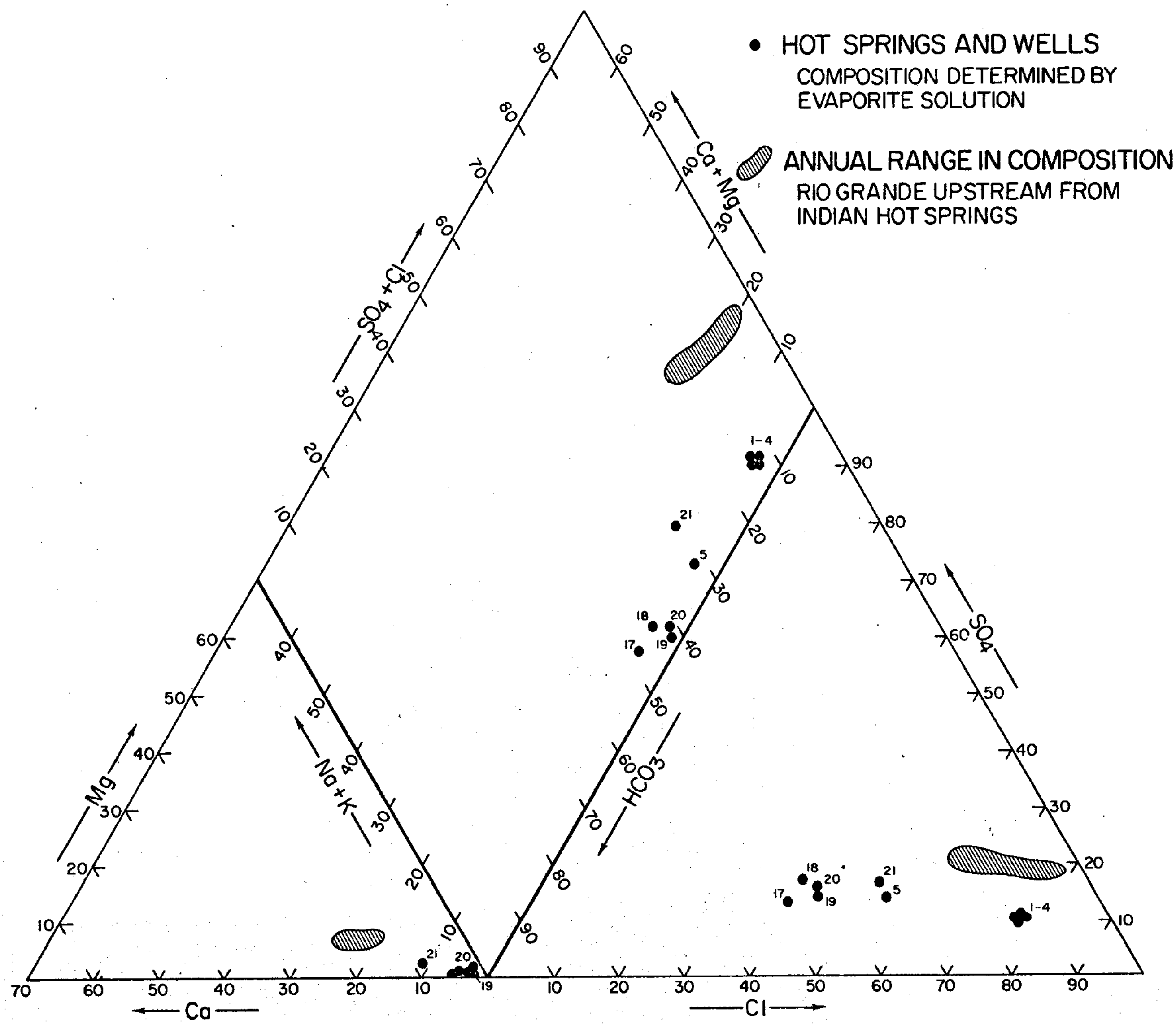

Figure 20e. Trilinear diagram of thermal and nonthermal waters discussed in text. Proportions are in moles per liter.

(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 it has similar ion proportions to springs in this study (fig. 20b) (Thompson and others, 1975). The geologic setting and reasonable estimation of water - host-rock reactions show that spring chemistry is determined mostly by solution of calcite and dolomite. Calcium, magnesium, and bicarbonate enter into solution by the following reaction.

$$(Me)\ CO_3 + H_2O + CO_2 = (Me^{++}) + 2\ HCO_3^-$$

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

Cold-spring waters emanating from carbonate rocks are composed almost entirely of calcium, magnesium, and bicarbonate with generally minor sodium, sulfate, or chloride (table 5 of White and others, 1963). 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 which require an additional source besides solution of carbonate materials.

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

$$Ca^{++} + Na - clay = 2\ Na^{+} + Ca - clay$$

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, such as oxidation of reduced sulfur (FeS_2 or H_2S). Oxidation reactions probably are occurring, but water in equilibrium with air contains only sufficient oxygen to produce about 15 mg/l sulfate by this process. Gypsum solution remains the only reasonable source of large amounts of sulfate and is consistent with the host-rock composition. Other sources for sulfate and chloride, such as magmatic gases, cannot be excluded but are unnecessary.

Recharge for the springs in Big Bend could come from the Rio Grande and for the springs at Peguis from the Rio Conchos. Because almost all of the Rio Grande flow below Presidio is from the Rio Conchos, the water chemistry of the two rivers is similar. Published analyses of the Rio Grande water upstream from the hot springs in Big Bend (U. S. Geological Survey, 1976) are similar to the Big Bend spring waters (fig. 20b). Because the Rio Conchos drains a predominantly carbonate terrain, recharge water from the river ought to be chemically similar to recharge water which percolates through limestone. The springs at San Carlos could be recharged by the Rio San Carlos, which drains both

Cretaceous carbonates and Tertiary volcanic rocks (fig. 18).

Sodium-Bicarbonate Waters

A second group of waters contains low to moderate total dissolved solids (300 to 1000 mg/l) with sodium and bicarbonate the dominant cation and anion, respectively (table 3; fig. 20c, #12,15,16,22). Sulfate and chloride concentrations are low to moderate; calcium and magnesium are uniformly low. Waters emanate from Tertiary rhyolitic volcanic and volcaniclastic rocks and basin-fill sediments derived from volcanic rocks. The volcanic rocks are commonly underlain by Cretaceous carbonate and clastic rocks that influence the chemistry of some of the springs (table 3; fig. 20c, #13,14).

Hot and cold springs and well waters derived from silicic igneous rocks throughout Trans-Pecos Texas are sodium bicarbonate waters, as are cold ground waters from silicic igneous rocks worldwide (table 1 of White and others, 1963). Spring waters from rhyolitic volcanic rocks in the Long Valley caldera in California have similar ion proportions (fig. 20d), although total dissolved solids in the Long Valley waters are 2 to 3 times those of the Trans-Pecos waters (Mariner and Willey, 1976). Lower calcium and magnesium concentrations may reflect the reduced solubility of calcite and magnesium absorption by clays (White, 1970) at the higher temperatures of thermal water in Long Valley. Spring waters in volcanic rocks in Yellowstone National Park (Truesdell and Fournier, 1976) are also sodium-bicarbonate waters, although they are again higher in total dissolved solids and have an additional source of chloride. These latter two areas have established subsurface temperatures above 200°C. Even though most if not all the Trans-Pecos waters were never that hot, the processes that determine general water chemistry are evidently similar. Hot waters in Trans-Pecos Texas are simply heated versions of the cold waters. No additional sources of dissolved constituents are necessary.

