

CHAPTER 4 - SOUTHEASTERN HUECO AQUIFER

Location and Extent

The southeastern Hueco Bolson is separated geographically from the Hueco-Tularosa Bolson at the El Paso/Hudspeth county line. A southeast trending linear feature, the bolson extends for 55 miles from the county line to its southeastern limit at Indian Hot Springs. The bolson is bounded on the north by the Finlay, Malone, and Quitman Mountains and Diablo Plateau. The Sierra de San Ignacio, Sierra de La Amargosa, Sierra de San Jose Del Prisco, Sierra de Las Vacas, and Sierra de Carrizalillo define its southern boundary (Figure 4.1). Total surface area of the southeastern Hueco Bolson is 829 mi². Approximately 61% of its land area is in the United States.

North of the river, the floor of the bolson slopes toward the southwest, from elevations of 4,600 to 3,600 ft near the Diablo Plateau escarpment and Quitman Mountains to elevations of 3,550 to 3,300 ft along the Rio Grande. South of the river, the floor of the bolson slopes from elevations of 4,450 to 4,100 ft along mountain fronts to the Rio Grande. The Rio Grande is the only perennial river between the El Paso/Hudspeth county line and Indian Hot Springs. A few springs in the mountains provide localized flows and seeps, but most surface flows in the highlands are ephemeral and are focused at arroyos which carry water only after heavy rainfall.

Stratigraphy and Water-Bearing Characteristics

Rock and sediment types

Saturated rocks in the highlands are recharged by precipitation (Figure 4.2). The Cenozoic basin fill, in turn, is recharged partially from Cretaceous and Tertiary rocks by cross-formational flow (Kreitler and others, 1986). That the interconnected bedrock-and-basin fill aquifers form an integrated flow system requires definition of aquifer nomenclature. Herein the

term "southeastern Hueco aquifer" refers to the saturated bolson and interconnected bedrock units that flank and underlie the southeastern Hueco Bolson. Groundwater divides in the mountains and plateaus define the limits of basinward recharge areas and the geographical limits of the aquifer (Figure 4.1).

The principal hydrostratigraphic units in the southeastern Hueco aquifer are the carbonate and clastic rocks of the Cretaceous Finlay, Cox, and Bluff Mesa formations (Figure 4.3). These rocks are exposed in the highlands and lie unconformably beneath the bolson sediments (Fisher and Mullican, 1990). Data are insufficient to determine if these consolidated rocks act as a single hydrostratigraphic unit or as a series of discontinuous and poorly interconnected hydrogeologic strata (Fisher and Mullican, 1990). The extensive tectonic history of the region and intense faulting, fracturing, and folding of Cretaceous strata may suggest that the rocks act as a heterogeneous, interconnected double continuum media with one continuum representing weakly-to-strongly interconnected fractures and the other representing the porous rock matrix. Evidence of extensive karstification of Cretaceous rocks is lacking in this area although Permian rocks to the northeast, in the Dell City area, show considerable karstification in outcrop and core.

The Cenozoic basin-fill sediments, which make up the second major water-bearing unit (Mullican and Senger, 1992) consist of minor sand lenses interstratified in a matrix of clay and silty-clays. Depositional environments ranged from alluvial fans to ephemeral lakes and saline playas (Gustavson, 1990). Vertical offset by Basin and Range faults and tabular and lenticular geometries of sand, silt, and clay deposits create significant intrastratigraphic discontinuities (Fisher and Mullican, 1990).

The thickness of the basin fill decreases from as much as 8,500 ft at the El Paso/Hudspeth county line to an infinitesimal thickness where the bolson thins out near Indian Hot Springs (Collins and Raney, 1991).

Saturated bolson fill is principally of the Fort Hancock formation. Fort Hancock sediments are mostly lacustrine clay, bedded gypsum, and sand, silt, and clay from both alluvial fans and local fluvial deposits (Collins and Raney, 1991). The Camp Rice formation, a second major lithologic unit, is thin and contains little water east of the El Paso/Hudspeth county line. Camp Rice deposits were formed in alluvial fan, fluvial, and lacustrine systems and are composed of sand and gravel and minor silt and clay. They are separated from older Fort Hancock deposits by an unconformable contact as much as 2.5 m.y. old (Vanderhill, 1986).

The Quaternary alluvium and terrace deposits adjacent to the Rio Grande were formed by sediment deposition by the river. These deposits and their hydrogeologic characteristics are discussed in Chapter 5, entitled "Rio Grande aquifer." Cross-formational flow between the Rio Grande aquifer and older bolson deposits is discussed in this chapter (Figure 4.2), but other details are omitted until later.

Aquifer properties

Transmissivity values in the Cretaceous strata and bolson fill north of the Rio Grande are all relatively low (Table 4.1). Well yields do not exceed 200 gpm usually and most well yields are less than 50 gpm.

Aquifer tests performed in wells screened or open in Cretaceous strata gave transmissivity values between 0.22 and 1.50 ft²/day (Mullican and Senger, 1992). These estimates were derived from wells concentrated about the southwestern Diablo Plateau. Aquifer tests in Cretaceous rocks on the northeastern Diablo Plateau gave higher average transmissivity values (see Chapter 6), between 0.32 and 6,700 ft²/d (Kreitler and others, 1986). Wells to the northeast are located along regional flexures and may have higher transmissivity values due to greater density of fractures.

Aquifer tests in wells completed in bolson silts and sands gave transmissivity estimates between 0.43 and 94 ft²/day (Mullican and Senger, 1992). Higher trans-

missivity values are characteristic of a higher percentage of sand and gravel in the basin (Table 4.1). Lower transmissivity values are characteristic of mud-rich sediments deposited in lacustrine and playa environments. Aquifer tests in basin fill indicate relatively low transmissivity values, sufficient only for livestock and domestic use.

We could not locate aquifer test data in the Mexican part of the southeastern Hueco aquifer; few if any may exist. There are at least 100 documented wells that draw water, at least in part, from bolson silts and sands (IBWC, 1989). Many of these are irrigation wells and most are aligned marginal to or on the Rio Grande floodplain (IBWC, 1989).

Many wells are concentrated about the larger arroyos. Arroyo deposits are coarse-textured and are surfaces for precipitation recharge. Placement of wells about arroyos provides a better chance of obtaining a prolific well with good quality water. Clayey-silts are dominant in the older basin fill and the few permeable sand-and-gravel lenses are usually saturated with saline water (Geo Fimex, 1970); not a good prospect for groundwater exploration and development.

That the density of wells at distances from arroyos is spotty provides indirect evidence of lower transmissivity values in the older basin fill. More credible evidence is provided by surface electrical resistivity sounding data (Geo Fimex, 1970). Vertical electrical soundings performed in the Mexican part of the southeastern Hueco Bolson showed that aquifer resistivities are usually less than 10 ohm-m (Figure 4.4). Such low values imply limited potential for development of well fields (Dobrin, 1976; Kearey and Brooks, 1984). Yields should be sufficient in most areas only to satisfy very limited agricultural and municipal water demands. Transmissivity values for saturated bolson fill in Mexico are probably similar to estimates derived in the American part of the southeastern Hueco Bolson.

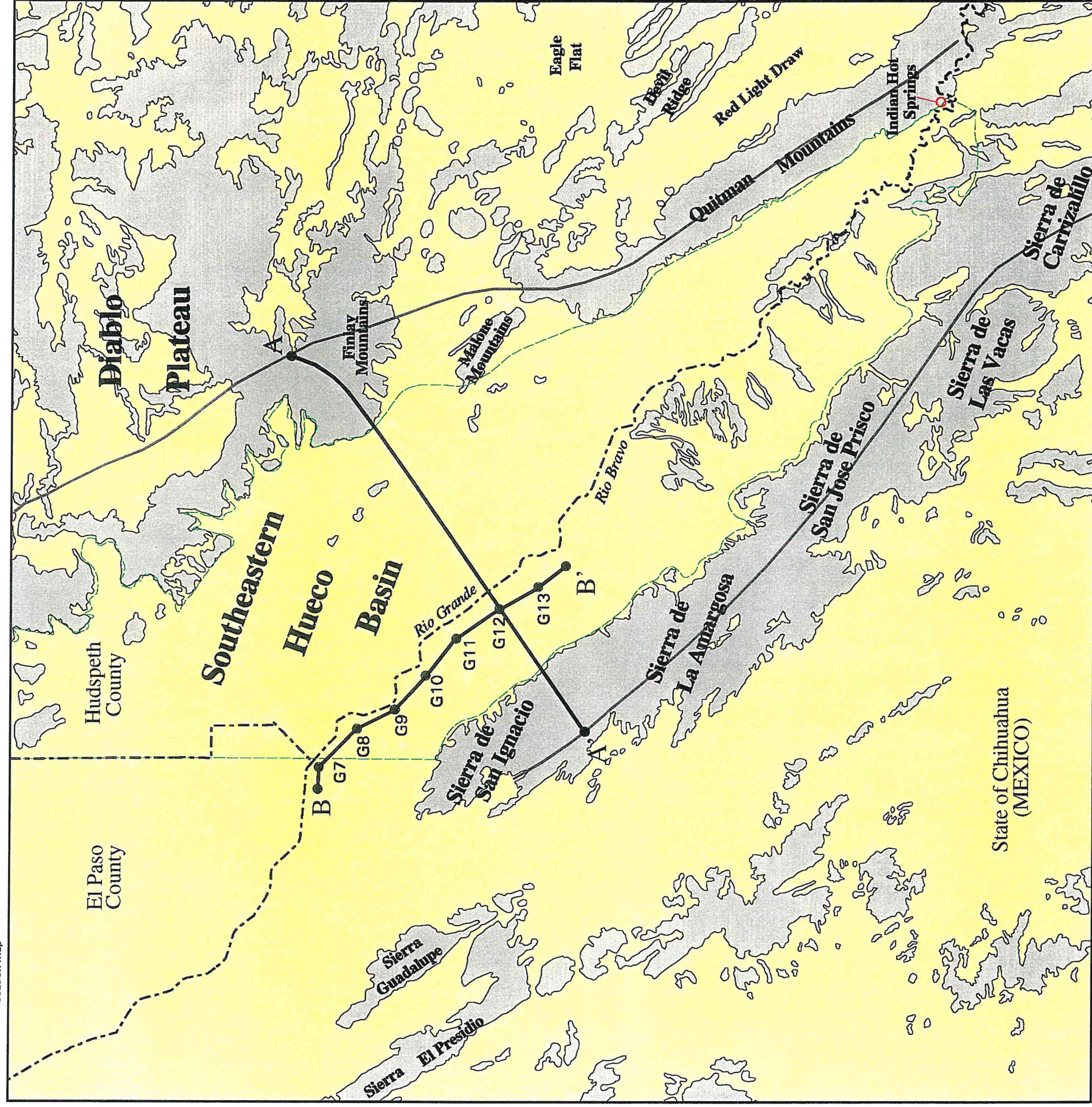
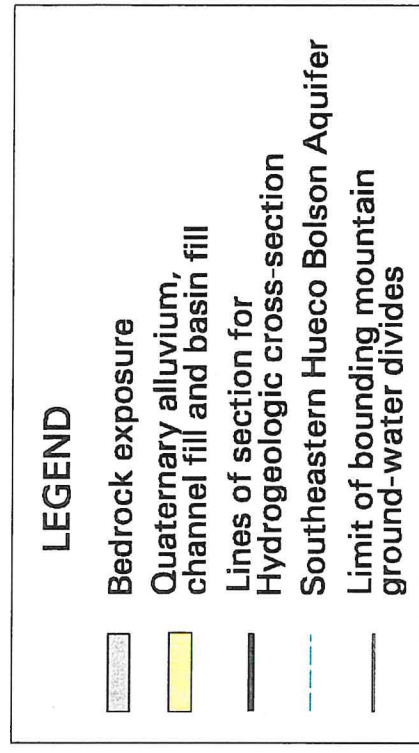
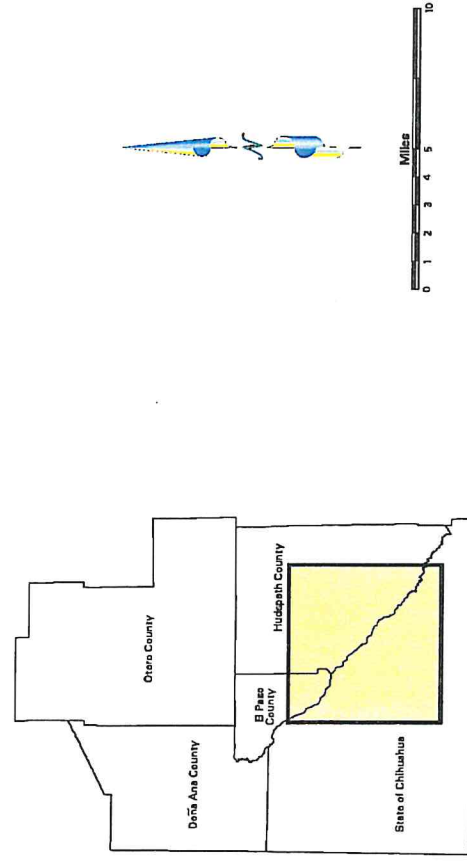


Figure 4.1. Location and extent of the southeastern Hueco Bolson aquifer. Map shows the limits of the southeastern Hueco Bolson aquifer as well as the limits of the bounding mountain ground-water divides that define the full extent of the aquifer (bolson fill and flanking highlands).

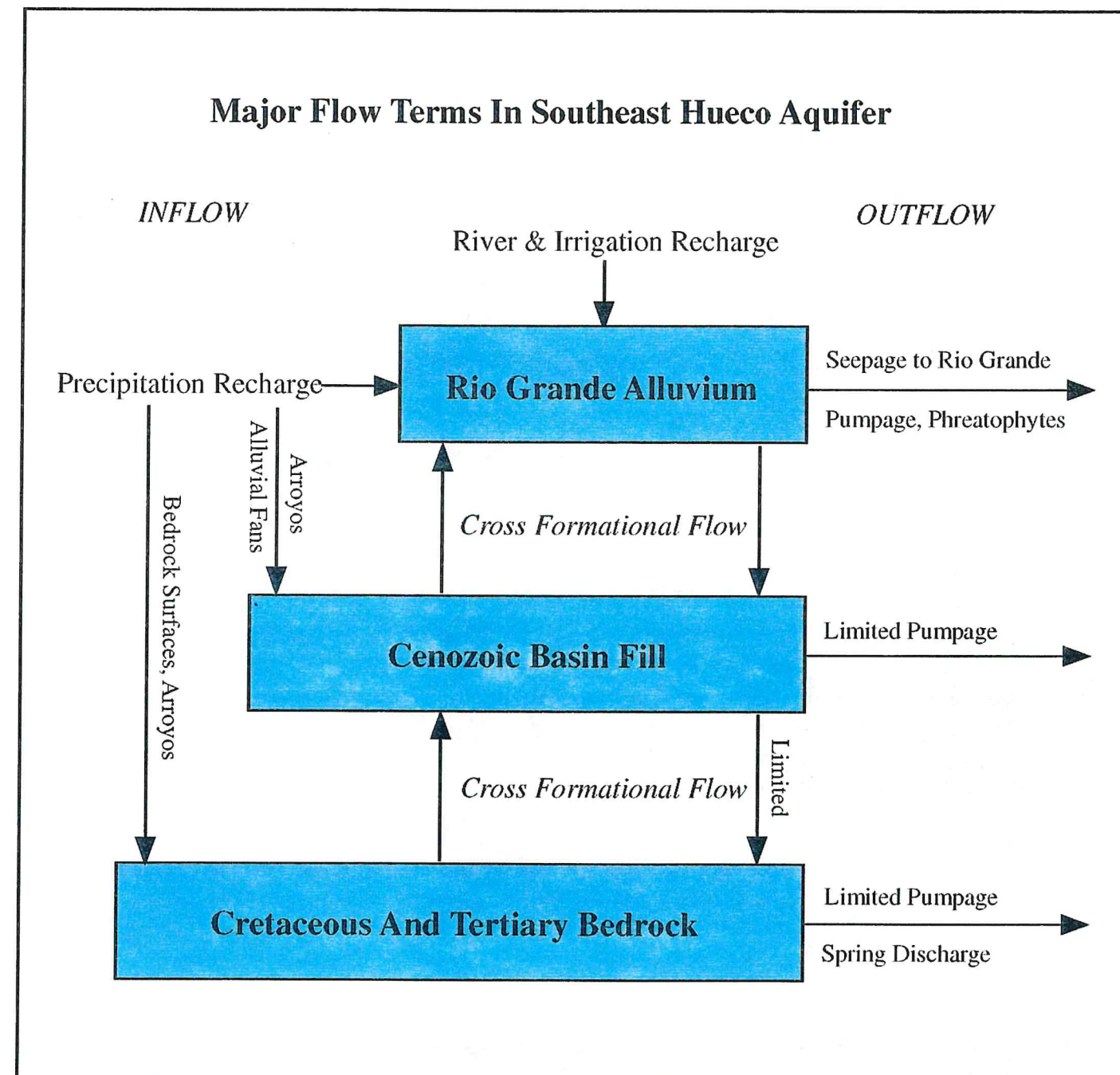


Figure 4.2. Major flow components in the southeastern Hueco aquifer.

