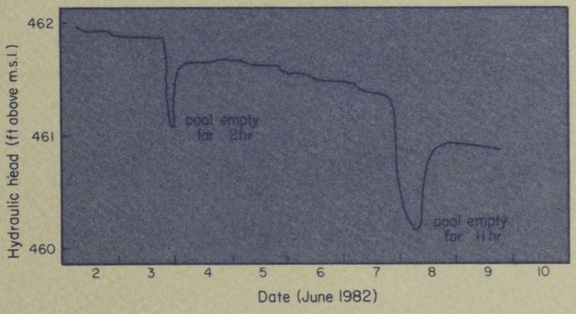
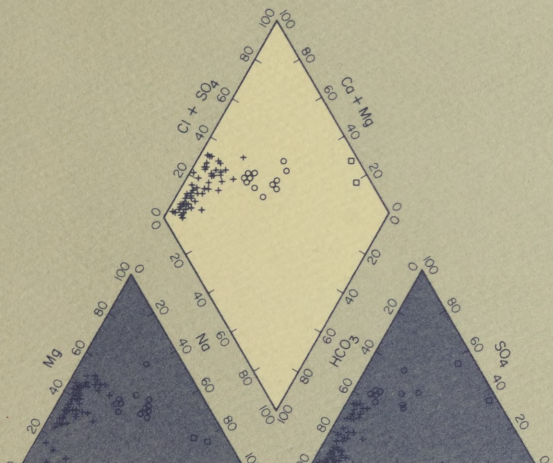
Report of Investigations No. 141

Hydrogeology of the Edwards Aquifer, Austin Area, Central Texas

Rainer K. Senger   Charles W. Kreitler





BUREAU OF ECONOMIC GEOLOGY
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W. L. Fisher, Director
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ABSTRACT

The Edwards Formation, on the downthrown side of Mt. Bonnell fault in the Austin, Texas, area (Hays and Travis Counties), is part of the northeastern extension of the Edwards Underground Reservoir, the primary source of water in numerous counties along the Balcones Fault Zone. Recharge to the aquifer is supplied mainly by creeks that cross the Balcones Fault Zone southwest of Austin. Barton Springs is the major point of discharge. Changes in water levels of wells in the area correlate positively with changes in discharge at Barton Springs, suggesting good interconnection. The potentiometric surface of the aquifer changes significantly from high flow to low flow at Barton Springs. During low-flow conditions, ground-water flow lines converge in the eastern part of the Balcones Fault Zone. Water levels are also much lower (less than 30 m) and indicate flow from the "bad-water" zone (water with 1,000 mg/L TDS or more from downdip in the Edwards Formation).

Water chemistry at Barton Springs also varies between high and low discharge. Concentrations of sodium, chlorine, sulfate, and strontium increase with decreasing discharge, indicating influx from the "bad-water" zone. This influx of highly saturated "bad-water" into the fresh-water aquifer theoretically results in a decrease in saturation state with respect to calcite and dolomite. The decrease in saturation state would enhance carbonate dissolution at the interface between fresh water and "bad-water" zones, thereby increasing permeabilities in this section of the aquifer. The Edwards aquifer generally contains a consistent calcium bicarbonate water. In some areas of the fresh-water section, however, leakage from the Glen Rose Formation increases the sulfate and strontium concentrations. Leakage occurs across fronts created by large displacements of faults that bring the Edwards Formation into contact with the Glen Rose Formation updip.

Keywords: Barton Springs, Edwards aquifer, Glen Rose Formation, leakage, "bad-water" zone, carbonate equilibria, karst hydrology.

INTRODUCTION

The Edwards Formation in Hays and Travis Counties, Texas, is part of the northeastern extension of the Edwards Underground Reservoir. In many counties along the Balcones Fault Zone (fig. 1), the reservoir is the primary source of municipal and private water supplies. The eastern and southeastern boundary of fresh water in the Edwards Underground Reservoir is marked by the "bad-water" line, which is the updip limit of nonpotable ground water containing total dissolved solids of 1,000 mg/L or more. A ground-water flow divide in Hays County, 15 mi (24 km) south of Austin, separates the Edwards Underground Reservoir into the Edwards aquifer, Austin region, and the Edwards aquifer, San Antonio region. The aquifer north of the ground-water flow divide and south of the Colorado River in the Austin region has major discharge points at Barton Springs, located along Barton Creek (fig. 2).