Sodium and bicarbonate are derived from hydrolysis of silicates or volcanic glass or are derived by cation exchange with zeolites in the volcanic rocks. Truesdell and Fournier (1976) give an appropriate chemical reaction:

$$CO_2 + H_2O + \text{Na silicate} = HCO_3^- + Na^+ + \text{H silicate}$$

This reaction also releases potassium, calcium, and magnesium into solution depending on their abundance and distribution in various silicate minerals in the rocks. Chemical analyses (Walton, 1975) of unaltered volcanic rocks show high sodium and potassium and low calcium and magnesium. Much of the volcanic section is zeolitized with high sodium but low potassium. 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 equilibrium with sodium-potassium-bearing feldspars and are a function of temperature. However, comparison of sodium-potassium ratios with measured and estimated temperatures of these waters does not show a clearcut relationship. Cation exchange of calcium for sodium may also be significant in keeping calcium concentrations low and sodium high. Hall (1963) suggested that cation exchange within a rhyolitic breccia converted calcium bicarbonate water to sodium bicarbonate water which discharged as hot springs in the Socorro, New Mexico, area. In any event, the cation chemistry of the spring waters is consistent with the rock chemistry.

Hydrolysis also releases silica into solution. Cold ground waters 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 kept in mind 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 and to 37 mole percent in Nunez Well, a cold well. Sulfate and chloride 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 in cold and hot waters 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 has little effect on the 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, #1,2,3,4,5,17,18,19,20,21). Calculated total dissolved solids for Indian Hot Springs are approximately 8000 mg/l; for

Rancho Cipres, approximately 4000 mg/l; and for Ojos Calientes, approximately 2200 mg/l (table 3). Even though 8000 mg/l is relatively dilute compared to sea water or many brines associated with evaporites (White and others, 1963, table 27), the waters are considerably more concentrated than 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 enriched in evaporite deposits. The three spring groups listed either fall within the Jurassic evaporite basin of Deford and Haenggi (1971) or at its edges where water could contact evaporites (fig. 21).

None of the hot spring waters is saturated with either halite or gypsum. Either the waters must not be in contact with evaporites for sufficient time to reach saturation or, more likely, waters circulating in other rock types mix with evaporite waters before discharge. If water contacting evaporites were saturated before mixing, the amount of water coming from evaporites could be small compared to nonevaporite water, and yet the dissolved solids would still be dominated by 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 to water from other rocks.

Waters derived from evaporites without 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

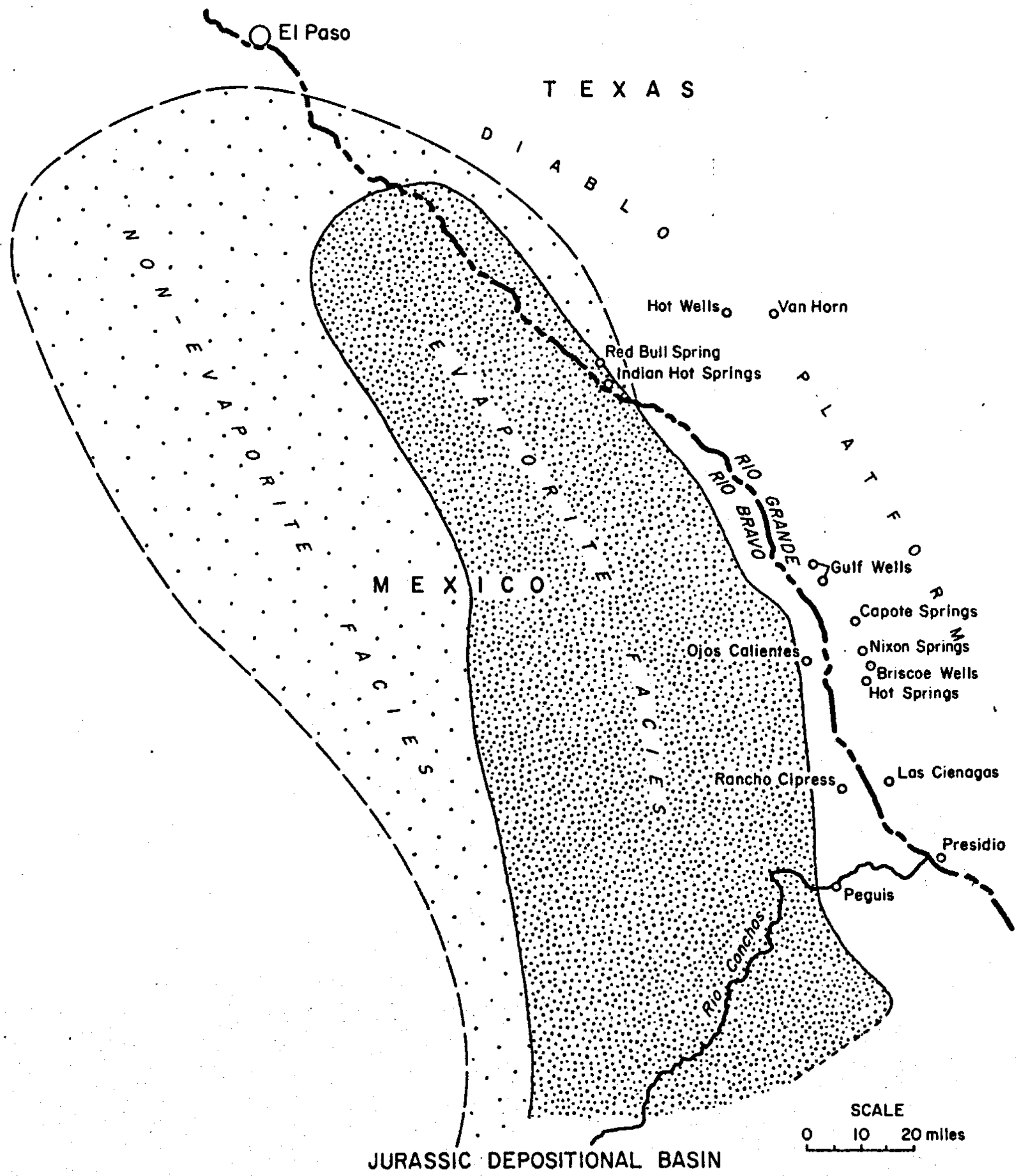

Figure 21. Location of hot springs and wells in Rio Grande area of Texas and Mexico and extent of Jurassic depositional basin including evaporite and nonevaporite facies. Jurassic basin from Deford and Haenggi (1971). Evaporite rocks now extend beyond original basin because of tectonic transport.

sodium-chloride ratios near 1, whereas other evaporite spring waters range up to approximately 2.6. Calcium could also be supplied by solution of limestone so that calcium-sulfate ratios ought to be more than 1. This problem is the same as noted for the carbonate waters. In this example, however, the evaporite component is much more dominant, and the shift in sodium-chloride ratios is considerably less. Simple mixing with nonevaporite waters with low dissolved solids could not produce these changes. Chemical reactions with other host rocks, such as the cation exchange postulated for carbonate waters, more readily account for the difference. Probably calcium is removed also 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 degasses carbon dioxide. For example Ojos Calientes waters have lower calcium and magnesium contents than either Indian Hot Springs or Rancho Cipres, even allowing for the lower total dissolved solids of Ojos Calientes compared to the other springs. All clearly evaporite-related springs deposit travertine at the surface and probably are also precipitating calcite in the shallow subsurface.

All of the springs that are 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 and 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 overshadowed by the high sodium, chloride, and sulfate contents derived from evaporites.