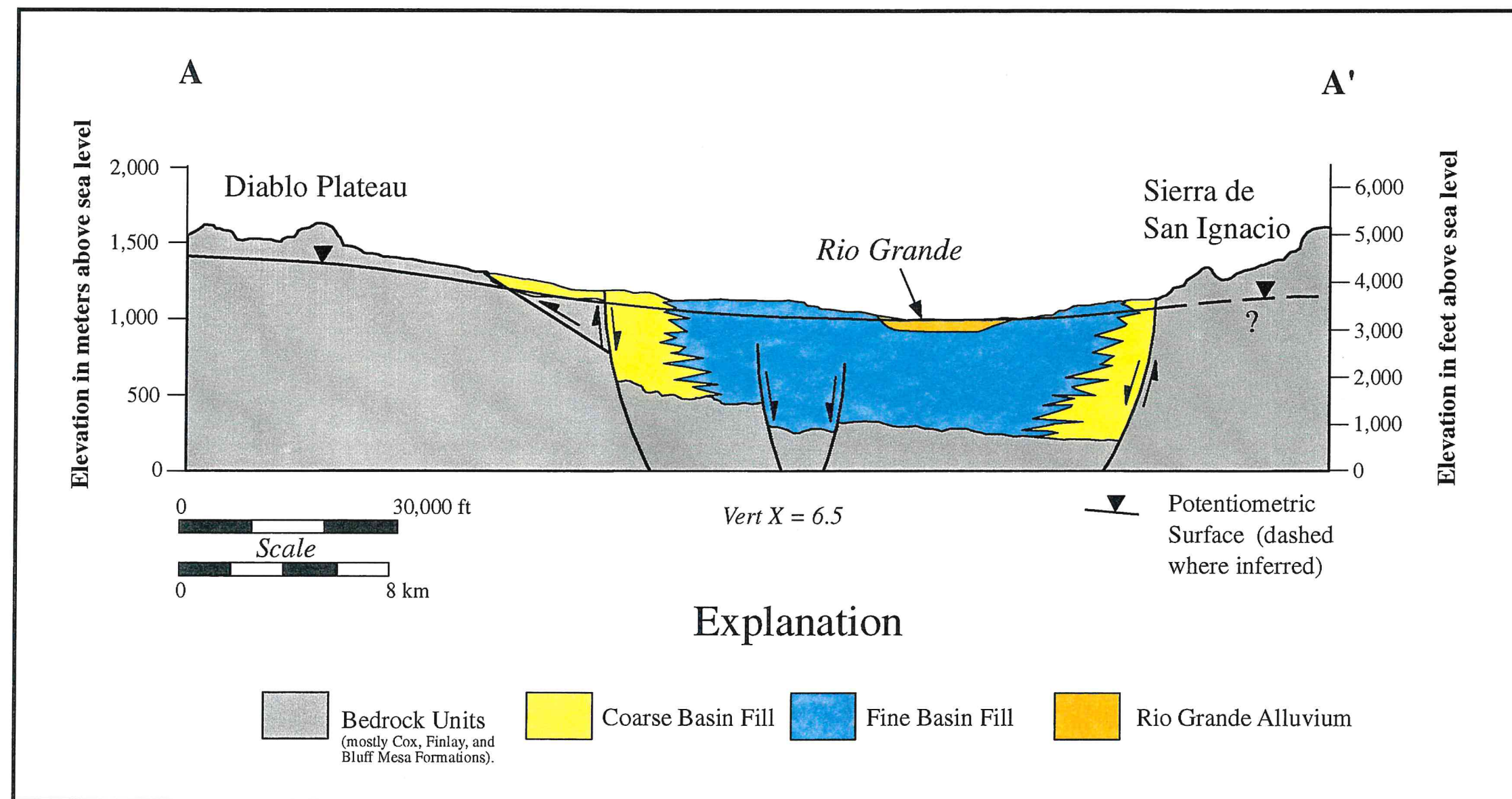


Figure 4.3. Generalized hydrogeologic cross section A - A' (line of section shown in Figure 4.1. Basin fill/bedrock contact selected from maps prepared by Collins and Raney, 1991, and from test-hole logs and geophysical logs in the Texas Water Development Board files).

Geoelectric Cross Section B - B'

Juarez Valley, Chihuahua Mexico

Explanation

G7



Geoelectric sounding station

 $8\Omega \text{ m}$

Resistivity in ohm-meters



Resistivity indicates good development potential, where saturated



Resistivity indicates poor to fair development potential



Resistivity indicates poor development potential



Resistivity and depth suggest bedrock strata

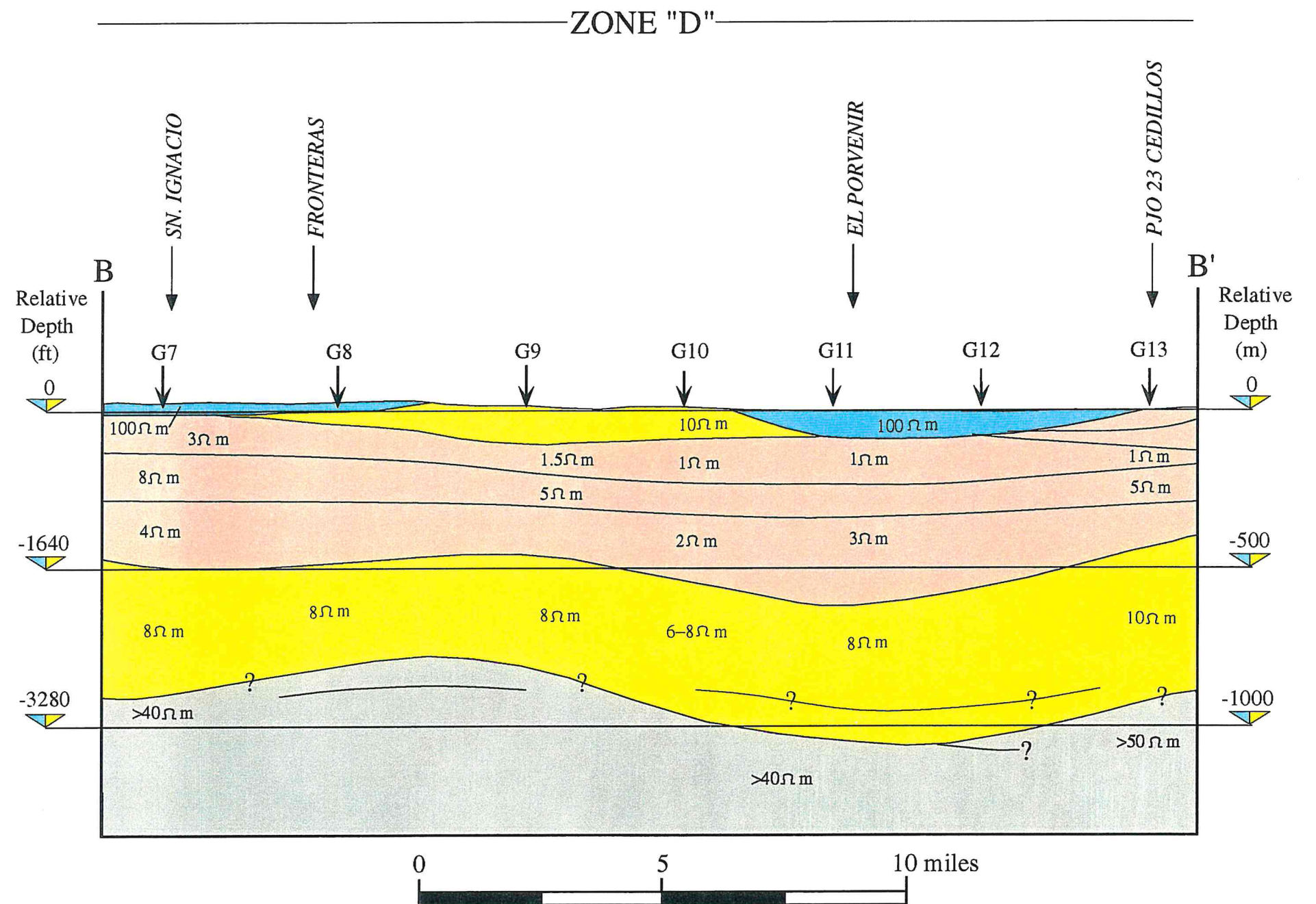


Figure 4.4. Geoelectric cross section B - B' across the southeastern Hueco aquifer, northern Chihuahua, Mexico. The map shows interpreted average real resistivities (modified from Geo Fimex, 1970; line of section shown on Figure 4.1).

**Transmissivity Results From Aquifer Tests
In The Southeastern Hueco Aquifer**

State Well Number	Water Bearing Strata	Transmissivity (ft ² /day)	Aquifer Test Method
48-35-702	Cretaceous	0.22	Theis Recovery (semilog)
48-43-101	Cretaceous	0.60	Theis Recovery (semilog)
48-43-501	Cretaceous	1.50	Theis Recovery (semilog)
48-34-903	Hueco Bolson	94.0	Theis Recovery (semilog)
48-42-501	Hueco Bolson	3.60	Theis Recovery (semilog)
48-34-802	Hueco Bolson	0.43	Theis Recovery (semilog)
48-42-101	Hueco Bolson	3.60	Ferris and Knowles (slug)
48-35-701	Hueco Bolson	7.83	Ferris and Knowles (slug)

**Table 4.1 Transmissivity values derived from aquifer tests in the southeastern Hueco Aquifer
(data from Mullican and Senger, 1992)**

Potentiometric Surface Map and Water Levels

North of the Rio Grande, the regional potentiometric surface map shows high hydraulic heads and ground-water divides along the Diablo Plateau, Finlay mountains, and Quitman mountains (Figure 4.5). Areas of high hydraulic head in the mountains and plateaus define focal points of recharge in the southeastern Hueco aquifer. Hydraulic head gradients in the Cretaceous and other bedrock strata are as much as 0.07 along ground-water divides and are as little as 0.04 along mountain fronts. Hydraulic head gradients in the bolson fill are about 0.008. Steeper hydraulic head gradients along mountain fronts are due to; (1) higher recharge rates in the mountains and along mountain fronts, and (2) average permeabilities of bedrock strata that are lower than average permeabilities of bolson fill (Table 4.1). By Darcy's law a high permeability material, allowing the same quantity of water to be transmitted as a low permeability material, has a smaller loss of hydraulic head along a flowpath vector.

South of the Rio Grande, the potentiometric surface slopes to the river from high topographic elevations along mountain fronts. Peak elevations of the mountain ranges probably mark the location of ground-water divides. Springs flow at high elevations from the mountains in Mexico. These probably discharge from locally perched flow systems that do not define hydraulic head in the zone of regional saturation. Data are not adequate to define regional hydraulic heads beneath the mountains in Mexico.

Depth to ground water in the southeastern Hueco aquifer is variable. The depths measured to the regional water table in Cretaceous rocks on the U.S. side varied from 76 to 627 ft, except at Thaxton Spring where ground water flows at land surface at the Diablo Plateau escarpment. Depth to ground water in the basin fill was measured between 93 and 479 ft (Mullican and Senger, 1992). Depth to ground water

beneath mountain ranges that bound the southeastern Hueco Bolson in Mexico is unknown.

The southeastern Hueco aquifer can almost be considered undeveloped, especially north of the Rio Grande. Low capacity domestic and livestock wells are used to satisfy the needs of the local population and livestock industry. Water-level data in time series are not available in the southeastern Hueco aquifer, except in wells in the Rio Grande alluvium (see Chapter 5). Water-levels over the last few decades probably haven't changed significantly in the aquifer.

Ground-Water Availability

Estimates of the quantity of fresh and slightly saline water for the southeastern Hueco aquifer cannot be derived because lithologic, geophysical, and water quality data are not sufficient to permit analysis. Total quantity of fresh, slightly saline, and moderately-to-highly saline ground water in the bolson fill is estimated by calculating the volume of saturated fill between the water table and bedrock surface and by multiplying this volume by 0.22, an average specific yield value for the bolson fill. Total amount of water stored in the bolson fill is about 76 million acre-ft. Of this amount, 42 million acre-ft are stored in the American part of the southeastern Hueco Bolson and 34 million acre-ft are stored in the Mexican part of the bolson.

The estimates for Mexico are less reliable because the water table map and bedrock configuration map were developed with fewer data. The analysis ignores the amount of water held in artesian storage (negligible compared to the drainage porosity) and does not include water stored in Cretaceous and Tertiary bedrock aquifers.

Recharge Areas

Tritium (^3H) and carbon-14 (^{14}C) data provide clues to the distribution of recharge and relative ages of ground water. Pre-1950 values for tritium in northern hemisphere precipitation were about 5 tritium units

(TU), where one TU is equal to one atom of ^3H in 10^{18} atoms of hydrogen. Tritium has a half life of 12.3 years (Mazor, 1991) and ^3H values less than about 0.5 TU usually indicate ground waters recharged before 1952, provided that extensive dilution by older ground waters has not occurred (Mazor, 1991). Tritium in northern hemisphere precipitation increased to more than 2,000 TU as a result of above-ground testing programs for nuclear weapons in the 1950's and 1960's. Tritium has decreased to near-background levels in recent years (Mazor, 1991).

With a much longer half-life of 5,730 years, ^{14}C is a radiometric dating tool that may be used to date ground water to 30,000 years or more provided appropriate adjustments are made to account for factors other than radiometric decay (i.e., mixing and/or isotope exchange) that alter the original isotopic signatures (Fontes and Garnier, 1979; Mook, 1980). The results of ^{14}C analyses are usually reported in units of percent modern carbon (pmc). In the ideal case, an initial ^{14}C concentration of 100 pmc in ground water, a parcel of 50 pmc ground water that has not undergone mixing or dilution is 5,730 years old. The carbon cycle is seldom ideal and rock in carbonate terrains may have an initial value of $60 \pm 5\%$ pmc due to rock-water and soil-water interactions in the vadose zone (Mazor, 1991). Initial values greater than 100 pmc in non-carbonate rocks (Fontes and Garnier, 1979; Mazor, 1991) are due to enhanced ^{14}C production associated with above-ground nuclear weapons testing programs of the 1950's and 1960's. Such irregularities require identification of the stable carbon-isotope signature of rock, among other data, for correction factors to be applied.

We use published radioisotope data to identify recharge areas and relative ground-water residence times (Fisher and Mullican, 1990). We do not attempt to derive adjusted ^{14}C ages because of uncertainty regarding many factors known to influence the chemistry of dissolved inorganic carbon in ground water (Fontes and Garnier, 1979; Mook, 1980). The complex tectonic history of the region and juxtaposition of

rocks of different lithologies and ages make application of correction techniques to the ground water in the area difficult. The absolute values of ^{14}C are used to provide information about the relative differences in ground-water ages within the basin and as an indicator of the areas where recharge occurs.

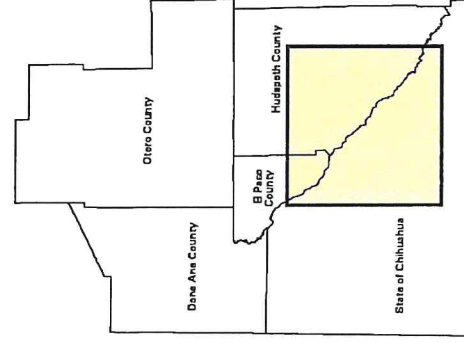
Radioisotope data and water-level information indicate that precipitation recharge to the basins occurs primarily within the upper mountains and plateaus, and across the broad alluvial fans that border mountain fronts (Figure 4.6). The data indicate there is a clearly defined trend toward lower ^{14}C and tritium within basin floor areas in the southeastern Hueco aquifer. Plots of ^3H and ^{14}C versus surface elevation suggest mixing of waters of different ages within the basin (Figure 4.6). Low percentages of modern carbon and tritium activities above 0.5 TU units in alluvial fans imply mixing of young and old ground water at the fans. Ground water moving downgradient from the mountains may converge at the fans and mix with recent precipitation recharge on the fans.

Major arroyos dissect the alluvial fans and bolson surfaces in the southeastern Hueco aquifer, sometimes penetrating the underlying Cretaceous rocks. These arroyos, sometimes over 200 ft wide, convey substantial quantities of runoff during episodic wet years and act as a third pathway for focused recharge downgradient from the principal recharge surfaces in the Diablo Plateau and alluvial fans.

Discharge Areas

Ground water is lost from the southeastern Hueco aquifer by spring discharge and by cross-formational leakage to the Rio Grande alluvium (Figure 4.2). Well discharge accounts for limited discharge from the basin except in the Rio Grande alluvium where irrigated agriculture is common.

Two hot springs (Indian Hot Springs and Red Bull Spring) are located near the southwestern end of the Quitman mountains along the trace of the Caballo



Location Map

LEGEND

- Bedrock exposure
- Quaternary alluvium, channel fill and basin fill
- Well control points
- Water-level contour
- Dashed where inferred
- NOTE:** Contour interval varies
Contour lines in feet

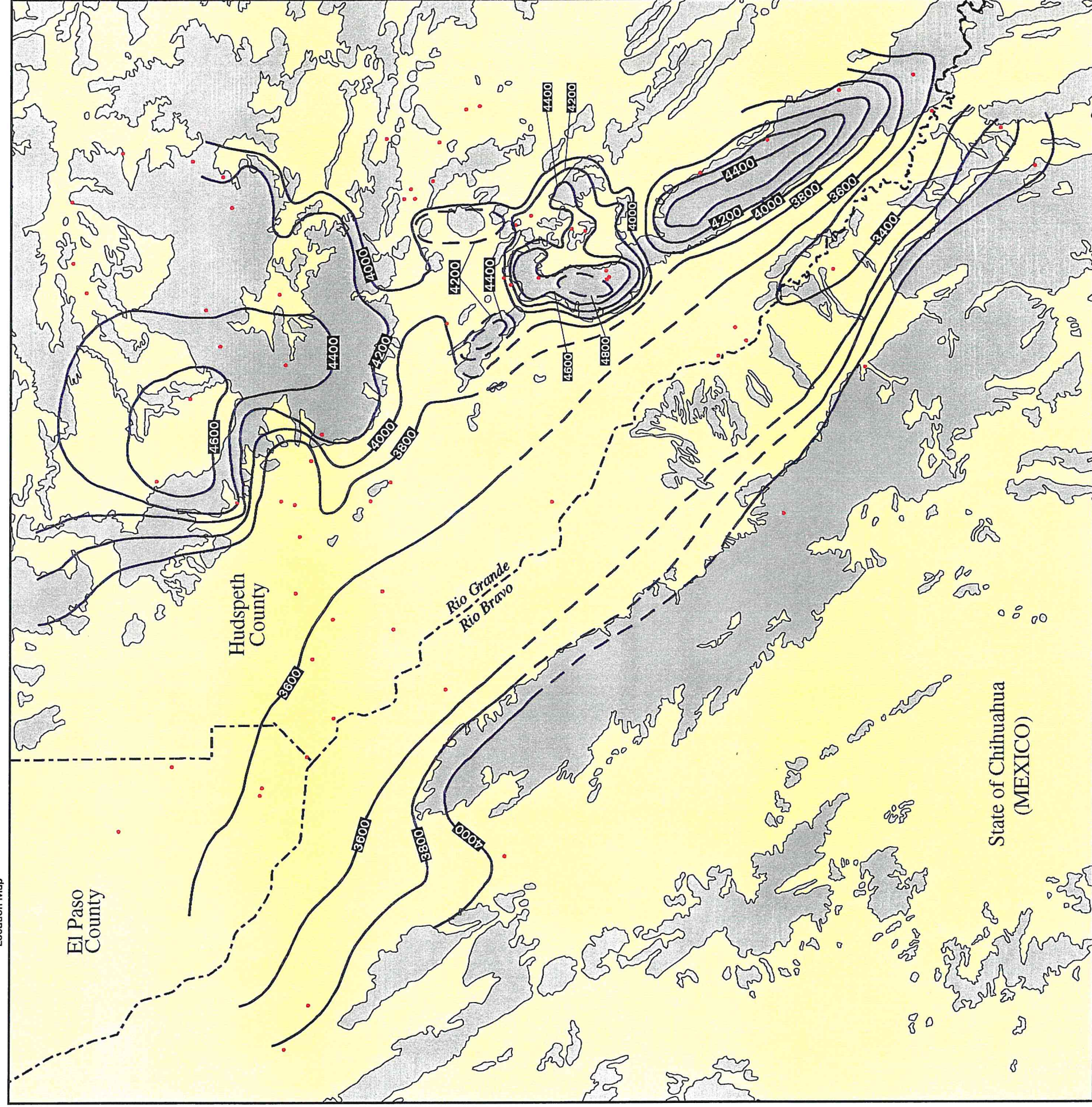


Figure 4.5. Regional potentiometric surface map for the southeastern Hueco aquifer and surrounding mountains and plateaus. (source of data, Fisher and Mullican, 1990; Texas Water Development Board; Comision Nacional del Agua; Instituto Nacional de Estadistica, Geografia e Informatica).