Urban development in the Austin area may affect natural systems and recreational features like Barton Springs. The Edwards aquifer in the Austin region is a potential source of drinking water, although currently it is not heavily used for domestic drinking water. Because of increased urban development, the aquifer may become more important as a source of drinking water. However, water in carbonate aquifers is typically considered to be vulnerable to contamination because of thin soil cover, fracture or vuggy porosity (which may permit a contaminant to pollute a large area), and lack of physical and chemical attenuation mechanisms commonly associated with intergranular flow.

Previous Work

The U.S. Geological Survey district office in Austin has been investigating the effects of urbanization on the quality of surface and subsurface water in the area. Slade and others (1982) reported general surface-water and ground-water conditions and discussed runoff phenomena and aquifer recharge and discharge. Guyton and Associates (1958) suggested the presence of a ground-water flow divide in Hays County that separates the Edwards aquifer in the Austin region from the Edwards aquifer in the San Antonio region (fig. 1). St. Clair (1978) investigated the effect of septic tanks on ground-water quality. Together with previous investigations of ground-

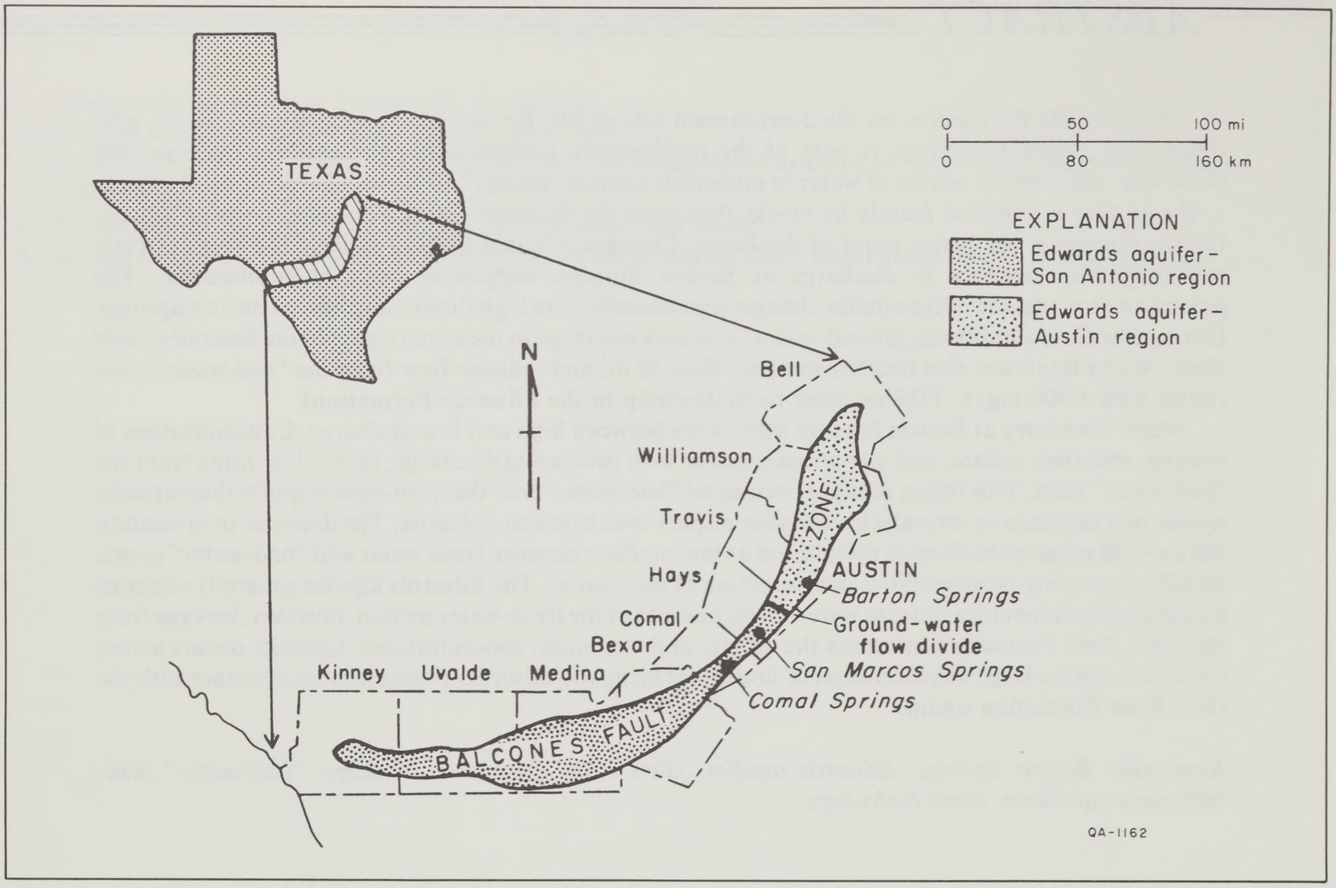

FIGURE 1. Division of the Edwards aquifer according to the Texas Department of Water Resources (1978).