Recharge of the Indian Hot Springs thermal system from the Rio Grande is conceivable considering the physical setting of the springs. Inheritance of the chemical characteristics of Rio Grande water is also possible. Analyses of

Rio Grande water taken monthly for 1 year (U. S. 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 low discharge in late spring and early summer, total dissolved solids in the Rio Grande exceeds 5000 mg/l, approximately 65 percent that of Indian Hot Springs. However, with increased discharge during early fall, the amount of total dissolved solids decreases; the annual average is only about 2600 mg/l. Thus, Rio Grande water would have to be more concentrated to equal the spring waters. There are subtle chemical differences also. The river water, although high in sodium, is low in potassium (10 to 20 mg/l--well below spring concentrations of 130-200 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. Their total dissolved solids (approximately 1200 and 1700 mg/l) are slightly greater than dissolved solids in nonevaporite waters but considerably less than dissolved solids in waters obviously derived from evaporites. Likewise, their sodium, chloride, and sulfate contents are relatively high but lower than those in the distinctive evaporite waters. Other major ions do not distinguish the Gulf wells, but they share some characteristics with evaporite waters which clearly distinguish them. In ion proportions, they are similar to Ojos Calientes waters and unlike most others. Also, potassium, lithium, and boron concentrations are distinctly greater than those of any of the nonevaporite springs and are only modestly diluted compared to 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 these ions.

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 occur in the subsurface. Water discharged from the Gulf wells probably has been in contact with, and derived a major part of their 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 (1000 mg/l), although slightly greater than that of most obviously nonevaporite springs, is lower than Indian Hot Springs. It has moderately high sulfate, sodium, and bicarbonate contents, but low chloride and lithium. Only a minor percentage of its water and dissolved salts can be derived from evaporites. Its circulation system is possibly too shallow to contact evaporites.

The hot springs at Peguis are also at the edge of the evaporite basin (fig. 21) but the water chemistry falls within the carbonate group discussed above. A 3660 m (12,000 ft) deep Pemex exploration well drilled 7 km north of the springs did not penetrate evaporites, although it is interpreted to have 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 like the evaporite waters. Evaporites apparently do not occur in the shallow subsurface where cold, shallow ground water could contact them. The chemistry of the evaporite waters is determined deeper within the circulation system than the chemistry of other thermal waters.

Calcite Saturation

Except for one hot spring and one cold spring, all the waters of this study are in equilibrium or oversaturated with calcite. Saturation indices (table 3) are equal to the log base 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, 1974). Water with indices between $\pm$ 0.1 are generally considered saturated; waters with indices above or below this range are over and undersaturated, respectively.

Calcite saturation should be expected for both travertine-depositing springs and springs emanating from limestone but not for sodium bicarbonate waters. Although sodium bicarbonate waters have low calcium content, they have relatively higher carbonate content compared to the calcium-magnesium-bicarbonate waters. Evidently, waters circulating within volcanic rocks or bolson sediments contact sufficient calcite to become saturated.

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. Precipitation of calcite then could occur by the following reaction:

$$Ca^{++} + 2\ HCO_3^- = CaCO_3 + H_2O$$

Some spring waters oversaturated with calcite do not precipitate calcite, which is presumably a problem of kinetics 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 Cienagas. Rapid loss of carbon dioxide from evaporite waters leads to oversaturation and precipitation. Indian Hot Springs waters illustrate this point. Most of the springs discharge at the bottoms of pools that bubble gas, presumably carbon dioxide. Stump Spring discharges rapidly to the surface, has the highest temperature, lowest pH, and highest partial pressure of carbon dioxide. 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) so that 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. Springs that are in equilibrium with calcite and have not lost much carbon dioxide must have reached equilibrium approximately at the spring surface temperature. These thermal waters either were never much hotter than their surface temperatures or reequilibrated near the surface during cooling. In the latter case, any record of equilibrium reached at higher temperature at greater depth will have been lost.

Partial pressures of carbon dioxide in West Texas waters range from 2.2×10^{-4}, near the atmospheric equilibrium value of 3.2×10^{-4} (Keeling, 1958), to 2×10^{-1}, several orders of magnitude greater than atmospheric equilibrium. The source of this excess carbon dioxide is uncertain. High partial pressures of carbon dioxide are commonly attributed to percolation of water through a soil zone enriched in carbon dioxide from 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, which is commonly cited in the Russian literature (Hem, 1970), implies extremely high temperatures at depth. Other possible sources of carbon dioxide (for example, reduction of sulfate by hydrocarbons) are unlikely.

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 (H or H^1), and of mass 2, deuterium (D or H^2). Oxygen has three isotopes, mass 16 (O^{16}), mass 17 (O^{17}), and mass 18 (O^{18}). Only O^{16} and O^{18} are commonly reported. Isotope ratios of both hydrogen and oxygen vary in nature and provide information about the origin and history of natural materials containing them.

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 (del or δ) is reported as a parts per thousand (per mil or $^{o}/oo$) difference between the isotope ratio of the sample and SMOW.

Natural waters are depleted in D and O^{18} compared to SMOW. The lighter isotopes are concentrated in the vapor during evaporation. During condensation, the heavier isotopes are concentrated in the condensate. Thus, atmospheric moisture is continuously depleted in D and O^{18} as it moves away from the ocean source. The amount of depletion increases with distance from the source, latitude, and altitude. Meteoric waters show a constant relationship between del D and del O^{18} shown in figure 22 as the meteoric water line. Sheppard and others (1969) have mapped the distribution of del D in meteoric water for 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 same area but are enriched in O^{18}. This oxygen shift is attributed to isotope 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 to those in water.

Results

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

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

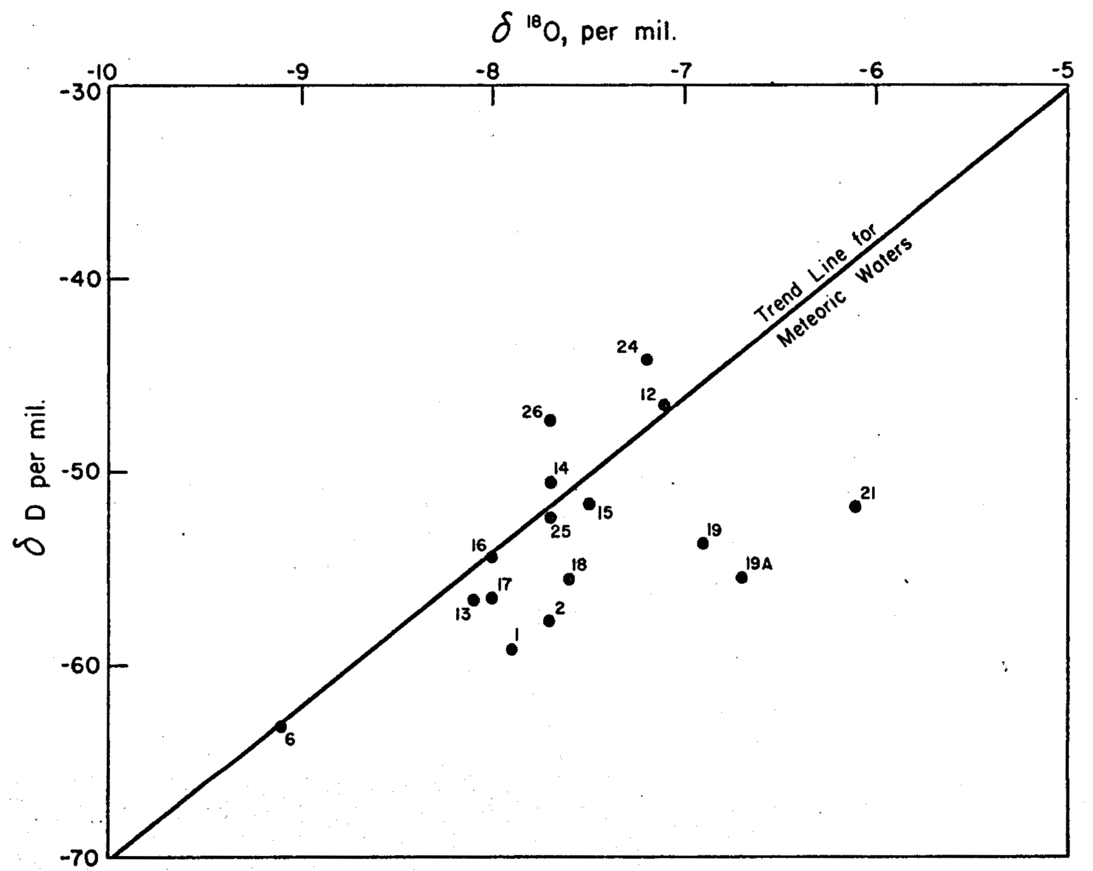

Figure 22. Isotopic composition of thermal and nonthermal waters, Rio Grande region of Texas and Mexico.