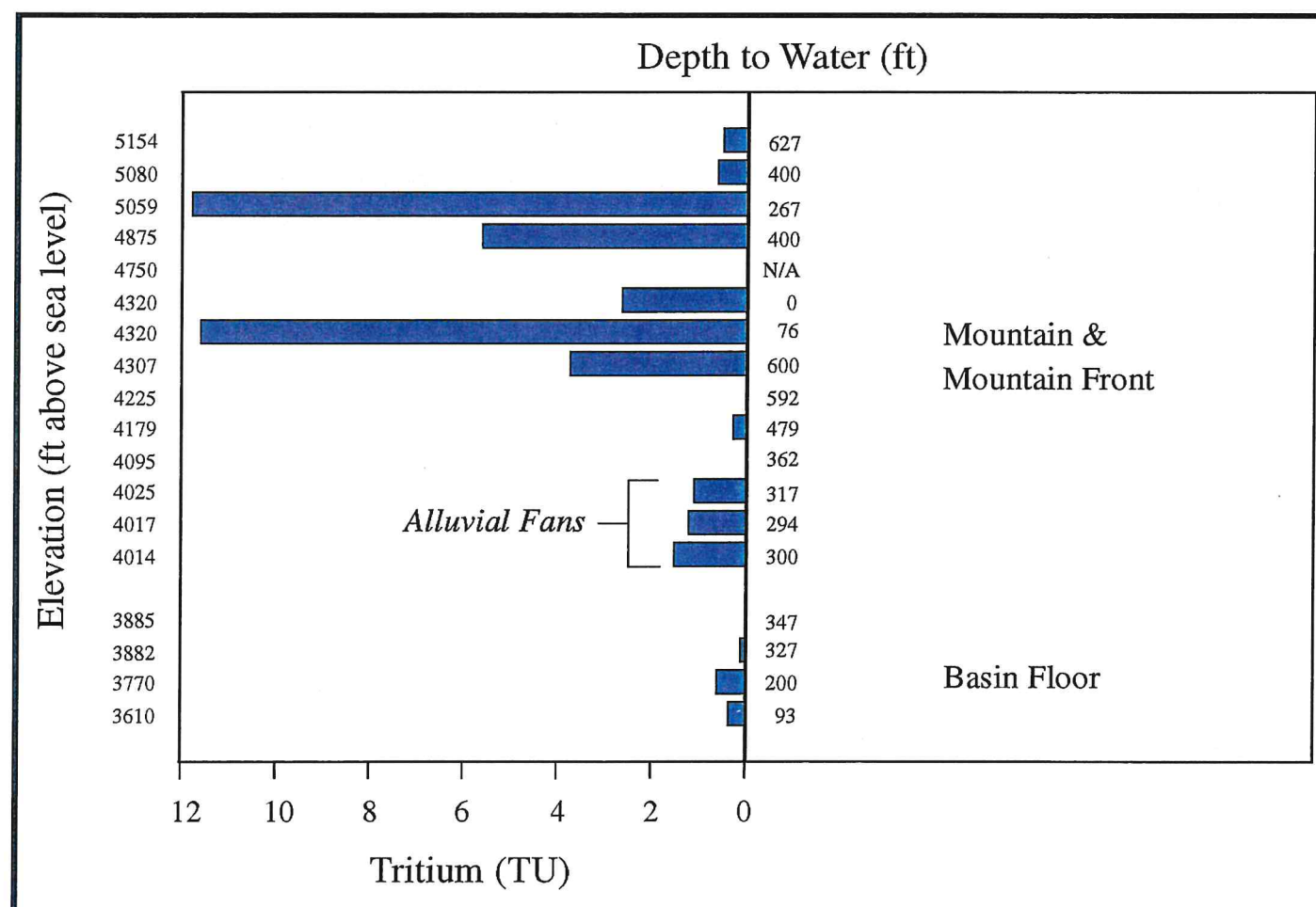


Figure 4.6a. Tritium activities (TU) in highland (mountain and mountain front) and lowland (basin floor) areas of the southeastern Hueco aquifer indicate that precipitation recharge occurs primarily within the upper mountains and plateaus (source of data, Fisher and Mullican, 1990).

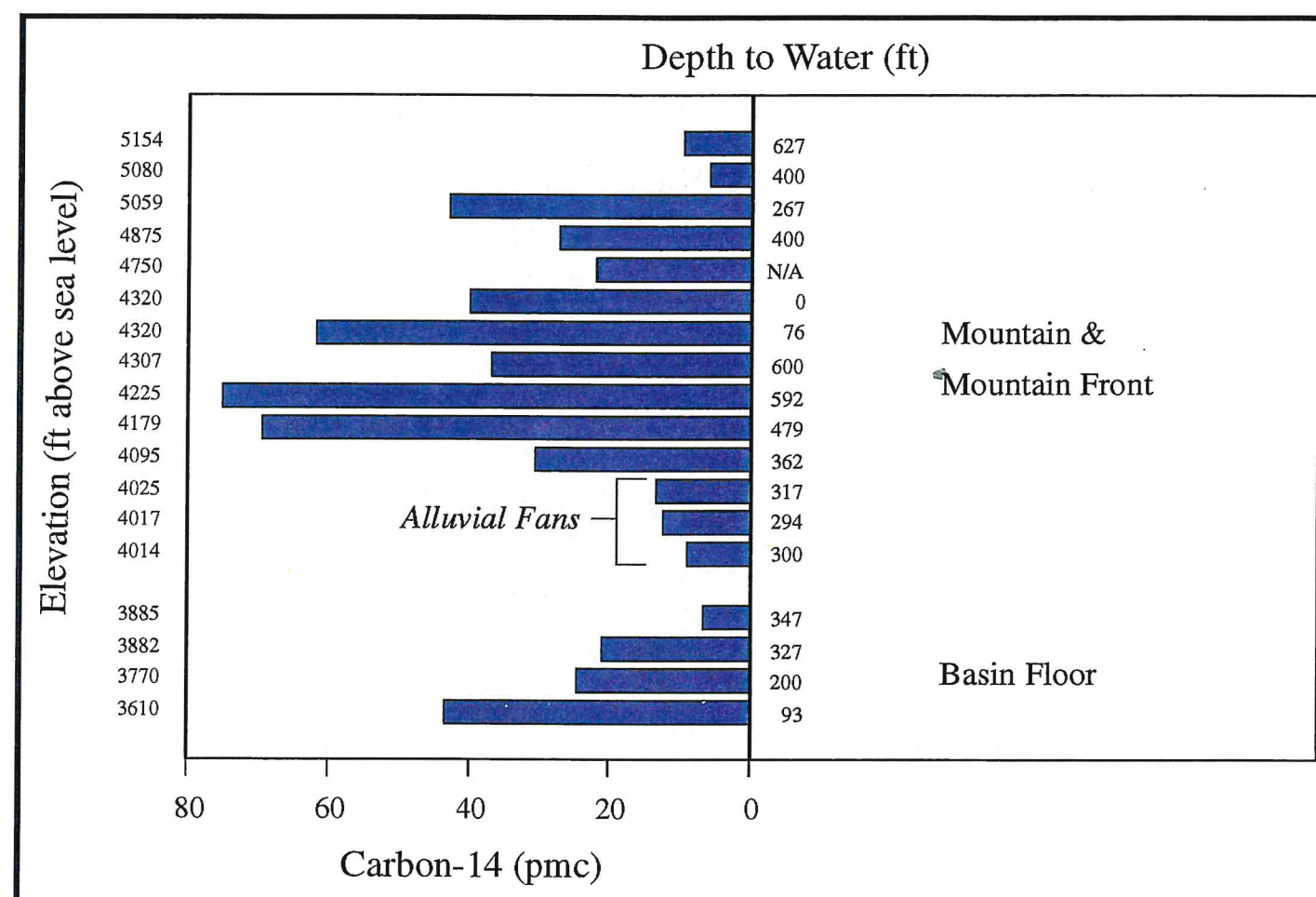


Figure 4.6b. Percentages of carbon-14 in highland (mountain and mountain front) and lowland (basin floor) areas of the southeastern Hueco aquifer indicate that precipitation recharge occurs primarily within the upper mountains and plateaus (source of data, Fisher and Mullican, 1990).

Fault, a northwest-trending normal fault that separates the Quitman Mountains and the southeastern Hueco Bolson. Water temperatures in the hot springs vary from 27 to 50°C (Darling and other, 1994). Several small, cooler springs are located in the Mexican part of the aquifer, mostly along faults in the highlands. Temperature in these springs vary from 22 to 25°C.

Cool springs are usually discharge areas for local ground-water flow systems (Mifflin, 1988). Temperature and discharge in springs that issue water from local flow systems are susceptible to large fluctuations due to changes in atmospheric temperature and moisture. Hot springs are usually discharge areas for regional flow systems. Temperature and discharge from hot springs in regional flow systems tend to vary only slightly from year to year (Mifflin, 1968; Winograd and Thordarson, 1975).

Most of the ground-water discharge from the southeastern Hueco aquifer occurs by cross-formational leakage to the Rio Grande alluvium (Figure 4.2). This ground water, in turn, eventually leaks into the Rio Grande or is discharged by well pumping from the alluvium. Consumptive use by phreatophytes accounts for another component of discharge (Figure 4.2). Salt cedar are densely thicketed in the Rio Grande alluvium below Fort Quitman and account for significant ground-water consumption along this stream reach.

Water Quality

General hydrochemistry

A stiff diagram (Plate 1) illustrates general water quality in the southeastern Hueco aquifer. Total dissolved solids are typically greater than 1,000 mg/L in the mountains, increasing to as much as 4,000 mg/L in the bolson. Ground water chemistry in the Rio Grande aquifer is discussed independently in Chapter 5.

The hydrochemical facies (Back, 1966) of southeastern Hueco aquifer ground waters on the American side

of the study area (Figure 4.7) varies from Ca-Mg-HCO₃ and Na-SO₄ along the Diablo Plateau to Na-SO₄-Cl beneath the floor of the basin. South of the Rio Grande, ground waters vary from Ca-Mg-HCO₃ beneath the mountain ranges to Ca-Mg-SO₄-Cl waters beneath the basin floor (Figure 4.7). Typically these ground waters have TDS that vary between 1,000 and 3,500 TDS. Indian Hot Springs is an exception; Na-Cl water with TDS higher than 7,000 mg/L discharges from Cretaceous carbonate and clastic rocks at the hot springs.

Origin of solutes

Hydrochemical signatures within the southeastern Hueco aquifer are controlled by the solubilities of aquifer materials, cation exchange, and simple mixing. The concentrations of Ca, Mg, and Na, for example, are controlled by weathering of carbonates, gypsum, and halite and by exchange of the divalent cations Ca and Mg for the monovalent cation Na (Fisher and Mullican, 1990; Darling and others, 1994). Slightly saline ground waters are dominated by the dissolution of gypsum whereas dilute ground waters are dominated by the dissolution of calcite and dolomite. Halite dissolution is indicated for moderately saline water at Indian Hot Springs by a plot of the molar ratio of (Na/Cl) against the concentration of Cl (Figure 4.8). (Na/Cl) molar ratios approach a value near 1.0 at Indian Hot Springs and usually range from 2.0 to as much 7.0 in more dilute waters (Figure 4.8). Ratios near 1.0 are the result of the release of equimolar concentrations of Na and Cl by dissolution of halite.

The molar ratio of chloride to bromide (Cl/Br) offers further support for halite dissolution as a significant control on hydrochemical compositions at Indian Hot Springs (Figure 4.9). These ratios are less than 750 in the southeastern Hueco aquifer and more than 3,000 at Indian Hot Springs. Chloride-bromide (Cl/Br) ratios reflect the origin of water as marine (~300 - 650), as a second-cycle solution of marine salt (>1,000), or as a residual brine from precipitation of halite (< 250;

Holser, 1979; Darling and others, 1994). The (Cl/Br) ratio of sea water remains nearly constant during evaporation up to the concentration at which halite precipitates (Holser, 1979; Drever, 1988). The larger bromide ion is preferentially excluded from the halite lattice structure during precipitation, and residual brine is consequently enriched in Br relative to Cl ([Cl/Br] ratio decreases) while halite is deficient in Br (Darling and others, 1994). The (Cl/Br) ratios of circulating meteoric ground water can increase by several factors as large masses of halite are dissolved, as indicated at Indian Hot Springs.

Within SO₄ - dominated waters, the predominant source of excess sodium is cation exchange with Ca the primary exchangeable divalent cation (Fisher and Mullican, 1990). Within HCO₃ waters, the dissolution of carbonate rocks provides most of the Ca required to drive the exchange. Subtracting the molar concentration of Cl from Na yields an estimate of the mass of Na attributable to the dissolution of halite (Na-Cl). A plot of (Na-Cl) against SO₄ produces a trend with a slope of approximately 2.0 in ground water north of the Rio Grande, a ratio expected in a 2 to 1 exchange of Ca for Na (Figure 4.10).

In summary, the origin of dilute and slightly saline ground waters in the southeastern Hueco aquifer corresponds to a set of common geochemical reactions, including dissolution of calcite, dolomite, gypsum, and simple cation exchange (Fisher and Mullican, 1990). These simple reactions are explained by the predominance of carbonate, evaporite, and clay minerals in Cretaceous rocks and basin fill, and by the tendency for exchange of bound sodium for calcium in solution, the dominant ion exchange reaction in dilute waters (Drever, 1988).

The geochemical signature of halite dissolution on the waters at Indian Hot Springs is more problematical, as no large halite deposits are found north of the Rio Grande. Halite deposits are ubiquitous in the area south of the Rio Grande however. El Cuervo Bolson,

located 25 miles southeast of Indian Hot Springs, is underlain by thick evaporite assemblages consisting of halite, gypsum, and anhydrite. This bolson aquifer is connected to Indian Hot Springs by a series of northwest trending faults and fractures in the Texas lineament and may be a source of warm and moderately saline ground water that discharges at the springs (Hibbs and Jones, 1996).

Ground-Water Movement

Numerical flow modeling

An important goal of this project is to identify areas where ground water moves, or may move across the international boundary. Transboundary ground-water flow is a sensitive international issues because the southeastern Hueco aquifer is adjacent to northwest Eagle Flat, the host basin for disposal of interstate sludge. Northwest Eagle Flat is also the proposed basin for disposal of low-level radioactive waste (Darling and others, 1994).

Ground-water movement across the international border could carry contaminants from one country to the other. A numerical profile model and pathline simulator are devised to predict potential transboundary ground-water flow. The model is used to predict ground-water residence times, ground-water pathlines, and the recharge rate that will bring the aquifer to flow capacity

Model orientation and design

The two-dimensional, cross-sectional model is developed along a 30 mile transboundary flowline oriented between the Diablo Plateau and the Sierra de San Ignacio (Figures 4.1 and 4.3). The steady-state model consists of 30 layers and 92 columns (Figure 4.11). Each of the 2,760 cells is 150 ft high by 1716 ft long. Boundary conditions were selected to correspond closely to actual hydrologic boundaries (Figure 4.11). A no-flow boundary was established at a depth of 100 ft below sea level at the brine/brackish water interface.

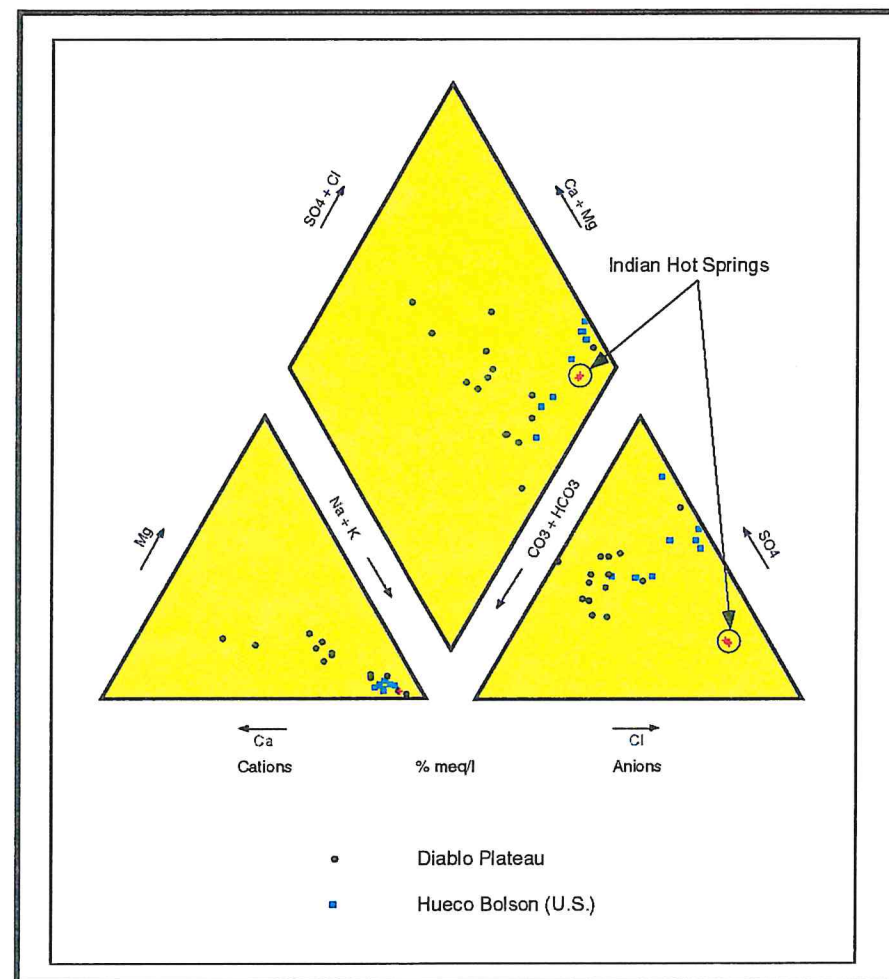


Figure 4.7a. Hydrochemical Piper plot for the bedrock (mountain and plateau) strata, bolson strata, and Indian Hot Springs in the U.S. part of the southeastern Hueco aquifer. Piper plot indicates distinct hydrochemical types for these water bearing strata in the southeastern Hueco aquifer (source of data, Fisher and Mullican, 1990; Texas Water Development Board).