water resources in Travis County that were conducted by the Texas Department of Water Resources and summarized by Brune and Duffin (1983), these reports provide important data on potentiometric levels and chemistry of surface and subsurface water.

Scope

This study, done in cooperation with the U.S. Geological Survey, Austin district office, concentrated on the hydrogeology and hydrochemistry of the Edwards aquifer and Barton Springs. The investigation was designed to (1) describe the stratigraphic and lithologic setting of the aquifer; (2) identify the dominant flow directions in the aquifer; (3) show the interconnection between Barton Springs and the aquifer; (4) document the hydrologic properties of the aquifer; (5) evaluate the chemical variations of Barton Springs water; and (6) characterize the water chemistry of the Edwards aquifer in the Austin area.

HYDROGEOLOGIC SETTING

Physiography and Climate

The Balcones Fault Zone marks the transition from the dissected remnants of the Edwards Plateau in the west (Hill Country) to the Blackland Prairie in the east. The physiography of this area is primarily due to differential erosion parallel to the numerous northeast-trending faults of the Balcones Fault Zone. Extensive faulting resulted in juxtaposition of different types of rock exhibiting varying degrees of resistance to erosion and supporting different assemblages of vegetation on the outcrop. Area topography is that of the Rolling Prairie province (Garner and Young, 1976). At some locations, the major creeks are entrenched into limestone valleys that have nearly vertical slopes.

The climate of the Austin area is subhumid. Short, mild winters are followed by short springs and long, hot