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

#	Spring or Well	δD SMOW ‰	δO^{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	Naegele Springs	-52	-7.7
26	Mexican Springs	-47	-7.7

A slight geographic variation is shown by the relative depletion of Red Bull and Indian Hot Springs water relative to those in the Presidio graben. The 10 per mil scatter in del 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 were never heated to sufficient temperatures for measurable exchange. The latter is consistent with geothermometry determinations presented in this section and is more likely.

The hydrogen-isotope composition of Peguis water is slightly different from the composition of all other waters of this study. The oxygen composition has not been determined, so 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 to Peguis area meteoric water. Although analyses of Rio Conchos water are not available to substantiate this hypothesis, the low del D shows that the Peguis thermal system is being recharged by the Rio Conchos. The isotope composition of Indian Hot Springs water indicates that recharge is probably not from the Rio Grande, however.

Most of the evaporite waters (fig. 21, #1, 2, 19, 19A, 21) do not plot on the meteoric water line. Gulf-Presidio water (fig. 22) cannot be distinguished from meteoric water; Gulf-Swafford water (fig. 22, #18) is approximately at the 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 that shown by most high-temperature spring waters.

Geothermometry calculations for the waters from Indian Hot Springs (fig. 22, #1, 2) and Rancho Cipres (fig. 22 #21) indicate subsurface temperatures no greater than about 60°C. The oxygen shift must have some source other than high-temperature exchange.

There are several possible explanations for the shift. Meteoric water near Indian Hot Springs probably has an isotopic composition like that of Red Bull spring water (fig. 22, #6). If so, Indian Hot Springs waters show both an oxygen and 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 isotope shift must have occurred during deep circulation, possibly by mixing of meteoric water and an isotopically heavy formation water within the evaporites. Persistence of formation water in Jurassic or older rocks during considerable structural deformation is difficult to conceive but not impossible.

Meteoric recharge water for Ojos Calientes, Gulf-Swafford, and Rancho Cipres ought to have a composition within the range of Presidio graben waters (fig. 22, #12, 13, 14, 15, 16, 24, 25, 26). If so, the oxygen shift for these waters does not follow an evaporation line. An explanation involving formation water 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^{o}C but is not possible for either Indian Hot Springs or Rancho Cipres waters.

In summary, all of the thermal waters of this study area are entirely or dominantly meteoric. All but the evaporite waters have retained meteoric isotope composition. Mixing of meteoric water and an isotopically heavy formation water could explain the isotope composition 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 isotope 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 hot springs waters of this area is similar to the chemistry of cold waters in the same area. The composition range and stable isotope ratios of the sodium-bicarbonate and calcium-magnesium-bicarbonate-sulfate waters overlap with 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. The similarity in chemistry indicates that most of the dissolved constituents are acquired early in circulation history and that deep circulation and heating have little effect on the water chemistry.

There are no cold spring waters chemically like 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 and are calcium-magnesium-bicarbonate waters. The final composition is then 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-Swafford, it may result from high temperature oxygen exchange with wall rocks.

All 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. No extra source besides rock leaching is required to provide all the dissolved constituents with the possible exception of carbon dioxide. Carbon dioxide may also be derived from rock leaching, but the exact chemical reaction is unknown. Sources for some constituents, such as chloride and sulfate from degassing of igneous magmas, are unnecessary, although such sources cannot be disproved.

That distinct chemical groups of thermal waters exist in a complex geologic terrain implies that flow paths are highly localized. That is, springs must be discharging water that was recharged and circulated within a very restricted geographic area. If circulation were broader, the springs would discharge more homogeneous waters. For example, in the northern part of the 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 Spring, Nixon Spring, Hot Springs, and Briscoe Well, have distinctive water chemistry that reflects rocks types exposed at the spring, in the shallow subsurface 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 as vertical changes in rock type would also tend to homogenize the thermal water.

This concept 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 at Capote Springs. Flow path B is entirely in limestone; water discharging at spring B' would be a calcium-magnesium-bicarbonate-sulfate water like those of the Big Bend National Park springs. Flow path C intersects both volcanic rocks and carbonates; water discharged from spring C' should have a composition intermediate between a sodium-bicarbonate and a calcium-magnesium-bicarbonate-sulfate water such as occurs in Briscoe Well. Mixing of 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 wallrock reaction on a homogeneous reservoir water at depth.