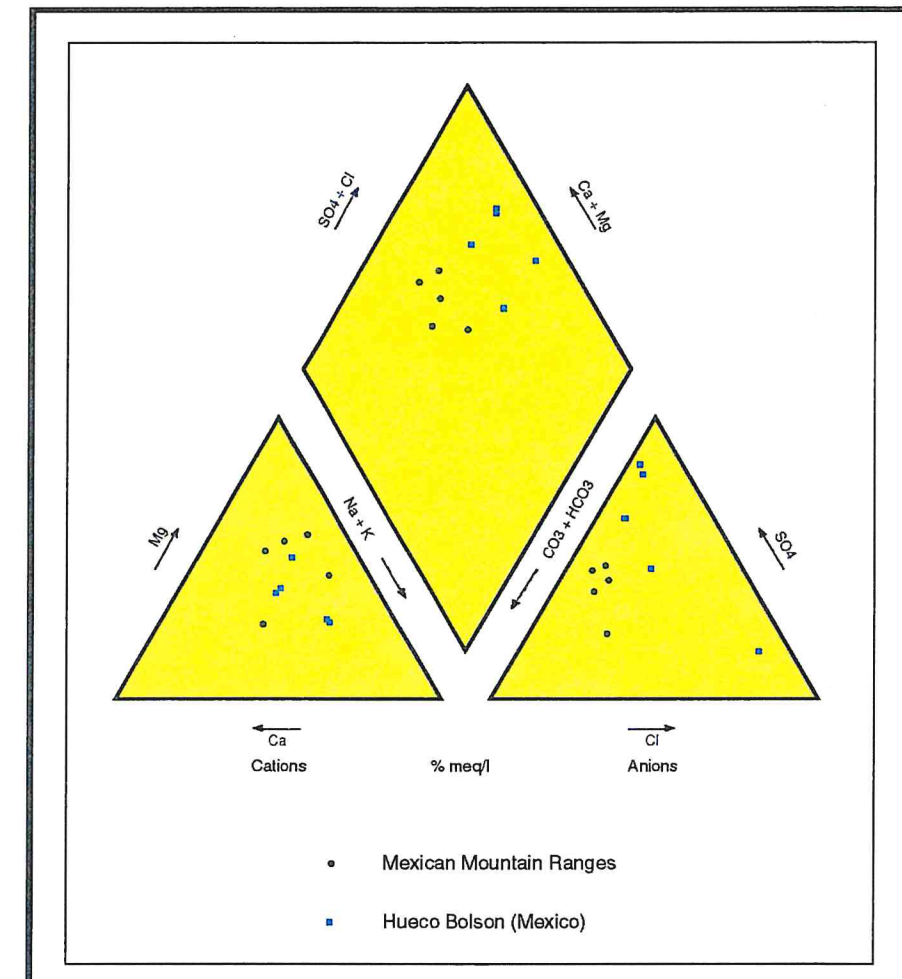


Figure 4.7b. Hydrochemical Piper plot for the bedrock (mountain) strata and bolson strata in the Mexican part of the southeastern Hueco aquifer. Piper plot indicates distinct hydrochemical types for these water bearing strata in the southeastern Hueco aquifer (source of data, Comision Nacional Del Agua; Instituto Nacional de Estadistica, Geografia e Informatica).

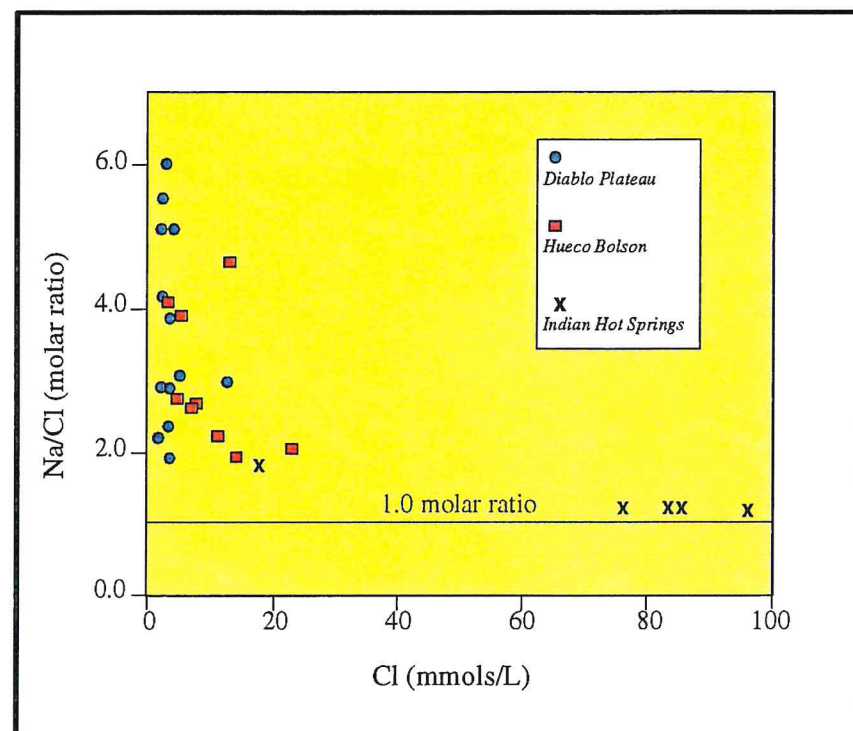


Figure 4.8a. Scatter plot showing (Na/Cl) molar ratios vs molar Cl for samples collected from bedrock strata, bolson strata, and Indian Hot Springs, U.S. (source of data, Fisher and Mullican, 1990; Darling and others, 1994; Texas Water Development Board).

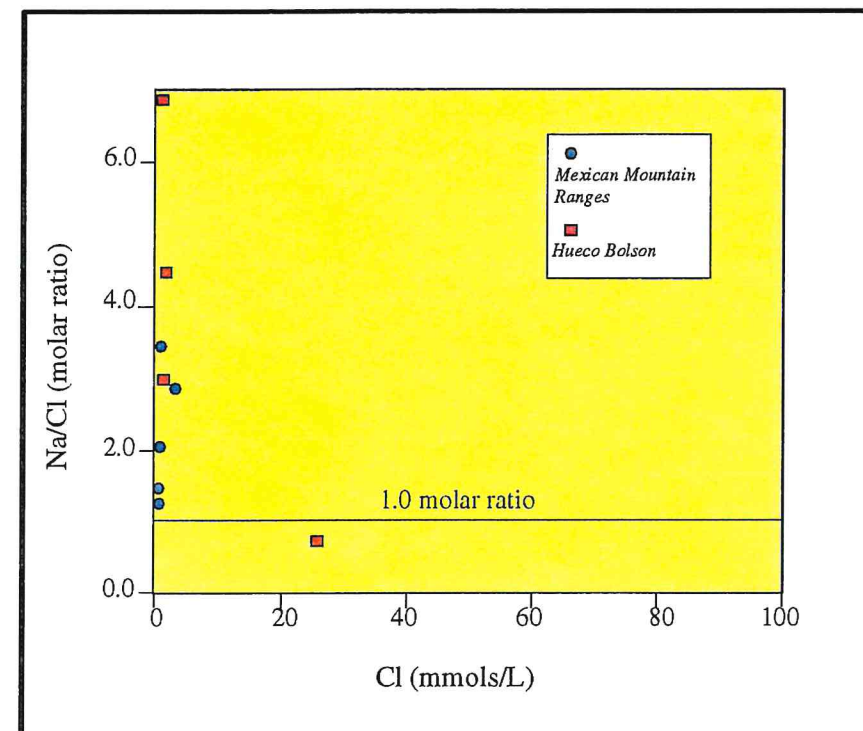


Figure 4.8b. Scatter plot showing (Na/Cl) molar ratios vs molar Cl for samples collected from bedrock strata and bolson strata, Mexico (source of data, Comision Nacional Del Agua; Instituto Nacional de Estadistica, Geografia e Informatica).

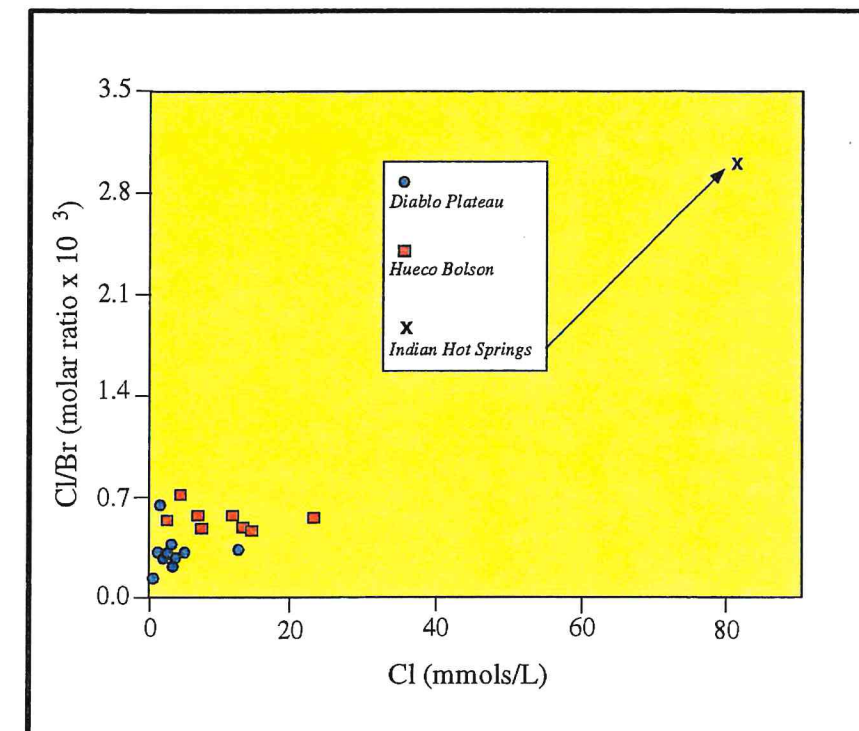


Figure 4.9. Scatter plot showing (Cl/Br) molar ratios vs molar Cl for samples collected from bedrock strata, bolson strata, and Indian Hot Springs, U.S. (source of data, Fisher and Mullican, 1990; Darling and others, 1994; Texas Water Development Board).

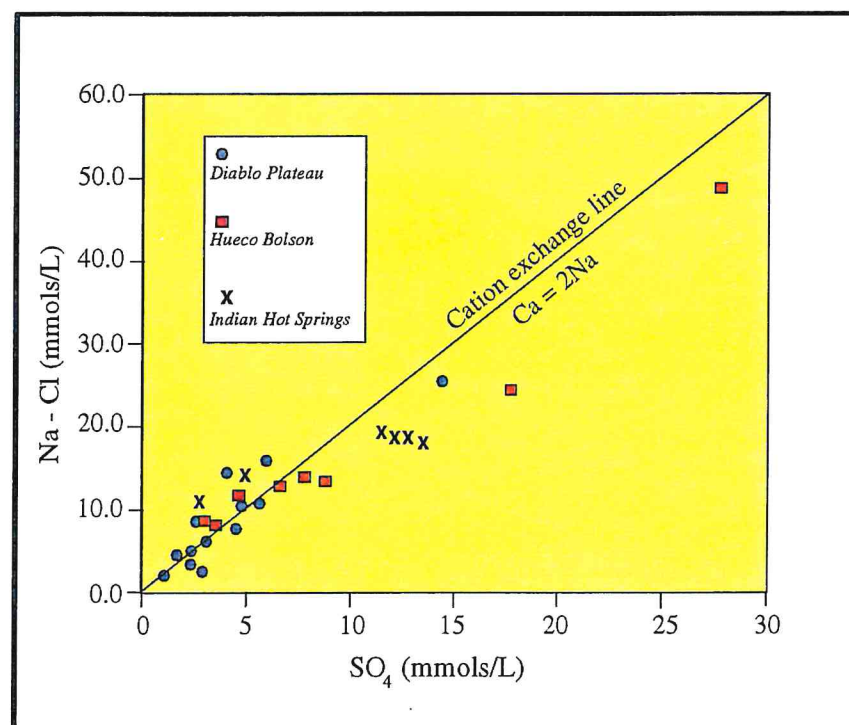


Figure 4.10a. Scatter plot showing (Na - Cl) molar quantity vs molar SO_4 for samples collected from bedrock strata, bolson strata, and Indian Hot Springs, U.S. (source of data, Fisher and Mullican, 1990; Darling and others, 1994; Texas Water Development Board).

Figures 4.8 through 4.10 show a series of scatter plots for ions in the ground water of the southeastern Hueco aquifer. Plots shown for the saturated basin fill in both the United States (Figures 4.8a, 4.9, and 4.10a) and Mexico (Figures 4.8b and 4.10b), and for the bedrock aquifers that are located in the flanking highlands. Indian Hot Springs waters plot distinct from other ground waters represented in these plots.

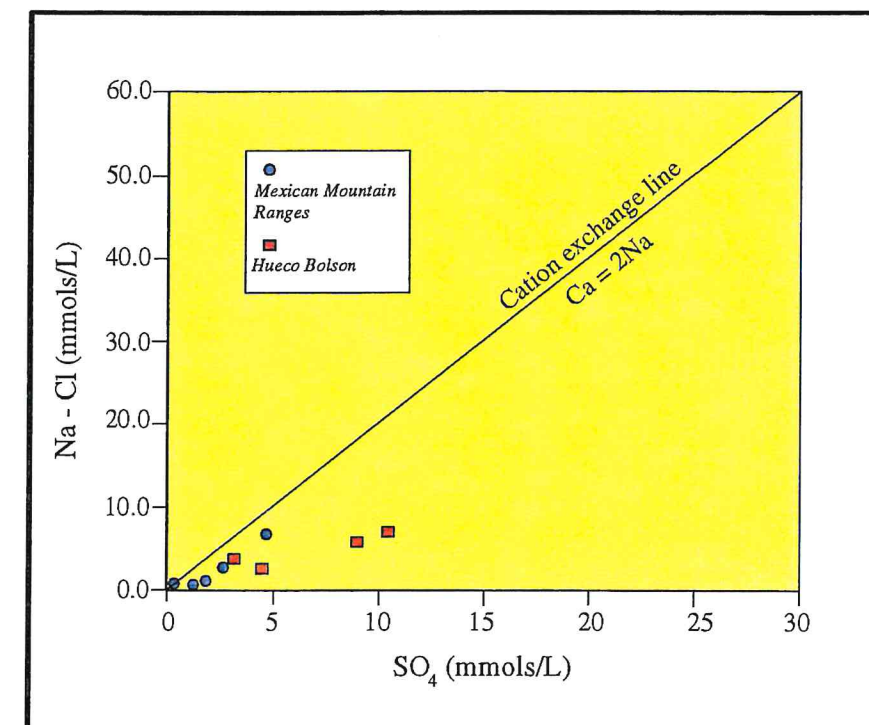


Figure 4.10b. Scatter plot showing (Na - Cl) molar quantity vs molar SO_4 for samples collected from bedrock strata and bolson strata, Mexico (source of data, Comision Nacional Del Agua; Instituto Nacional de Estadistica, Geografia e Informatica).