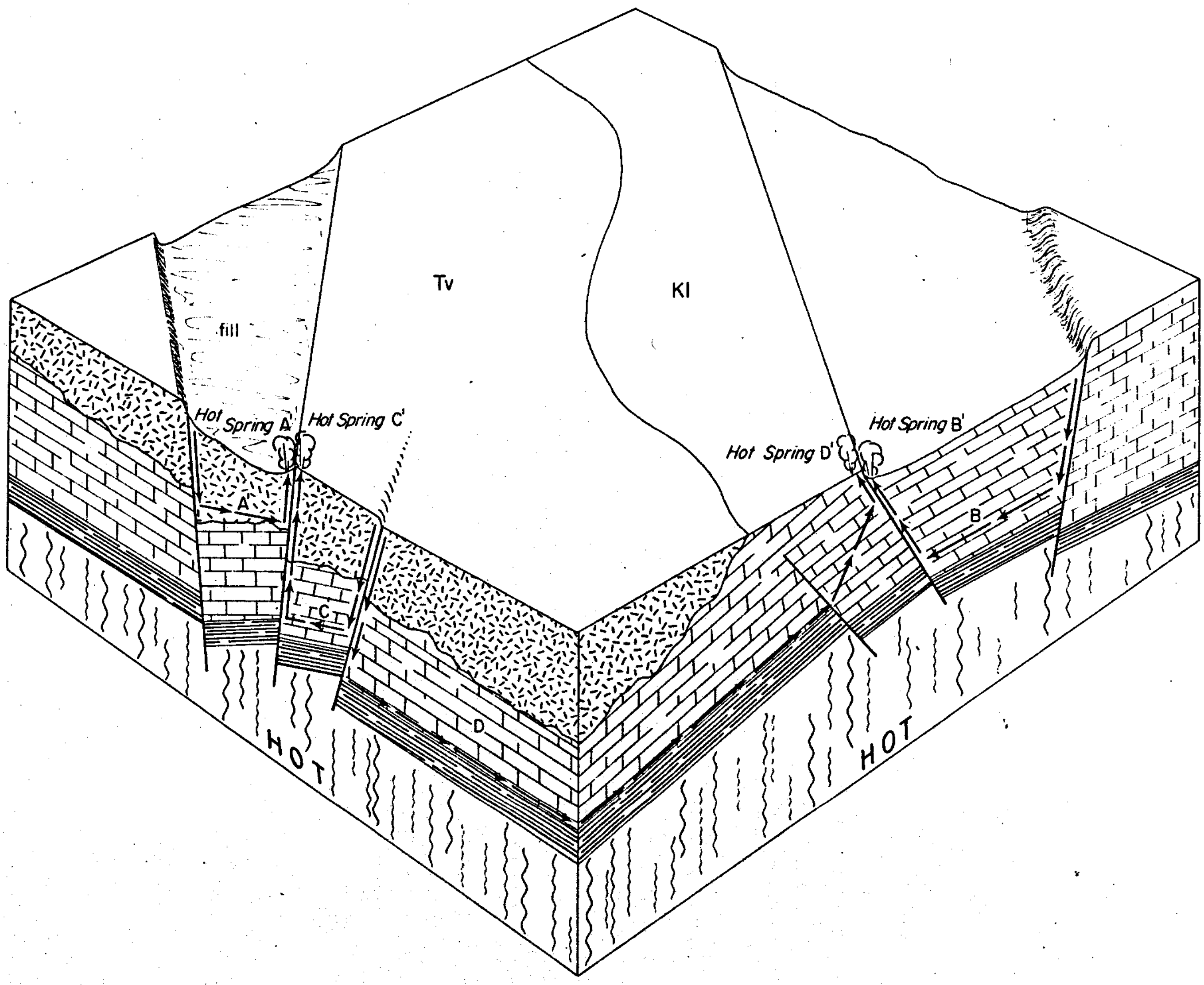

Figure 23. Postulated circulation of water in thermal convection systems showing fault control of flow paths and lithologic control of thermal water chemistry.

GEOTHERMOMETRY

Introduction

One of the few well-developed and commonly used tools in geothermal exploration is geothermometry. Geothermometry is used to determine maximum subsurface temperatures from water chemistry by assuming that some aspect of the chemical composition of the geothermal water is controlled by maximum temperature. Even though 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) the 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 reequilibration or change in composition at lower temperature; and (5) the thermal water does not mix with cooler shallow ground water. Fournier and others (1974) warn that the assumptions are not always valid and that interpretation of results is simpler for springs with high temperatures and discharges than for springs of low temperatures or discharges.

There are several qualitative geothermometers, but only two are sufficiently calibrated to be quantitative. They are 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, therefore, the silica content reflects the maximum temperature of the water. Absolute solubility, however, is also a function of the silica phase. Quartz is least soluble, followed by chalcedony, cristobalite, and amorphous silica, which is most soluble.

The method, developed by Fournier and Rowe (1966) and Mahon (1966), commonly assumes that equilibrium with quartz is attained at high reservoir temperatures and 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 temperature. However, for systems at lower temperatures, equilibrium may not occur or equilibrium may be with another silica phase. At temperatures around 100°C and even up to 180°C (Arnorson, 1975), equilibrium may be with chalcedony. Water in contact with rocks containing amorphous silica may contain 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 kinetics of reaction, 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 precipitation of silica as water cools to about 180°C is rapid but that the rate drops greatly below that temperature.

Sodium-Potassium-Calcium Method

The sodium-potassium method (White, 1965; Ellis, 1970) and the sodium-potassium-calcium method derived from it (Fournier and Truesdell, 1973) 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 that incorporates calcium but still assumes equilibrium with feldspars. They found a linear relationship between the function $\log \left(\frac{Na}{K}\right) + \beta \log \left(\frac{\sqrt{Ca}}{Na}\right)$ and inverse temperature. The factor Beta (β) is derived from the stoichiometry of feldspar reactions and is either 4/3 if the water equilibrated below $100^{\circ}C$ or 1/3 if equilibrated above $100^{\circ}C$.

Problems with the sodium-potassium and sodium-potassium-calcium method can arise if equilibrium does not occur or if equilibrium is with minerals other than feldspars. Although equilibrium can be established at low temperatures (less than $50^{\circ}C$) given sufficient time, equilibrium probably should not be assumed for waters not heated well above $100^{\circ}C$. Equilibrium with minerals other than feldspars, such as zeolites which occur in rocks of this study, does not follow the same temperature relationship as feldspars. Subsurface temperatures for such waters cannot be predicted by the sodium-potassium-calcium method. Precipitation of calcite by travertine-depositing springs without reequilibration of sodium and potassium leads to estimated temperatures that are too high (Fournier and Truesdell, 1973). Dilution affects sodium-potassium-calcium temperatures less than it does silica temperatures because sodium-potassium-calcium temperatures are partly based on ratios of ions.

Qualitative Geothermometers

There are many qualitative geothermometers that ideally indicate whether thermal waters are part of relatively high or low temperature systems or hot water versus vapor-dominated systems (White, 1970). These geothermometers include calcium-bicarbonate ratios, magnesium concentrations and magnesium-calcium ratios, sodium-calcium ratios, chloride-fluoride ratios, stable isotope ratios, and siliceous sinter versus 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.

Results

Table 5 presents subsurface reservoir temperatures by the silica method assuming equilibrium with quartz and chalcedony and by the sodium-potassium and sodium-potassium-calcium methods using β equals 1/3 and 4/3. Interpretation of the geothermometry results can best be made by relating them to the geologic setting and geochemistry of the thermal waters.

Table 5. Calculated subsurface temperatures (°C) by Si and Na-K and Na-K-Ca methods.[a]

#	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)[b]	135(154)[b]	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	Mimbre 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

[a]See text discussion for interpretation of subsurface temperatures.

[b]Using published SiO_2 concentrations.

Geothermometry of Cold- to Intermediate-Temperature Springs and Wells

Subsurface-temperature estimates range up to 159°C assuming quartz equilibrium and to 135°C assuming chalcedony equilibrium. This is within the temperature range suggested by Arnorson (1975) as a transition between quartz and chalcedony equilibrium. The hottest springs and wells (Ojos Calientes and Gulf wells) have the highest silica contents (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. First, the high-silica cold waters may actually be thermal waters that have cooled before sampling either through conduction of heat into wall rocks or by dilution with nonthermal ground waters. However, several of the cold waters are from wells drilled near or into fault zones (for example, Vizcaino, Mimbre, and Nuñez) where thermal water could be upwelling. It seems unlikely that thermal waters in this setting could have cooled completely to ambient temperatures. Dilution with nonthermal water implies that the silica content should be still higher, implying still higher reservoir temperatures.

A second possibility seems more reasonable: that the waters are unrelated to thermal activity and that the high silica content results from dissolution

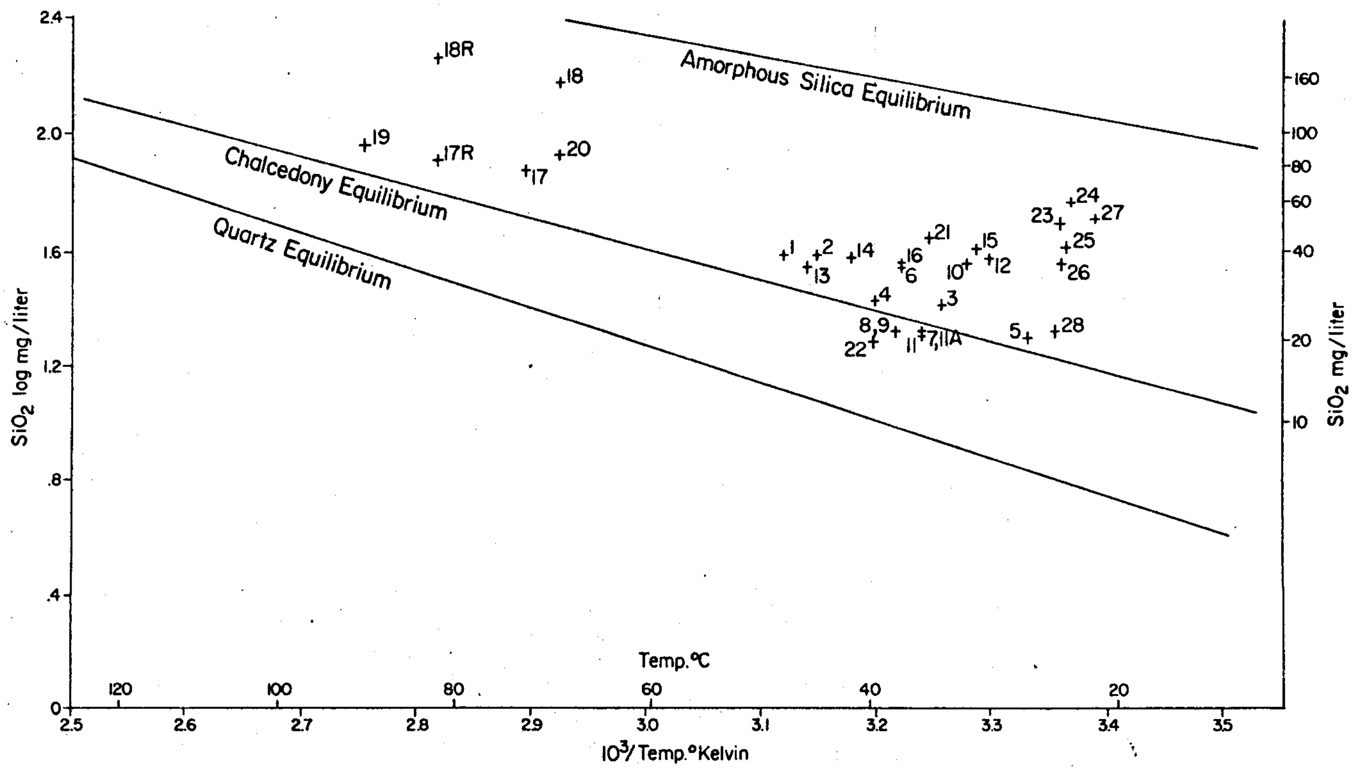

Figure 24. Measured temperature versus log silica concentration in thermal and nonthermal waters of this report. Silica phase equilibrium lines from Fournier (1976). The letter "R" after number indicates reported value from table 4.

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 that are rich in volcanic glass and opal (Walton, 1975); both the volcanic glass and opal are soluble forms of silica. Hydrolysis of volcanic glass or silicate minerals releases silica into solution. If the waters were never heated sufficiently to reach equilibrium with quartz, which seems likely, high silica concentrations could persist. The most logical explanation for many of the high-silica waters in Trans-Pecos Texas is the dissolution of opal or amorphous silica from igneous rocks and the persistent nonequilibrium of the waters with respect to quartz. This hypothesis is consistent with published analyses (table 1 of White and others, 1963) of cold ground water emanating from silicic igneous rocks for which silica concentrations ranged from 10 to 76 ppm.

This conclusion creates problems for interpreting the geothermometry of thermal springs. 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 give a misleading picture of the geothermal potential of an area. A common assumption is that dilution of thermal waters with nonthermal waters will lower the silica content. Dilution with these cold waters could, in fact, raise the silica content of most of the hot waters of this study.

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, there is no apparent increase in silica content with increase in temperature. Most of the samples contain approximately 40 mg/l. This relationship implies two possibilities: (1) thermal waters for most of the samples come from the same reservoir, which has approximately 40 mg/l silica or (2) the samples represent waters that have circulated to similar depths in areas of similar geothermal gradient. Both factors are probably important. All but two of the thermal springs or wells with a silica content of approximately 40 mg/l come from the northern part of the Presidio Bolson or the Indian Hot Springs area. For each area, the first possibility of a related reservoir may be important. Between the two areas, the second possibility of similar depth of circulation in a similar heat-flow province is probable despite the considerable differences in water chemistry between springs of the two areas.

Are these samples in equilibrium? If they are, what does this fact imply about reservoir temperatures? The cold waters in the northern Presidio basin, which presumably recharge the thermal waters, have similar if not greater concentrations of silica so that the silica in the thermal waters could simply be inherited. For the Indian Hot Springs area and several springs from Big Bend National Park that have lower silica content, this is not a problem. 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 equilibrium at low temperatures is a tenuous assumption. However, equilibrium with chalcedony 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 use in 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 not much above their surface temperatures. Many of the springs have surface temperatures of approximately 40° to 47°C. Two of the springs with lowest flow rates, Nixon and Las Cienagas, which presumably could have cooled the most from their reservoir temperatures, 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 secondary chalcedony nodules (Groat, 1972), and volcaniclastic sediments contain a variety of secondary silica minerals (Walton, 1972).

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 exclusively in contact with Cretaceous limestones containing chert. They have high rates of flow (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 previously listed. The narrow range in silica content of the warm spring waters could indicate only minor dilution of thermal water with a maximum silica content 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 contents. Mixing might not lower the silica contents 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 content, 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 equilibrium with feldspars (Fournier and Truesdell, 1973), but many of the spring waters have been in contact with evaporites or zeolitized volcanic rocks which contribute sodium, potassium, and calcium. Thus, results for the West Texas waters would not be meaningful. The problem is illustrated by the divergence between many silica and sodium-potassium-calcium temperatures, for example, Ojos Calientes.

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

Many estimated temperatures for sodium-bicarbonate waters determined by the β equals 4/3 method agree reasonably well with chalcedony temperatures for the same waters; however, many others do not. Attempts to explain the sodium-potassium-calcium results can lead to circular reasoning. It is better to recognize that the method does not work 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 short residence time. The cause for the slight isotopic shift for Indian Hot Springs and Rancho Cipres waters is not certain but probably is not high-temperature exchange.

In this study, the high silica content in many cold ground waters, 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 approximately 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 Presidio and Hueco Bolsons areas 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. The Presidio and Hueco Bolsons are in areas of recent faulting and relatively high thermal gradient and have the only known high-temperature thermal systems. There is no evidence for recent faulting in the Big Bend area, and heat flow and thermal gradients are apparently lower than in the Presidio area.

Geothermometry of Gulf Wells and Ojos Calientes

The geothermometry of Gulf wells and Ojos Calientes needs to be discussed separately because they are the only known thermal systems which show potential for power generation. Estimated silica temperatures are 112° and 94°C for Gulf-Presidio, 159° and 135°C for Gulf-Swafford, and 134° and 107°C for Ojos Calientes for quartz and chalcedony equilibrium, respectively (table 5). Published silica content for Gulf-Swafford (White and others, 1977) gives 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 a transition between chalcedony and quartz equilibrium. Interpretation of these temperatures is arguable. Silica content versus surface temperature plots of Ojos Calientes and Gulf-Presidio, like the intermediate temperature springs, are near the chalcedony equilibrium curve suggesting that their maximum temperatures are only slightly higher than their surface temperatures.