Model Boundary Conditions & Permeability Zones

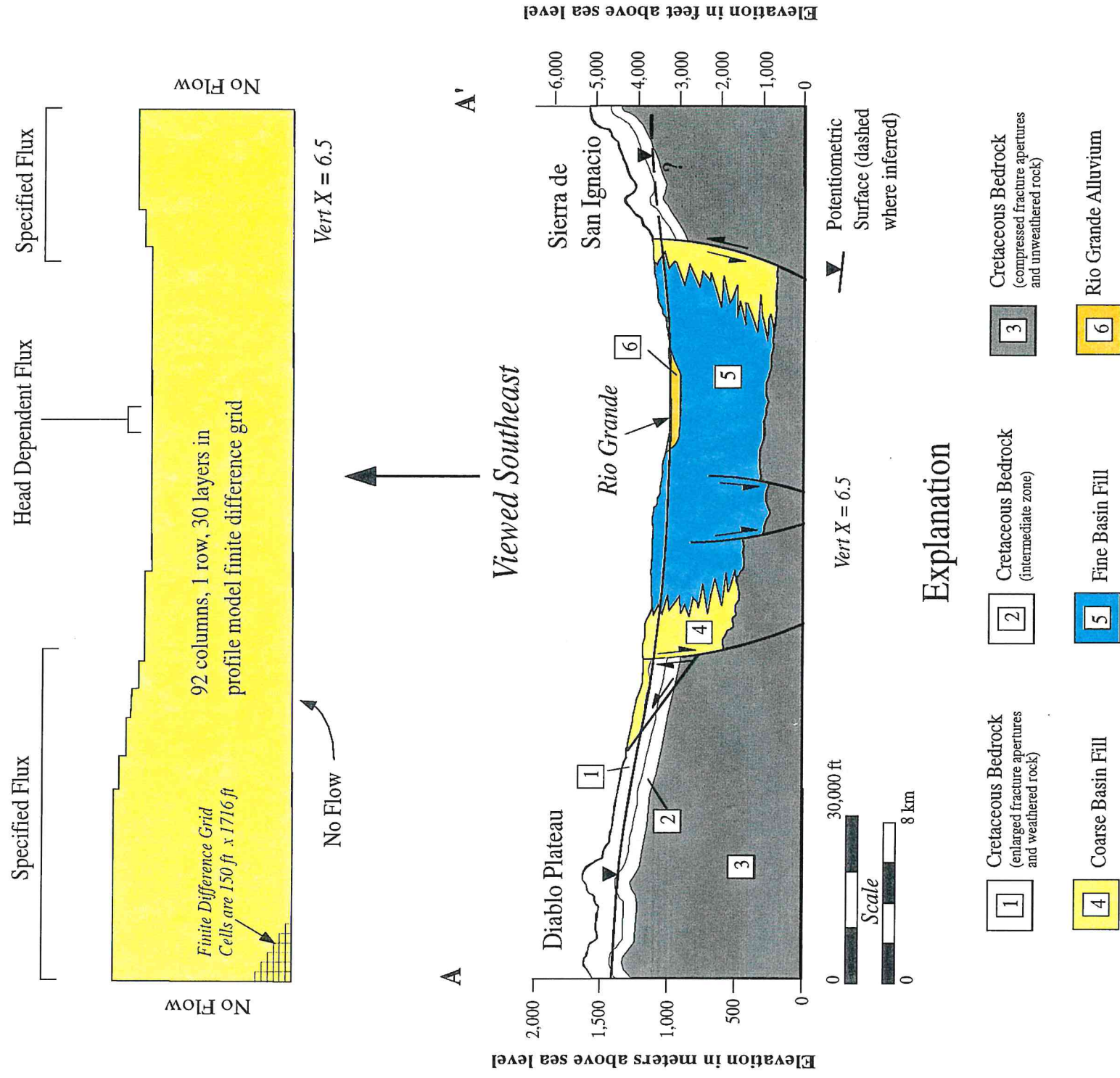


Figure 4.11. Diagram showing the gridding scheme, the aquifer zones, and the boundary conditions selected for the numerical profile model oriented between the Diablo Plateau and the Sierra de San Ignacio (key: zone 1 = Cretaceous rock with enlarged fractures and weathered rock; zone 2 = Cretaceous rock, intermediate zone; zone 3 = Cretaceous rock with compressed fracture apertures and unweathered rock; zone 4 = sand rich basin fill; zone 5 = mud rich basin fill; zone 6 = Rio Grande alluvium).

The northern boundary of the model is no flow, which corresponds to a ground-water divide on the Diablo Plateau. A prescribed flux boundary replenishes the aquifer to the south of the divide. The southern boundary of the model corresponds to an assumed ground-water divide at the Sierra de San Ignacio in northern Mexico. Near the Rio Grande, head-dependent flux boundaries were selected to correspond to low-lying areas near the river where discharge by evapotranspiration and river leakage occurs (Figure 4.11).

Hydraulic conductivities assigned to the model grid were selected from published values (Davis, 1969; Wolff, 1982; Bedinger and others, 1986; Kernodle, 1992; Mullican and Senger, 1992) and from lithologic descriptions of rocks and sediments (Kreitler and others, 1986; Gustavson, 1990; Collins and Raney, 1991). Six permeable rock and sediment zones were specified in the model (Figure 4.11). Zones 1, 2, and 3 were specified in Cretaceous water bearing rocks in the Diablo Plateau and Sierra de San Ignacio (Figure 4.11). These rocks were assigned hydraulic conductivity values of 0.1, 0.03, and 0.007 ft/day (Table 4.2). Cretaceous strata correspond to weathered, slightly fractured carbonate and clastic rocks with expanded fracture apertures (upper zone 1), a intermediate zone (zone 2), and unweathered, slightly fractured carbonate and clastic rocks with compressed fracture apertures (lower zone 3). No evidence of significant karstification of rocks is found in either outcrop or core in the area and flow is fracture and matrix controlled.

The higher permeability pathways near mountain surfaces are assumed to be associated with the tendency of fractures to close with depth due to the increase of mechanical stress (Figure 4.11). Bedinger and others (1986), for example, suggest that hydraulic conductivity is as much as one to three orders of magnitude higher in the upper 100 to 1000 ft of land surface due to weathering, jointing, and expansion of fracture apertures that succeeds erosional unloading of rock overburden. Empirical laws broadly applicable to these permeability trends show the relationship between permeabil-

ity distribution with depth in fractured rocks (Snow, 1968; Carlsson and Olson, 1977):

$$K(z) = K_s \cdot 10^{-z/l} \quad (4.1)$$

$$K(z) = K_s \cdot Z^{-1.6} \quad (4.2)$$

where:

$K(z)$ = hydraulic conductivity as a function of depth

l = a parameter in the range of 100 to 500 m

z = depth in meters beneath land surface

(+ downward)

K_s = hydraulic conductivity at land surface

Only small expansion of fracture apertures due to erosional unloading will impart significantly greater permeability to fractured rock units. Shown in simple terms for a parallel plate fracture model (Snow, 1968):

$$K = \frac{W^3}{12\Delta} \quad (4.3)$$

where:

k = intrinsic permeability

w = uniform fracture aperture width

Δ = uniform spacing between fractures

or for a cubic fracture model (Snow, 1968):

$$K = \frac{W^3}{6\Delta} \quad (4.4)$$

The cubic relationship between fracture aperture width and permeability shown in (4.3) and (4.4) implies that small compression of fractures due to increasing rock overburden can cause a significant decrease in hydraulic conductivity with depth beneath mountain surfaces, and provides a theoretical basis for this assumption in the numerical model. Ancillary data

in the southeastern Hueco aquifer indicate that Cretaceous rocks tested at depths between 600 and 800 ft had lower permeabilities than Cretaceous rocks tested at shallower depths (Mullican and Senger, 1992). Whether this reflects lower permeabilities due to compression of fracture apertures and absence of weathering at greater depth, or a change in lithology, is unknown.

Basin fill deposits that comprise zones 4 and 5 were assigned hydraulic conductivity values of 1.0 and 0.2 ft/day (Figure 4.11; Table 4.2). Higher permeabilities were assigned to the sand rich deposits that flank the highlands and form as alluvial fans (zone 4). Lower permeabilities were assigned to the sandy-silt and clay rich playa and riverine deposits that formed in the Hueco Bolson (zone 5). The Rio Grande alluvium (zone 6), noted for its relatively high permeability, was assigned a hydraulic conductivity value of 10.0 ft/day (Table 4.2). Cenozoic basin fill and Rio Grande alluvium were assigned horizontal to vertical anisotropy ratios of 10:1 (Kernodle, 1992).

Effective porosity values required for ground-water travel time estimates included (Table 4.2): Cretaceous carbonate and clastic rocks (0.08 weathered, 0.05 intermediate, and 0.02 unweathered), sand rich basin fill (0.18), mud rich basin fill (0.25), and Rio Grande alluvium (0.20). These values were compiled from literature values (Wolff, 1982; Bedinger and others, 1986), and from lithologic descriptions of the rocks and sediments (Albritton and Smith, 1965; Kreitler and others, 1986; Gustavson, 1990; Collins and Raney, 1991).

Model results

Model calibration was constrained by using recharge rates of between 0.3 - 3.0% of mean annual precipitation (12 in/yr) as suggested for approximate mountain-front recharge rates in this area by other authors (Kelly and Hearne, 1976; Orr and Risser, 1992). Calibration of the model was achieved by closely matching measured and simulated heads under the additional constraint that travel times estimated by the particle tracking simulator agreed with ground-water residence times

estimated by environmental isotopes. Calibrated recharge rates in the Diablo Plateau that averaged 0.14 in/year (1.2% of mean annual precipitation) provided a good match between measured and simulated heads in the American part of the southeastern Hueco aquifer (Figure 4.12). In the Mexican part of the aquifer, published head data are not available beneath the Sierra de San Ignacio and water levels are predicted based on final recharge rates at the Diablo Plateau (Figure 4.12 & 4.13).

Particle tracking results show effects of higher permeability materials on ground-water flow (Figure 4.13 & 4.14). At the Diablo Plateau, particles tend to flow along higher permeability bedrock units specified close to mountain surface, except near the northernmost ground-water divide where vertical hydraulic gradients drive ground water beneath the higher permeability bedrock zones (Figure 4.14). Likewise, these zones do not influence particle trajectories in northern Mexico because of the propensity for vertical flow.

In both the northern and southern portions of the model, the alluvial fans (zone 4) influence particle trajectories and act as sinks for ground-water flow (Figure 4.14). Particles near the lowermost model boundary move vertically upward to the higher permeability alluvial fans in order to follow paths of least hydraulic resistance. The fans act as convergence zones for both short and long flowpaths and thereby function as mixing zones of old and young ground water. Model results are in agreement with environmental isotopes that indicate old ground water at the alluvial fans (established by small percentages of modern carbon) that mix with smaller amounts of young, tritiated ground water (Figure 4.6). Once in the fans, the particles move laterally into lower permeability basin fill and then laterally and vertically upward to low-lying discharge areas near the Rio Grande (Figure 4.14).

Travel times in the model suggest that ground water moving from the models northern boundary may be old (e.g., 20,000 yrs) when it reaches the alluvial fans

Profile Model Parameters, Southeastern Hueco Aquifer			
Model zone and rock or sediment unit	Hydraulic conductivity (ft/day)	Effective Porosity	Horizontal to vertical anisotropy ratio
#1: Carbonate and clastic rocks, weathered and expanded fractures	0.1	0.08	1
#2: Carbonate and clastic rocks, intermediate zone	0.03	0.05	1
#3: Carbonate and clastic rocks, unweathered and compressed fractures	0.007	0.02	1
#4: Sand rich basin fill	1.0	0.18	10
#5: Mud rich basin fill	0.2	0.25	10
#6: Rio Grande alluvium	10.0	0.20	10

Table 4.2. Numerical profile model parameters, by zone.

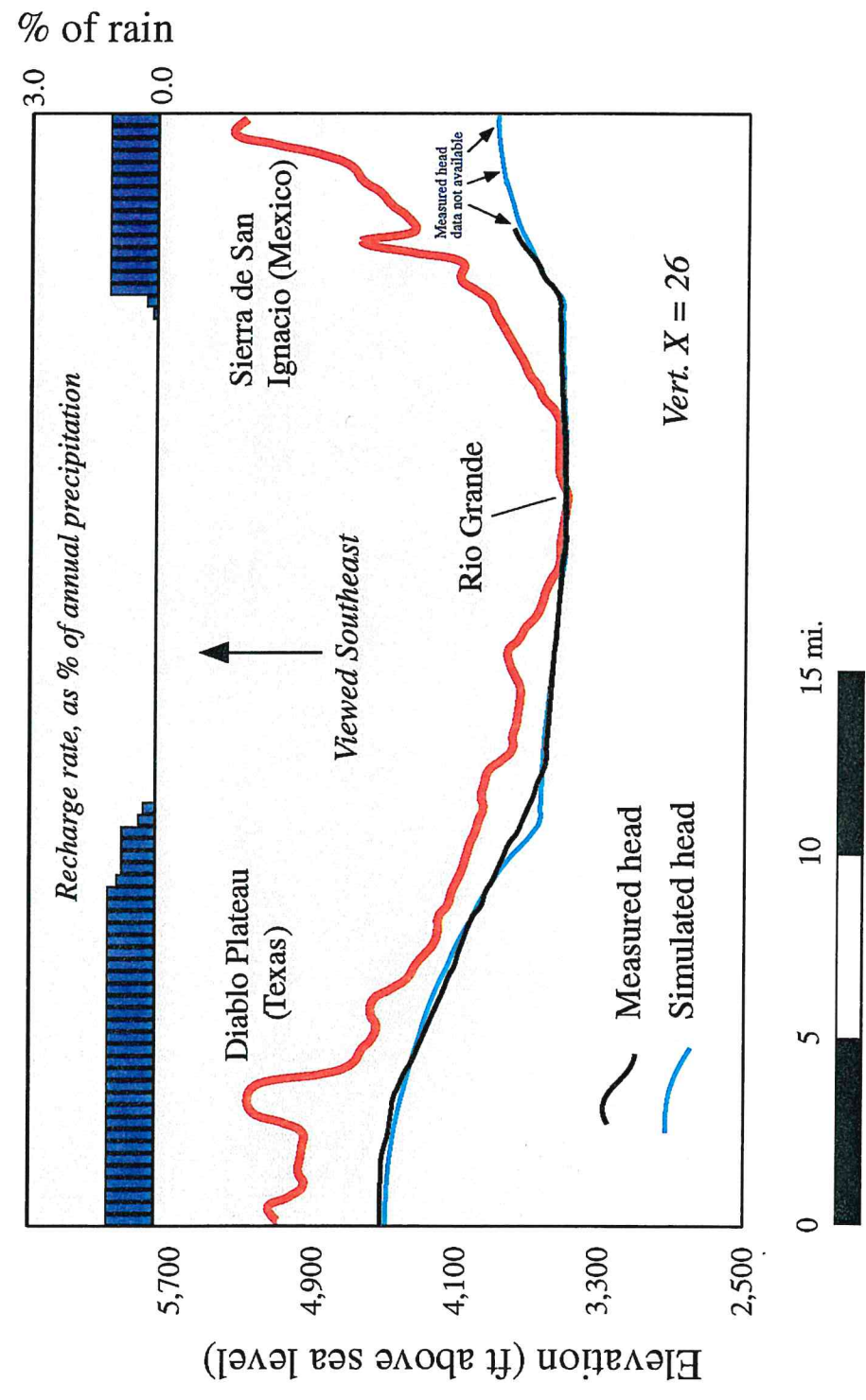


Figure 4.12. Recharge rates and comparison between measured and simulated hydraulic heads in the transboundary profile model.

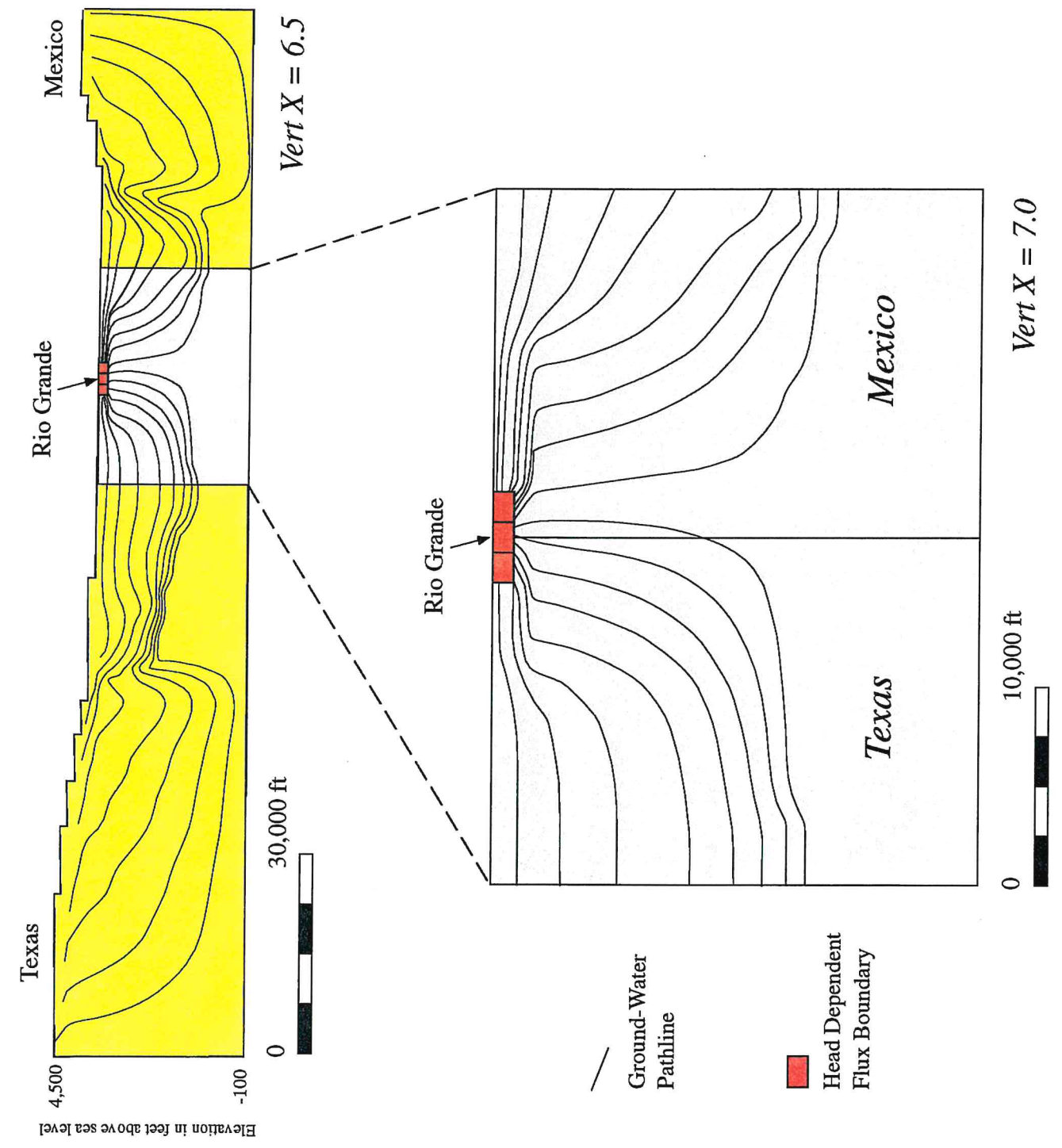


Figure 4.13. Particle trajectories do not form a perfect mirror image where they upwell beneath the international boundary. Pathlines suggest movement of ground water slightly southward into Mexico.