The results of other geothermometers are also equivocal. Temperatures by the sodium-potassium-calcium method are all around 200^{o}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 results suggest high temperatures (greater than 100^{o}C) for Ojos Calientes waters which show an isotopic shift; isotopic results possibly suggest high temperatures for Gulf-Swafford waters which show an oxygen shift, but the results do not suggest high temperatures for Gulf-Presidio water which is 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 their surface discharge temperatures. However, the waters probably have reequilibrated with limestone during cooling from their maximum temperatures. Travertine deposition simply indicates this reequilibration. Reequilibration should be expected because the producing reservoir for both Gulf wells is Cretaceous limestone and Ojos Calientes rises along a fault cutting Cretaceous limestone. For the same reason, magnesium content and magnesium-caloium 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 which 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-Swafford is far from equilibrium with either quartz or chalcedony. Its high silica content and implied high equilibrium temperatures are surprising in two respects. First, the well produces water at approximately

$80^{o}C$. Wells in high-temperature reservoirs should produce water near reservoir temperatures. The temperature log of Gulf-Swafford shows that the hot-water-producing zone must tap convective water that circulates from still greater depth. It is possible that the sampled water has cooled through conduction or by mixing with cooler ground water at the higher level, but mixing would require that the reservoir silica content and temperature be still greater. The other surprising aspect 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 is somewhat more dilute but only by about 25 percent for most constituents other than silica. It is conceivable that diluting water is low in silica but that the concentrations of other dissolved constituents are not much less than the concentrations in the thermal water. Another explanation is that Gulf-Presidio is farther from the source of thermal water than is Gulf-Swafford. Silica has precipitated during cooling, producing the difference in concentration.

No clear conclusions can be made from the available information. Ojos Calientes and Gulf-Presidio may be in equilibrium with chalcedony with maximum subsurface temperatures around $100^{o}C$. Quartz equilibrium at temperatures around $130^{o}C$ or dilution of greater concentrations is possible. On the basis of its geologic setting and water chemistry, Gulf-Swafford 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-Swafford have some source other than high reservoir temperatures. Because of the potential 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 such as 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 rates of discharge can originate either through deep circulation of meteoric water, which is heated only to its approximate discharge temperature, or by mixing of a much higher temperature water with cold, shallow ground water.

Determination of whether or not mixing has occurred is commonly made with the sodium-potassium-calcium geothermometer. Water with sodium-potassium-calcium temperature within 25°C of the discharge temperature is in equilibrium and has probably not undergone dilution, whereas water very much out of equilibrium probably has. Because of the previously described problems of applying the sodium-potassium-calcium method, this test cannot be used for either Ojos Calientes or the Gulf wells. More qualitative tests, such as variations among related waters of nonreactive constituents (for example, chloride variation with temperature), on the meager available data, suggest that water from Ojos Calientes has not been diluted, but that waters from the Gulf wells have undergone mixing.

Waters that have undergone mixing 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 in the two springs differ appreciably. Also, the lower pH and higher bicarbonate content and partial pressure of carbon dioxide of Ojos Calientes #4, the cooler of the two, suggest that this spring should have experienced less dilution than did Ojos Calientes #3. There are numerous springs at Ojos Calientes, and analysis of only two of them is not conclusive. Additional sampling to evaluate mixing is 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, the wells are physically separated, and the waters may not be genetically related so that mixing models using the two may not be valid. Nevertheless, their chemistry suggests that water from Gulf-Presidio is a more dilute version of water from Gulf-Swafford. Dilution would have had to take 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. Under these circumstances, mixing is particularly difficult to evaluate. Because of the implications for geothermometry, the possibility of mixing ought to 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. No hot springs in the Rio Grande rift have silica contents above approximately 100 mg/l with the exception of springs in the vicinity of the Valles caldera in northern New Mexico (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 content 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 the Valles area, the hottest springs and highest silica concentrations occur in several thermal areas in the rift in Colorado. Mount Princeton Hot Springs, Poncha Hot Springs, and hot springs in the upper San Luis Valley have maximum surface temperatures from 60° to 82°C and maximum quartz equilibrium temperatures from 100° to 125°C (Pearl and Barrett, 1976). Pearl and Barrett did not provide data concerning silica content, but these temperatures indicate approximately 60 to 80 mg/l silica. Olson and Dellechaie (1976) give 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 content of Ojos Calientes and Gulf-Presidio. The silica content of thermal water from Gulf-Swafford is considerably greater than any of the thermal systems in the identified Rio Grande rift. By this standard the Trans-Pecos Texas region at least has the geothermal potential of the Rio Grande rift.

A major difference in interpretation is that Pearl and Barrett assumed quartz equilibrium for all springs. Chalcedony equilibrium is more likely for some springs in Texas and quartz equilibrium is arguable for even the highest temperature thermal waters. Pearl and Barrett interpreted sodium-potassium-calcium and mixing model geothermometers to give temperatures up to 200^{o}C. Evaluations by these methods cannot yet be made for the thermal waters of this study. Nevertheless, available evidence shows that the Rio Grande area of Texas and Mexico deserves as much consideration as any part of the Rio Grande rift.

GEOTHERMAL MODEL AND AREA EVALUATION

Geothermal Model

With available evidence, a model of ground-water flow, depth of circulation, source and magnitude of heat, and geothermal potential can be constructed for Trans-Pecos Texas. Local meteoric water circulates downward through an intersecting net of permeable fractures created by late Tertiary to, in some areas, Recent crustal extension. Some rock types, such as massive, cavernous limestones, are permeable without fracturing, although most others are not. Water reaches depths where it is heated by the normal thermal gradient. The heated water is less dense and 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 proportion 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 the Presidio and Hueco Bolsons. However, some hot springs occur in areas without evidence of recent faulting. In addition, Ojos Calientes, the hottest thermal activity in the area, occurs along a major normal fault for which there is no evidence of recent movement. Cooler hot springs on the opposite side of the basin from 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 the 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. Chemistry of the springs indicates numerous separate thermal convection systems.

Source of heat is the Earth's normal thermal gradient which increases from Great Plains values in the eastern part of Trans-Pecos Texas to higher, at least Basin and Range values, along the Rio Grande. Still higher heat flow, as along the Rio Grande rift in New Mexico, is possible and even suggested by available data. Thermal gradients are at least $30^{o}C/km$ and possibly higher 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 increased thermal gradient caused by blockage of normal heat flow by low-conductivity rocks, particularly the thick, water-saturated basin-fill sediments.

Presidio and Hueco Bolsons

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 according to available evidence. Presidio and Hueco Bolsons represent the most likely extension into Texas of the Rio Grande rift and its associated high heat flow. 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.

In Presidio Bolson there are two groups of thermal waters distinguished by their estimated reservoir temperatures. Most belong to a lower temperature group with estimated subsurface temperatures of $60^{\circ}C$, assuming chalcedony equilibrium. $60^{\circ}C$ is well below any presently feasible energy production use but possibly useful for space and process heat. The second group, including Gulf wells and Ojos Calientes, has distinctly higher reservoir temperatures, a minimum of $100^{\circ}C$ and possibly up to $180^{\circ}C$.

Using a reservoir temperature of $60^{\circ}C$, an average annual surface temperature of $20^{\circ}C$ and an estimated thermal gradient between 30° and $40^{\circ}C/km$, the depth of circulation of the lower temperature spring group is between 1000 and 1300 m (3300 to 4300 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 temperatures. Thick Cretaceous shales may also have low conductivity so that areas outside the sediment-filled portions of the basin might also have anomalously high thermal gradients.

A test of this estimate can be made from Capote Springs using its surface and estimated reservoir temperature and water chemistry. The chemistry of Capote Springs indicates that its water circulates entirely in volcanic rocks. Cretaceous shales should occur at a depth of no more than 400 m (1300 ft); water in contact with the shales should have markedly different chemistry. Circulation to a depth of 400 m in an area with a thermal gradient of $40^{o}C/km$ would raise water temperature only about $16^{o}C$, to about $36^{o}C$, almost identical to the springs discharge temperature but below the estimated reservoir temperature of $60^{o}C$. Either the estimated temperature is too high, circulation is to a greater depth, or the thermal gradient is greater than $40^{o}C/km$, possibly because of blockage of normal thermal gradient by the shales.