Particle Tracking Results

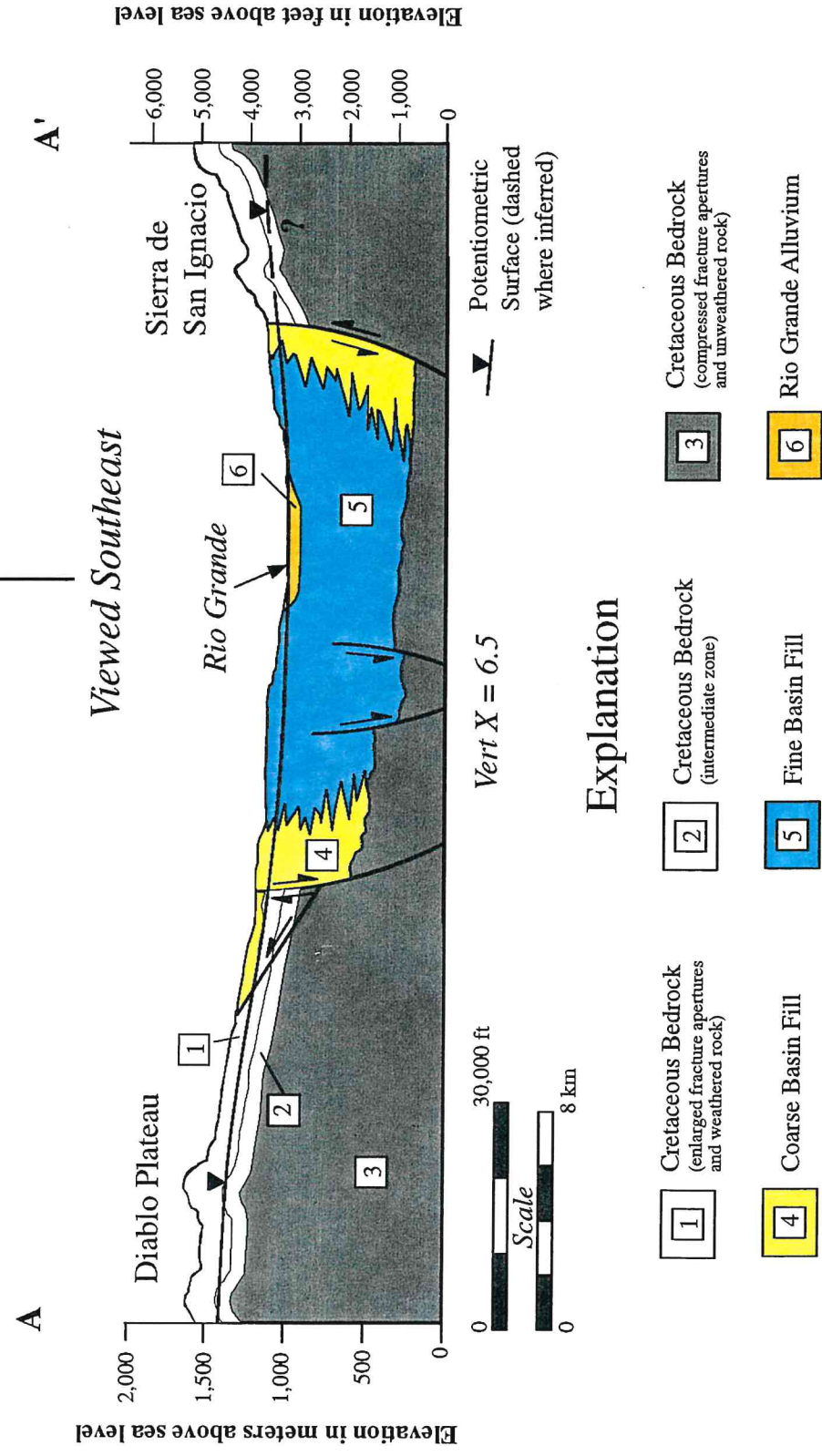
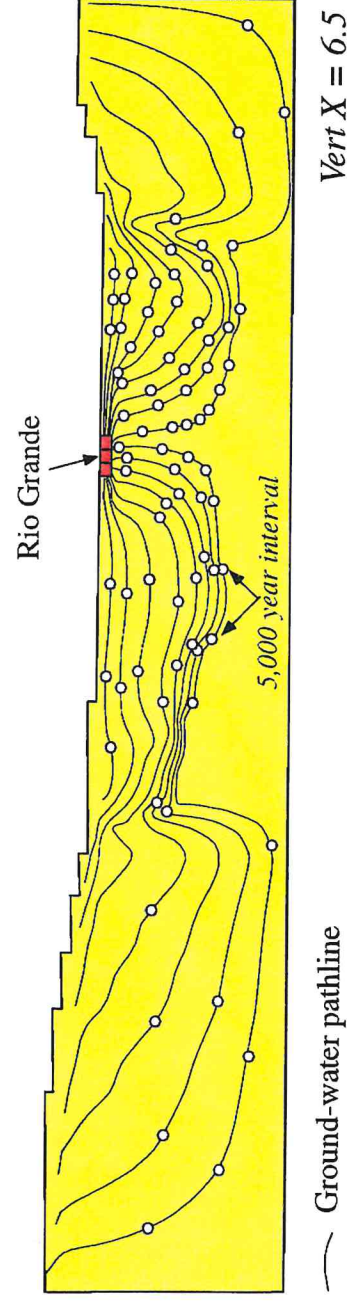
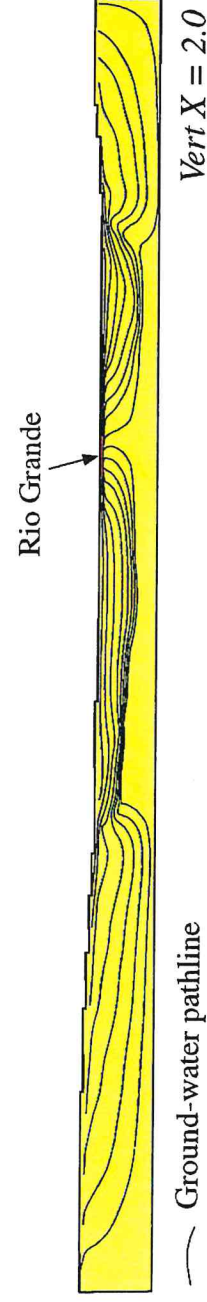


Figure 4.14. Simulated pathlines in the numerical profile model. Alluvial fans act as convergence zones for both short and long pathlines. Pathlines marked by 5,000 year travel times (squares) at segments along the pathline.

(Figure 4.14). Ground water moving from intermediate mountain elevations and from lower mountain elevations varies from moderately-old (1,000 to 8,000 yrs) to very young (100 yrs) when it reaches the alluvial fans. Total estimated travel times from principal recharge areas to the Rio Grande vary from about 15,000 years (alluvial fans Rio Grande) to about 60,000 years (Diablo Plateau ground-water divide Rio Grande).

Particle trajectories do not form a perfect mirror image where they upwell beneath the international boundary (Figure 4.13). Pathlines tend to suggest movement of ground water slightly southward into Mexico before upwelling near the Rio Grande. The asymmetric shape of modeled pathlines is a function of the asymmetric southeastern Hueco Bolson. Predicted movement is not substantially transboundary under simulated steady-state conditions.

Model analyses

Several analyses are performed to assess the response of the model to changes in recharge rates and boundary sinks. Model analyses include:

Scenario 1: Recharge rates in the Sierra de San Ignacio are decreased by a factor of 5.6.

Scenario 2: Heavy aquifer pumping is prescribed north of the Rio Grande.

Scenario 3: Recharge rates are specified to bring mountain flow systems to flow capacity.

Scenario 1

The lack of hydraulic head data in the Sierra de San Ignacio did not allow us to match measured and simulated heads in Mexico. Recharge rates were assumed equal in the Diablo Plateau and Sierra de San Ignacio. Flat lying surfaces in the Diablo Plateau may be more effective recharge areas than the steep terrains in the Sierra de San Ignacio (Figure 4.12). The steeper

mountain surfaces in Mexico may reduce infiltration rates and favor surface runoff.

To test the effect of lower recharge rates in Mexico due to topographic influences, the recharge rates in the Sierra de San Ignacio are decreased to 0.025 in/year (0.21% of mean annual precipitation) from an initial recharge rate of 0.14 in/year (Figure 4.15). The lower recharge rate is selected arbitrarily to assess effects on pathlines and transboundary ground-water flow. Recharge rates in the Diablo Plateau remain unchanged (Figure 4.15).

Movement of ground water from the United States into Mexico is more pronounced in the analysis (compare Figures 4.13 and 4.16). Ground water moves nearly 3,200 ft into Mexico, or about 2,000 ft more than in the initial model run. The Rio Grande alluvium acts to refract pathlines in Mexico and creates a more imperfect pathline image on both sides of the international border.

Scenario 2

Movement of ground water across the international border to pumping cones of depression is tested. A number of wells are placed in basin fill, 15,400 ft north of the Rio Grande. Water wells are pumped at a steady rate.

Drawdown cones of depression caused by well pumping are sufficient to lower heads in the southeastern Hueco Bolson and induce infiltration from the Rio Grande (Figure 4.17). Drawdown cones do not cause hydraulic head detachment from the river. Pathlines originating at higher elevations in the Sierra de San Ignacio move across the international border to the pumping cones of depression (Figure 4.18). Pathlines that originate at lower elevations in Mexico move to discharge areas near the Rio Grande.

Radial flow patterns are implicitly ignored in a two-dimensional profile model. The physics of drawdown in the profile model do not replicate precisely the phys-

ical drawdown configurations that accompany well pumping. Profile models are sometimes used even when pumping wells are oriented along the line of profile (Sanford and Konikow, 1985). Even so, results must be scrutinized as limited. Constraints are that pathlines move across the international boundary to areas where hydraulic head is below river stage, expected in areas where pumping causes substantial drawdown.

Scenario 3

Flow capacity is defined as the maximum amount of water that a flow system can accept and transmit. The phrase "rejected recharge" typically is applied to an aquifer at flow capacity where precipitation infiltration or other sources of potential recharge exceed the capacity of the saturated zone to accept additional recharge (Figure 4.19). Usually this results in regional saturation of the water table at land surface. The characteristics of a flow system that influence flow capacity are slope of the terrain and permeability of the aquifer (Mifflin, 1968).

Moisture is not available for flow capacity to be attained in most arid aquifers. Climatological change as a result of global warming or cooling may cause an arid system to later reach flow capacity. Abandoned landfills and dumps may become inundated, creating a potentially threatening water quality problem.

Paleohydrologic evidence of flow capacity in arid aquifers is observed in the southwestern Basin and Range province (Mifflin, 1968). Flow capacity was established in some of these aquifer during the Pluvial periods of the late Pleistocene Epoch, when precipitation was higher. Recent concerns about global warming intensify concerns about possible effects on ground-water quality. The southeastern Hueco aquifer, like most desert aquifer, is not at flow capacity today.

Recharge rates needed to attain flow capacity in the southeastern Hueco aquifer highlands are predicted by placing a series of drains at land surface elevation, and

by increasing recharge rates until flow capacity is attained (head is at land surface and springs are omnipresent). Specified flux is prescribed so that the recharge rates exceed the terrain's capacity to take the prescribed recharge, and potential recharge is rejected. Drains along the mountain surfaces act as conduits for rejected recharge (Figure 4.20).

Recharge rates needed to attain flow capacity are about 0.41 in/year (3.4% of mean annual precipitation) along the Diablo Plateau and 1.49 in/year (12.4% of mean annual precipitation) along the Sierra de San Ignacio (Figure 4.21). These recharge rates are 293% and 1064% higher than initial calibration recharge. Results imply that a moderate increase in the recharge rates at the Diablo Plateau might be sufficient to bring the Diablo Plateau to flow capacity. The substantial increase in the recharge rate required to bring the Sierra de San Ignacio to flow capacity is probably not realistic and suggests that this system will not attain flow capacity in Mexico.

Model limitations

The paucity of data along the model profile limits the use of the model beyond that of an interpretive tool for estimating ground-water flowpaths and velocities. The model presents a simplified picture of the hydrostratigraphy of the area, as defined by major structural and geologic features such as the southeastern Hueco Bolson. The simulated hydraulic gradient was suitably matched with the actual hydraulic gradient where head data were available, but the model's reliability is limited by the lack of information on vertical and horizontal hydraulic conductivity, effective porosity, and hydrostratigraphy. The limiting factors that are most pertinent to this modeling effort include the assumptions that:

- Fractured rock, at large scales, is equivalent to a porous medium.
- Ground-water flow is restricted to the plane of the profile model.

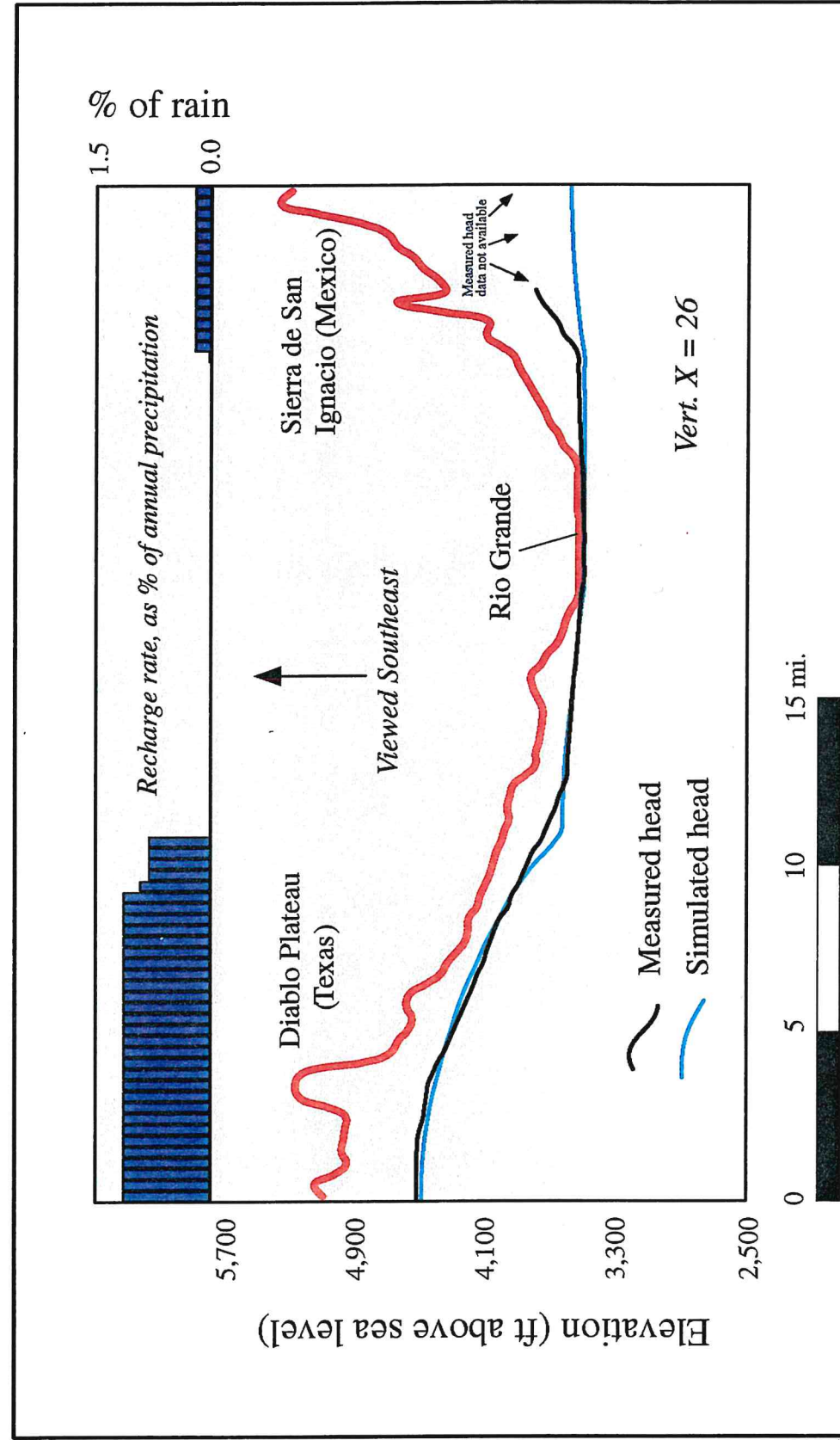


Figure 4.15. Recharge rates and comparison between measured and simulated hydraulic heads in model scenario 1. Prescribed recharge rates are much higher in the United States.

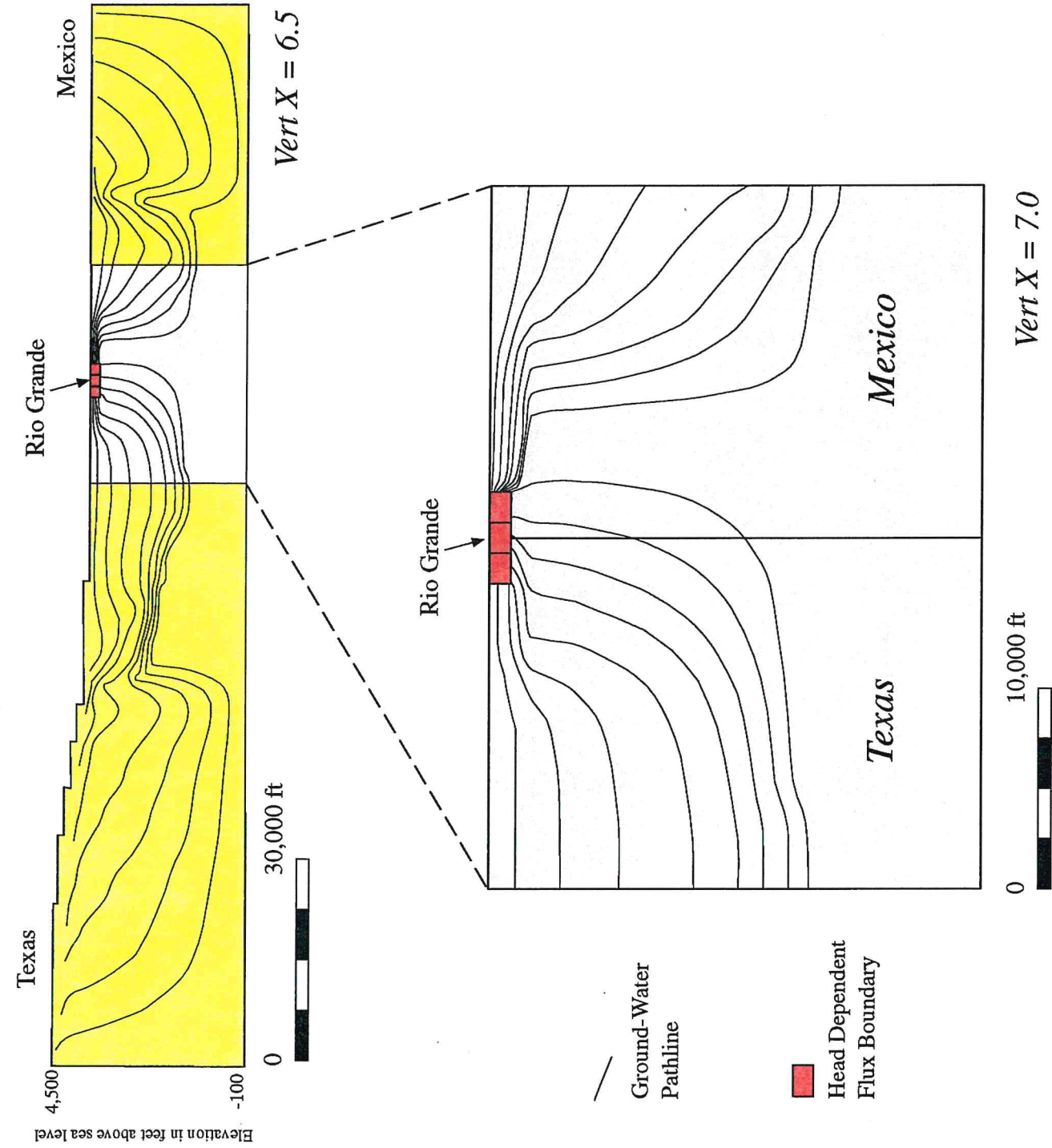


Figure 4.16. Pathlines suggest greater movement of ground water into Mexico with lower recharge rates in the Sierra de San Ignacio, model scenario 1.