Estimated depth of circulation for the higher temperature waters of Ojos Calientes and Gulf wells varies considerably with assumed subsurface temperatures and thermal gradients. With a subsurface temperature of $100^{o}C$ and a thermal gradient between 30^{o} and $40^{o}C/km$, the required depth of circulation ranges between 2000 m to 2700 m (6550 ft to 8750 ft). If subsurface temperatures are as high as $180^{o}C$, depth of circulation must be between 4000 m to 5000 m (13,000 ft to 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 4300 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 with 2740 m (9000 ft) of fill indicated by gravity and seismic data along the east side of the Franklin Mountains just east of El Paso (Mattich, 1967). A well drilled by El Paso Water Utilities penetrated 1330 m (4363 ft) of bolson fill (Davis and Leggat, 1967). Total bedrock relief may greatly exceed either figure. Gravity profiles by Decker and Smithson (1975) indicate a depth of 4 km to the Hueco Bolson in the El Paso vicinity. Earth-resistivity and aeromagnetic data of Gates and Stanley (1976) show that the basin shallows 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 presumably is 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 in Hueco Bolson north of El Paso and still higher values to the west. Decker and Smithson measured heat flow values of 2.5 to 3.1 at Orogrande in New Mexico at the northern end of Hueco Bolson. Heat flow in the Texas part of Hueco Bolson may be similar, although no determinations have 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 at only 60°C. The depth of circulation required to reach this temperature should be on the same order as in Presidio Bolson, about 1000 m to 1300 m. If deeper circulation systems exist, they do not discharge to the surface. More sophisticated exploration techniques will be necessary to discover the thermal water.

The higher temperature thermal systems of 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. Definition of the Rio Grande rift probably should include this region. Obviously, geophysical studies to determine deep crustal structure are necessary to confirm this inclusion.

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 not in Presidio or Hueco basins 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 aids in interpreting 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 are in apparent agreement and are close to measured discharge temperatures (approximately 40°C). Evidently the spring waters do not cool significantly during ascent, consistent with their large discharge.

Hot springs in the Big Bend region discharge water that is heated apparently 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 0.7 to 1 km is sufficient to reach 40°C. Temperatures high enough to have potential for geothermal energy probably occur only at relatively inaccessible depths. The hot springs in the park area are used best for the enjoyment of tourists.

The area from Lobo Valley to Marfa, although it does not have either hot springs or wells, has been inferred to be a potential geothermal area because of high silica content of 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, these values would indicate subsurface temperatures around 125°C. As discussed in the section on geochemistry of thermal waters, however, the high silica content probably reflects solution of amorphous silica in volcanic rocks (tuffs and tuffaceous sediments), or basin fill derived from volcanic rocks. There is no evidence that the well waters represent anything but shallow circulation. 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. It is a deep basin with recently active normal faults (Belcher and Goetz, 1977) with at least 1000 m of normal fault displacement along the west side. The basin becomes shallow to the south. A water test well 40 km northwest of Marfa penetrated no more than 380 m (1250 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. 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 the water table is well below land surface because of heavy ground-water withdrawal for agriculture. It is conceivable that hot water comes up 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 is unrecognizable because of mixing or conductive cooling. Nevertheless, if thermal water exists in the Lobo Valley area, it is surprising that no wells have tapped any.

Heat-flow and thermal-gradient studies mostly indicate values similar to those of the Great Plains (Decker and Smithson, 1975). Kleeman (1977) did show one higher heat-flow value (2.4 or 3.0) near Van Horn. The slight ambiguity in heat flow and the similarity in setting and proximity to the Presidio graben warrant further investigation. However, according to the available evidence, the area is not a potential geothermal area.

Salt Basin, the northern extension of the Lobo Valley - Presidio series of grabens, also has recent fault scarps (Belcher and Goetz, 1977). Geophysical measurements (White and others, 1977) indicate a maximum of 760 m (2500 ft) of fill. There are no hot springs or wells. Silica content of ground water ranges from 10 to 20 mg/l. The low values are probably caused by the absence of silicic volcanic rocks in the area. Adjacent highlands are composed of carbonate sediments. For the same reasons given for the Lobo Valley area, Salt Basin is not a potential area for geothermal development.

One line of evidence contradicts this conclusion about Salt Basin. Decker and Smithson (1975) measured a heat flow of 2.0 at Cornudas, New Mexico, just across the State line in the highlands bordering the 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, however, also determined a heat-flow value of 1 north of Van Horn at the south end of the Salt Basin. Thermal gradients in oil tests in the area are low (AAPG, 1975). An explanation for the difference is not apparent unless large changes in heat flow are possible over short distances. Possibly ground-water flow has disturbed the measured heat flow, although Decker and Smithson did not consider either value anomalous.

ADDITIONAL WORK

Sufficient potential is shown by the geothermal evidence in Trans-Pecos Texas that additional work is warranted. More evaluation is needed rather than a development program. Most of the known hot springs or wells do not have much potential. Only Ojos Calientes and Gulf wells suggest high temperatures at depth, and even in these there is no evidence as yet of sufficiently high temperatures at accessible depths. The problem is that geothermal exploration, in general, is not well developed, and in Trans-Pecos Texas is still in reconnaissance stage.

The geologic background and most of the geochemical data necessary for exploration exist. Trans-Pecos Texas is exceptionally well mapped, and many geologic and geochemical data have been gathered for this report. What is lacking are geochemical data to evaluate the effects of dilution on the highest temperature thermal waters and geophysical information to define heat flow and crustal structure. Geochemical investigation of mixing of thermal and non-thermal waters, as discussed in the section on geothermometry, would be one useful step.

The greatest amount of new insight can probably be gained from geophysical studies. However, geophysical methods designed explicitly for geothermal exploration are not well developed, and some developed methods aimed at discovering buried magma chambers (Eaton, 1976) are not applicable in Texas. Nevertheless, much useful information could be gained from geophysical research. A direct geophysical exploration method might be to find additional areas of convective upwelling by measuring thermal gradients in existing wells or in shallow wells drilled specifically for geothermal exploration in and near major 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 ones that do not reach the water table, would indicate upwelling thermal waters. Some thermal waters do not reach the surface but have been discovered accidentally; thus, there should 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 would be complicated by the heterogeneity of the rocks and ground water in the areas of study (Gates and Stanley, 1976).

Although more thermal water should be discovered by these methods, there may be no 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 aimed at existing thermal convection systems. However, hot springs and wells are not ideal exploration subjects. Most hot springs of the Rio Grande area represent meteoric water which circulates to moderate depths and is heated to moderate temperatures. Depths of circulation have been estimated, but data which could define those depths accurately are unavailable. Deep circulation in low-thermal-gradient areas of the world produces hot springs, although in those areas temperatures sufficient for power production exist only at economically inaccessible depths. On the other hand, adequate temperatures could 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 structure at depth is necessary to determine whether or not sufficiently high temperatures exist and in what settings they are most shallow. Such information can be obtained in two ways: indirectly by geophysics or directly by drilling. Geophysical studies ought to precede expensive drilling.

Applicable methods include gravity, magnetic, and seismic studies to delineate subsurface structure and heat-flow studies to determine thermal gradients. For the same reasons that thermal gradients in shallow wells could be useful in detecting hidden thermal convection, they are not good indicators 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 enhance 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 and 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 reencountered. 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 are not reached 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.

Commonly 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 sites for measuring heat flow and also discovering areas of anomalously high but deep thermal gradients are possibly 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, 1976). Distinguishing this kind of high thermal gradient from that caused by convection may be difficult and even ambiguous. Yet examination of the permeability of the material penetrated in a well should provide some distinction. In general, heat flow studies both in the basins and in crystalline rocks in the adjacent highlands should at least be able to determine regional heat flow and provide some evidence whether or not the area has any real geothermal potential.

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