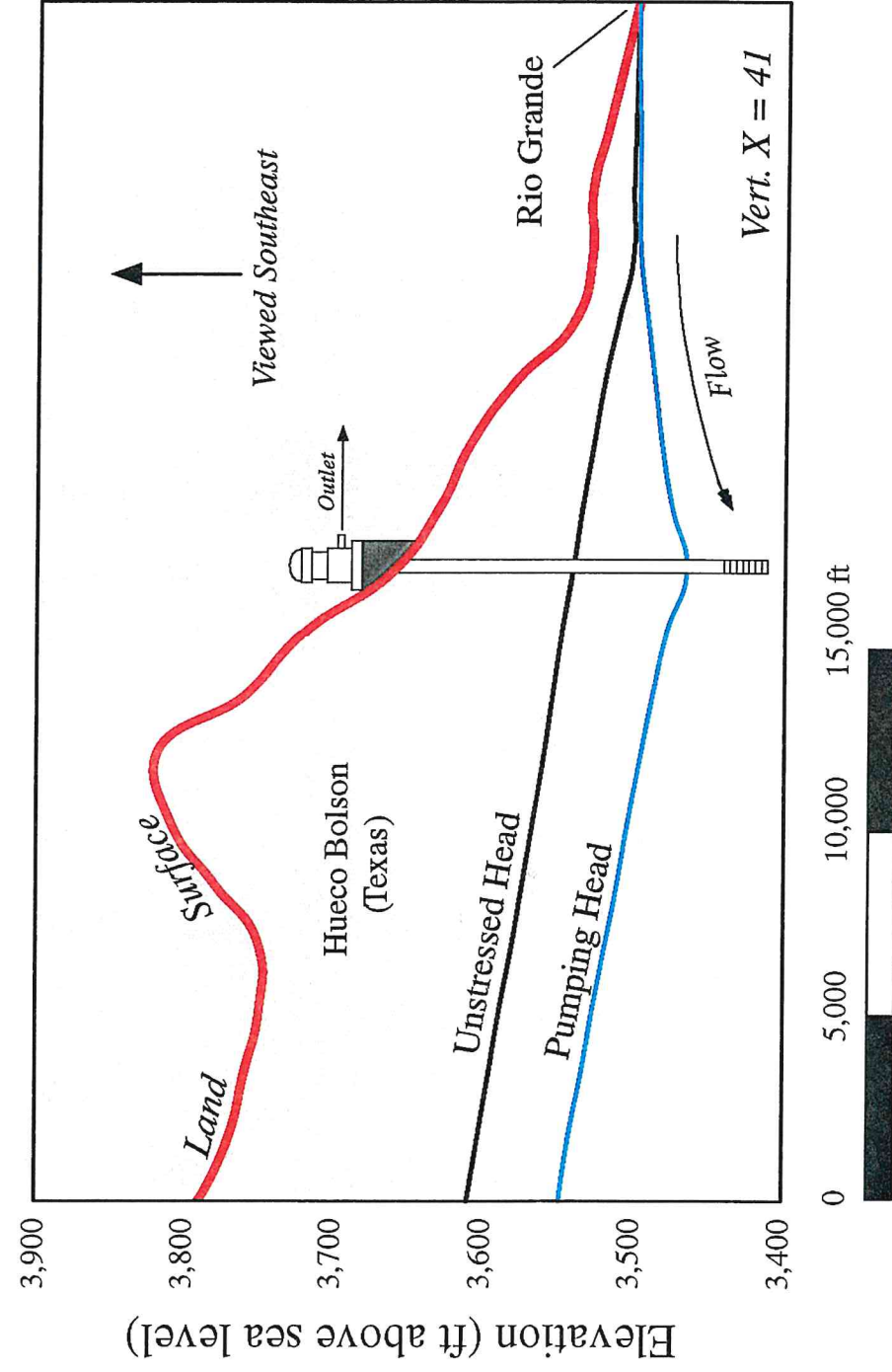


Figure 4.17. Comparison of unstressed hydraulic head and pumping hydraulic head in the numerical profile model, model scenario 2. Pumping cones of depression lower hydraulic head in the Hueco Bolson and induce infiltration of surface water from the Rio Grande.

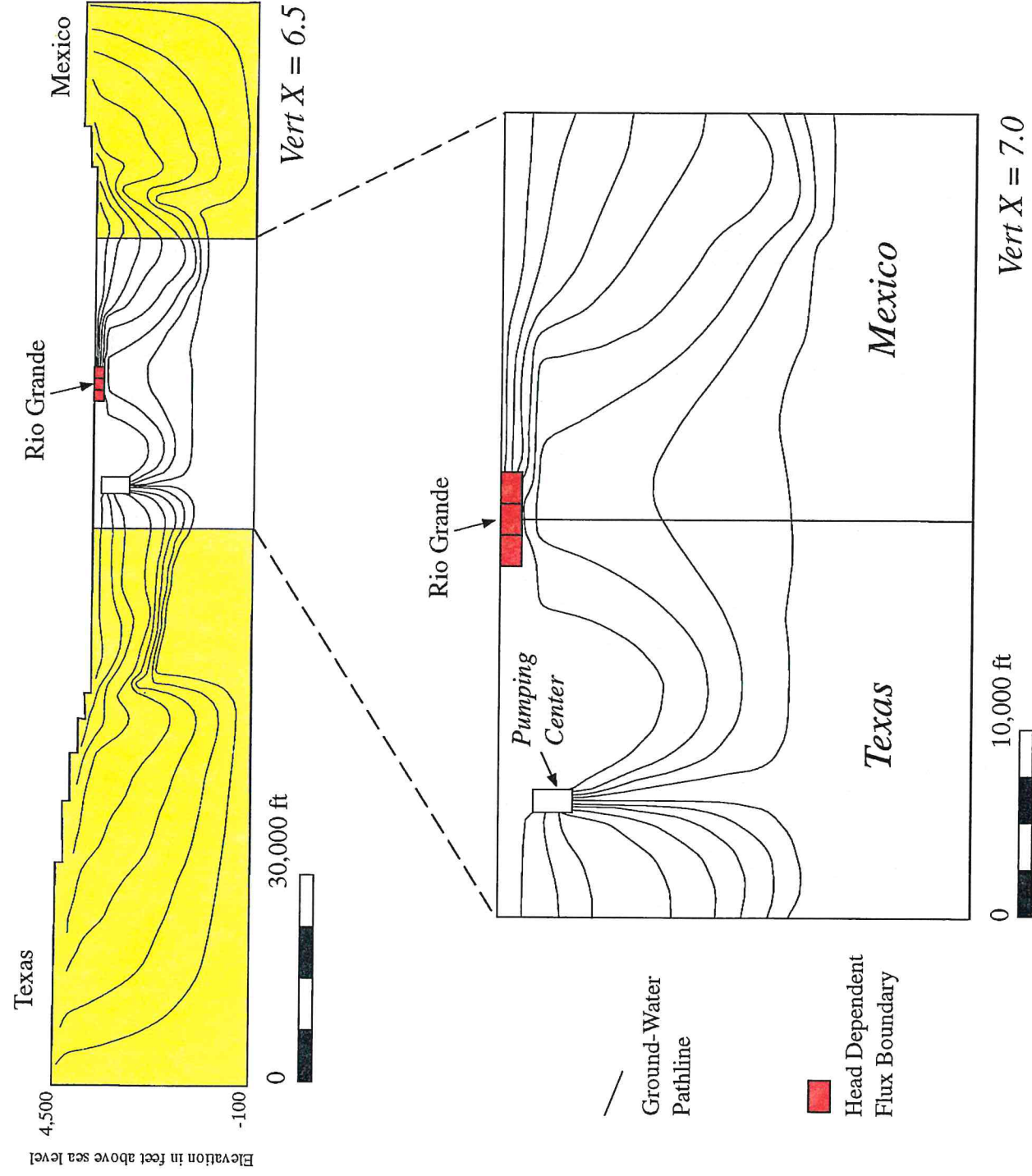


Figure 4.18. Pathlines originating in Mexico move across the international border to pumping cones of depression in the Texas portion of the southeastern Hueco Bolson in model scenario 2.

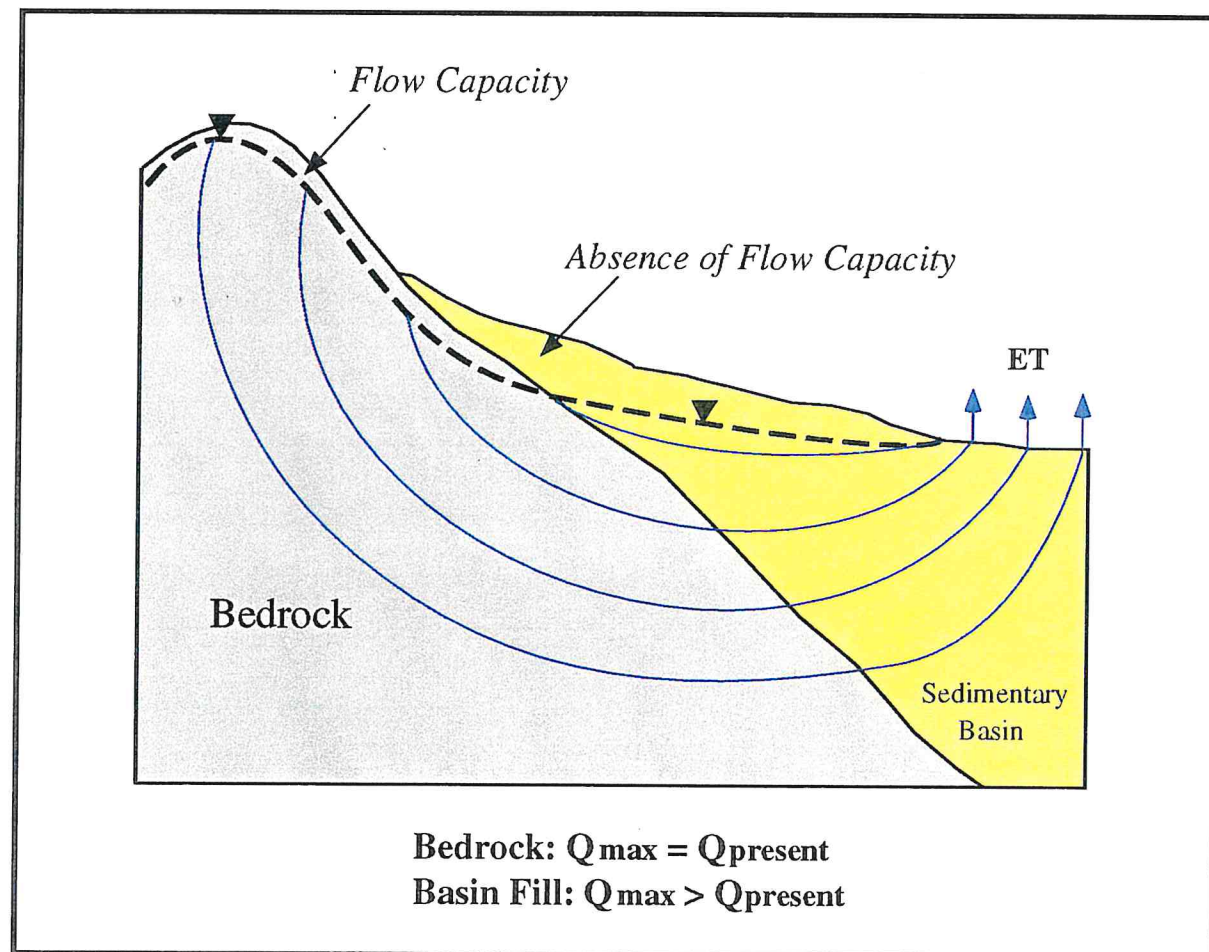


Figure 4.19a. Diagram showing flow capacity in the highland bedrock aquifer and absence of flow capacity in the sedimentary basin aquifer (modified from Mifflin, 1968).

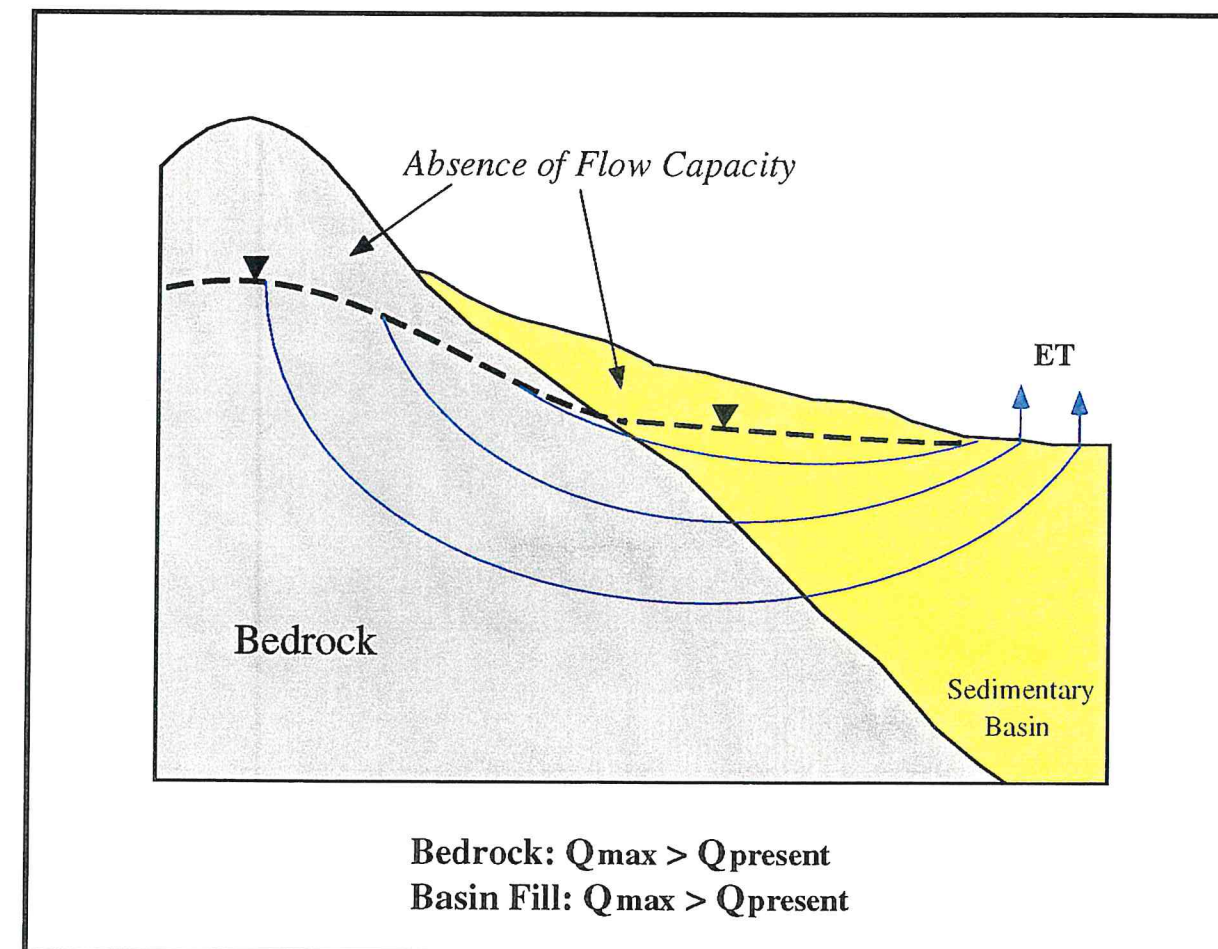


Figure 4.19b. Diagram showing absence of flow capacity in the bedrock and basin aquifers. Mountains in desert basins are typically not at flow capacity (modified from Mifflin, 1968).

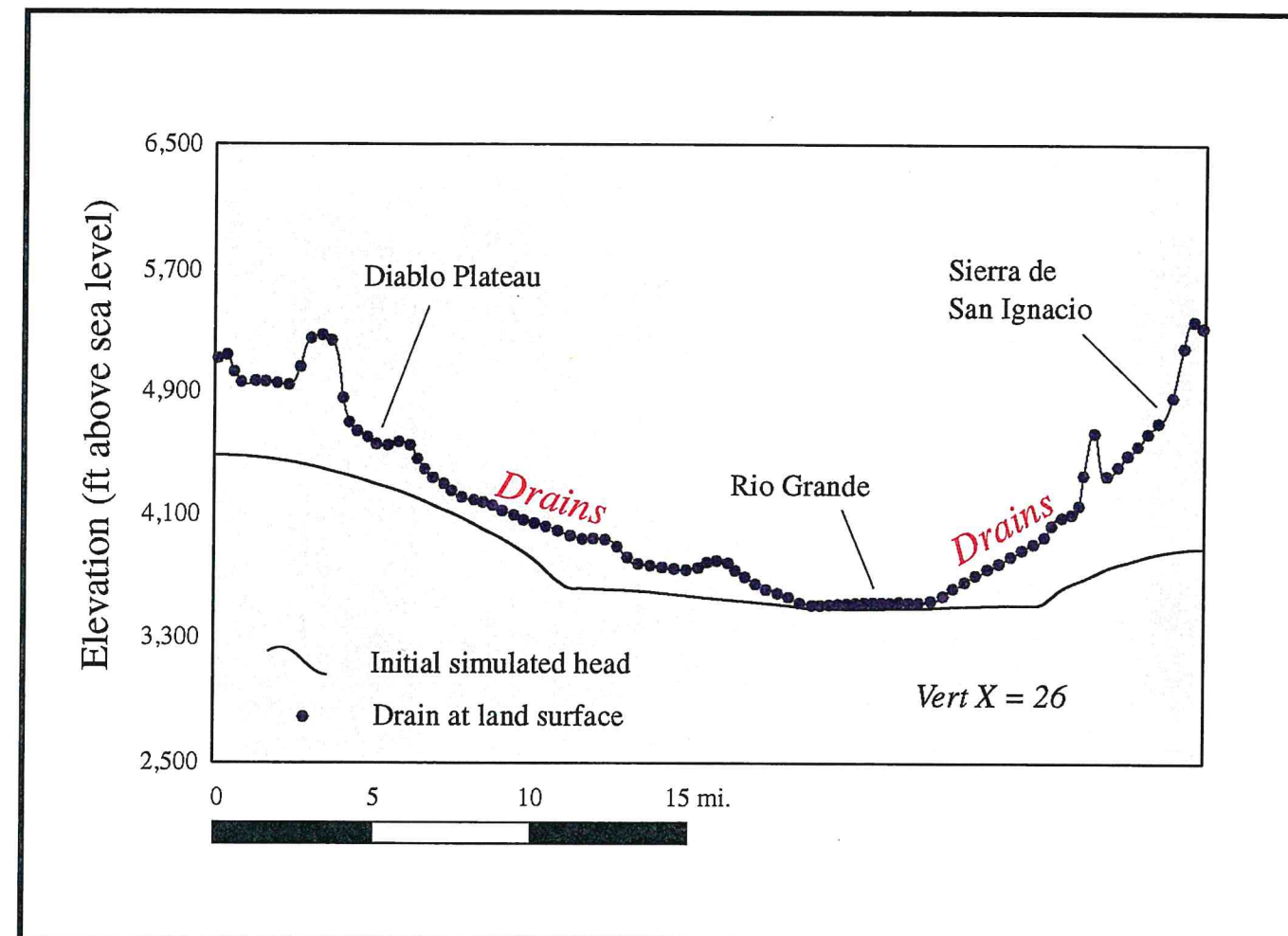


Figure 4.20. Drains specified at land surface elevation in the numerical profile model, model scenario 3.

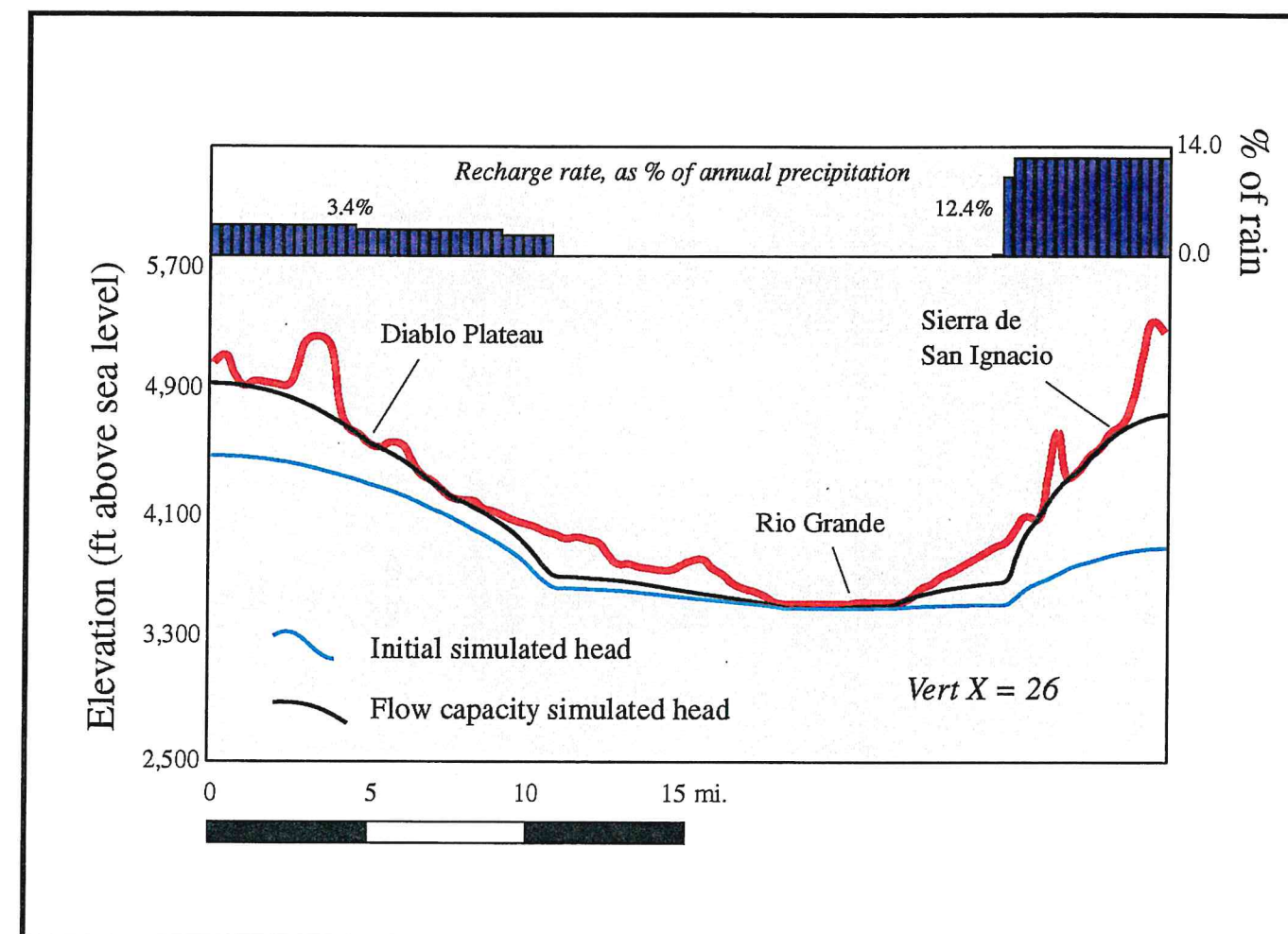


Figure 4.21. Comparison of initial simulated hydraulic head (see Figure 4.12) and simulated hydraulic head at flow capacity; and recharge rates at flow capacity, model scenario 3.

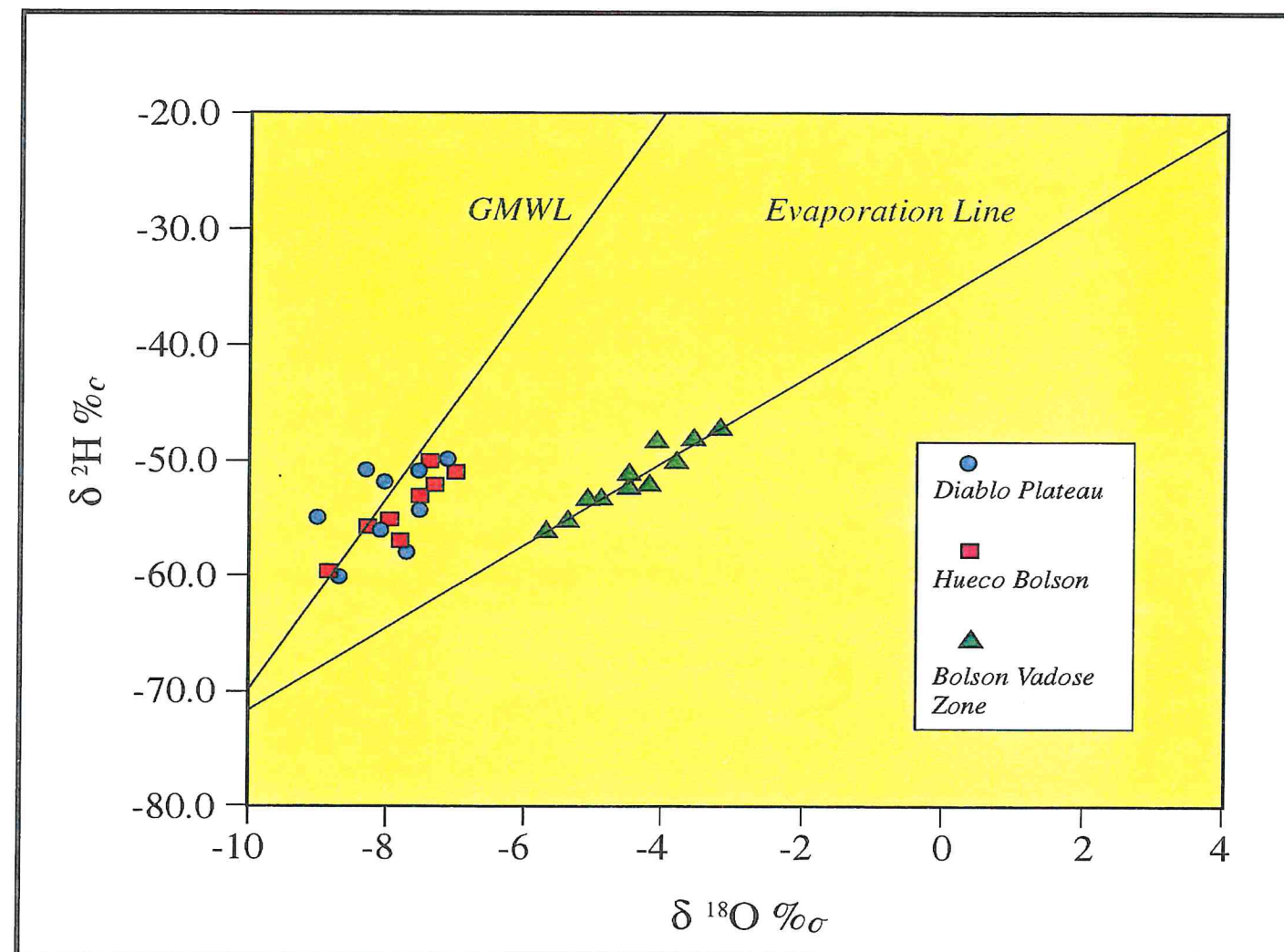


Figure 4.22. Binary plot of $\delta^2\text{H}$ versus $\delta^{18}\text{O}$ values for vadose zone and saturated zone waters in the southeastern Hueco aquifer. Stable hydrogen and oxygen isotope ratios in ground water (Diablo Plateau and Hueco Bolson) plot generally along the global meteoric water line (GMWL), indicating little or no enrichment associated with evaporation. Stable isotope ratios in vadose zone waters fall along an evaporation line. The lower slope is attributed to enrichment from the substantial excess of evaporation over precipitation. At distances from arroyos that act as sites of focused recharge, results imply no precipitation recharge along the basin floor due to high evaporation rates (source of data, Fisher and Mullican, 1990; Bureau Of Economic Geology, unpublished field data).

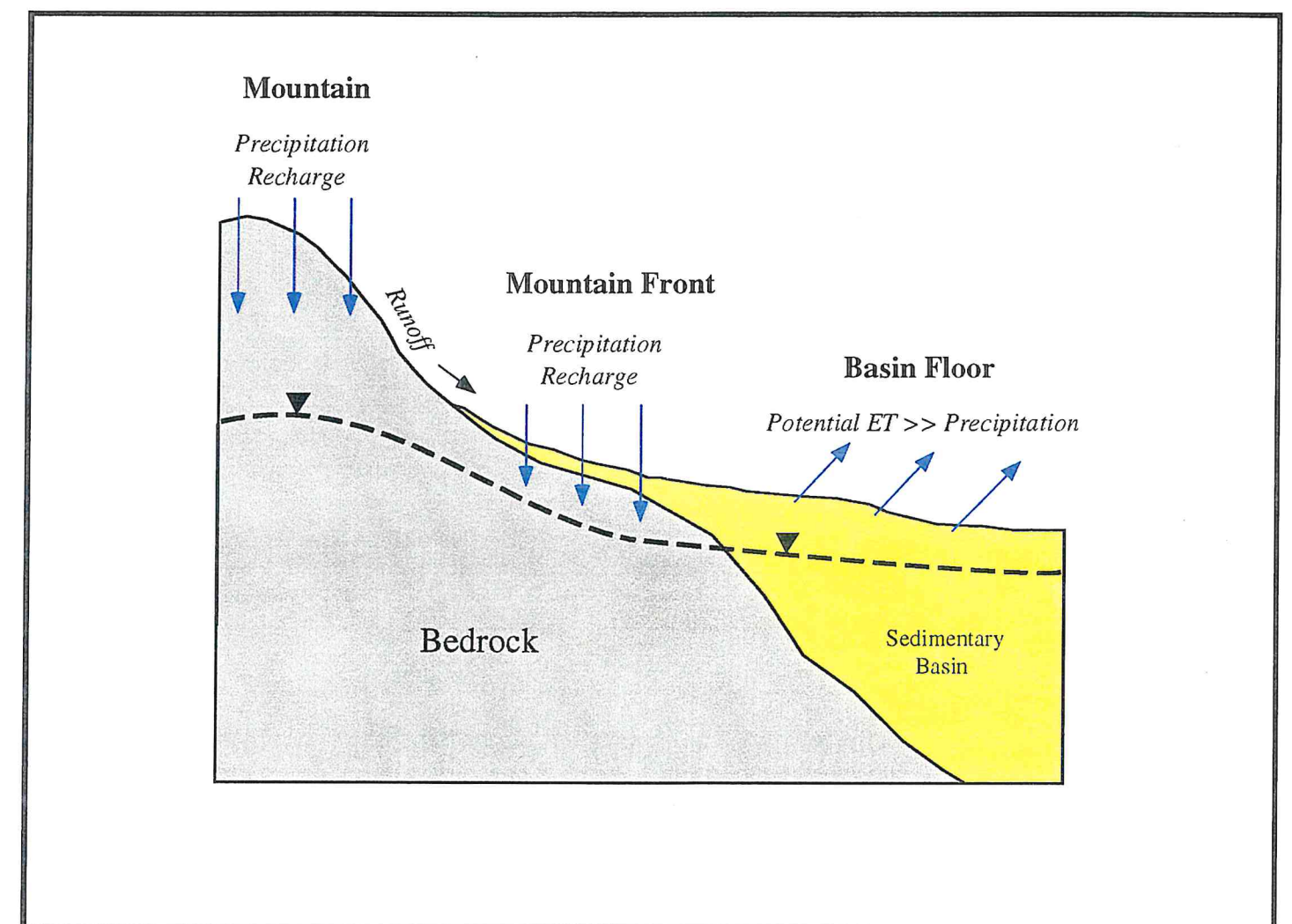


Figure 4.23. Conceptual model showing mountain and mountain front recharge and the absence of precipitation recharge along the basin floor due to the substantial excess of evapotranspiration over precipitation.

- Recharge rates at the Sierra de San Ignacio are approximately equal to recharge rates at the Diablo Plateau.

- Each of the zones has a constant vertical and horizontal hydraulic conductivity and effective porosity.

Of these, the most limiting is the last assumption. It is certain that rock units in any particular zone are laterally and vertically heterogeneous. These zones were defined by the boundaries between rock and sediment types. Within Cretaceous rocks for example, structural attributes and transitions in the potentiometric surface were used to separate the water-bearing units into zones. The simplistic definition of zones that have uniform hydrogeologic properties is required because borehole and aquifer test data are not available at most depths simulated in the model. Nevertheless, the model provides useful insights on pathlines, residence times, and potential transboundary movement of ground water in the asymmetric southeastern Hueco aquifer.

Susceptibility to Contamination

Bedrock units exposed in the southeastern Hueco aquifer are moderately susceptible to contamination. The Diablo Plateau has a moderate ground-water pollution potential (DRASTIC index) that ranges from 95 - 124 for agricultural sources (TWC, 1989). The DRASTIC index for general, municipal, and industrial sources is lower, ranging from 65 - 94 (TWC, 1989).

The bolson has a higher DRASTIC index, ranging from 110 - 124 for agricultural sources and from 80 - 94 for general, municipal, and industrial sources (TWC, 1989). Some qualification of this ranking is required. Even though the potential for contaminants at land surface being carried with infiltrating precipitation to the saturated zone is possible along arroyos (established by radioisotopes, Figure 4.6), the potential for contamination along areas of the basin floor that are not juxtaposed to arroyos is generally small. The rela-

tively dry climate, specific retention of the soil, and intensity and distribution of rainfall does not provide adequate moisture for wetting fronts to reach the saturated zone except along arroyos.

Evidence of the lack of precipitation recharge through soil profiles beneath the basin floor is provided by a comparison of stable isotopes for ground waters in the Diablo Plateau and Hueco Bolson with soil water extracted from the unsaturated (vadose) zone in the bolson (Figure 4.22). Stable isotopes in ground water typically plot along the global meteoric water line (GMWL) of Craig (1961), whereas the waters extracted from the vadose zone plot along an evaporation trend. The vadose zone waters are enriched with respect to the heavier stable isotopes ($\delta^{18}\text{O}$ and $\delta^2\text{H}$) due to the substantial evaporation of lighter water molecules (i.e., those that contain $\delta^{16}\text{O}$ and $\delta^1\text{H}$) from soil profiles (Figure 4.23). Having a slightly different vapor pressure, the forms of water molecules that contain the lighter isotopes tend to evaporate preferentially leaving behind a solution enriched with respect to heavier water molecules. This process is known as fractionation.

Results show a clearly defined stable isotope signature for ground waters in the southeastern Hueco aquifer that is genetically different from the stable isotope signature of water extracted from the vadose zone. The data indicate that ground waters are meteoric and undergo little isotopic fractionation by evaporation (Figure 4.23). Ground water is probably recharged rapidly in limited quantities after episodic precipitation and runoff events that provide sufficient moisture for ephemeral saturation of pediments and arroyos that overlie and flank the mountains.

Limited contamination potential may not be a valid assumption where anthropogenic structures are in place. Quarry's, excavations, and abandoned and poorly constructed water wells all can serve as conduits for contamination from land surface.

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