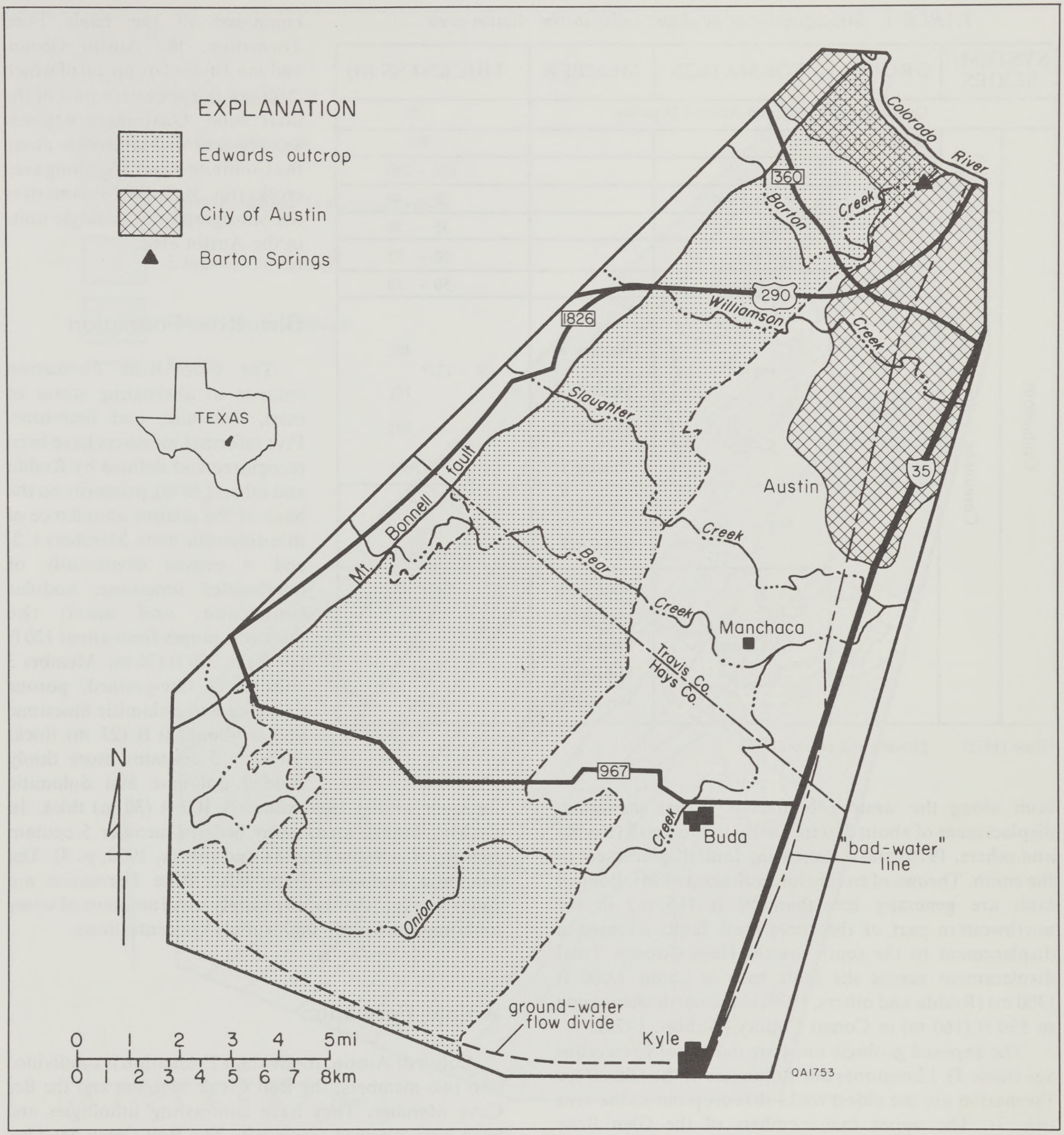

FIGURE 2. Location of the study area.

summers. Humidity is moderately high and the prevailing winds are southerly. The mean minimum temperature, 41° F (5° C) occurs in January, and the mean maximum temperature, 95° F (35° C), occurs in July. The average annual rainfall, calculated on measurements taken in 1941 through 1970, amounts to 32.5 inches (82.5 cm). Major rainstorms occur during the spring and fall.

Geology Related to Hydrology

The Balcones Fault Zone is a belt of northeast-trending, dip-slip, normal faults that displace gently eastward-dipping Cretaceous rocks down to the south-west in this area. Mt. Bonnell fault is the largest

TABLE 1. Stratigraphy of geologic units in the Austin area.

SYSTEM/SERIES	GROUP	FORMATION	MEMBER	THICKNESS (ft)
Quaternary Terraces and Alluvial Deposits				20
Cretaceous — Gulf Series	Taylor			300
Cretaceous — Gulf Series	Austin			130 - 250
Cretaceous — Comanche Series		Eagle Ford		20 - 40
Cretaceous — Comanche Series		Buda		35 - 50
Cretaceous — Comanche Series		Del Rio		60 - 75
Cretaceous — Comanche Series		Georgetown		50 - 55
Cretaceous — Comanche Series	Edwards*	Edwards‡		
Cretaceous — Comanche Series	Edwards*	Person*	Member 4‡	40‡
Cretaceous — Comanche Series	Edwards*	Person*		70 - 125*
Cretaceous — Comanche Series	Edwards*	Person*	Member 3‡	10‡
Cretaceous — Comanche Series	Edwards*	Kainer*	Member 2‡	40‡
Cretaceous — Comanche Series	Edwards*	Kainer*		300*
Cretaceous — Comanche Series	Edwards*	Kainer*	Member 1‡	200‡
Cretaceous — Comanche Series		Walnut	Bee Cave	30
Cretaceous — Comanche Series		Walnut	Bull Creek	35
Cretaceous — Comanche Series		Glen Rose	Member 5	90 - 100
Cretaceous — Comanche Series		Glen Rose	Member 4	125
Cretaceous — Comanche Series		Glen Rose	Member 3	70
Cretaceous — Comanche Series		Glen Rose	Member 2	120
Cretaceous — Comanche Series		Glen Rose	Member 1	250

*Rose (1972) ‡Rodda and others (1966)

fault along the western boundary. It has maximum displacement of about 720 ft (220 m) in the north (Rodda and others, 1970) and a decreasing fault displacement to the south. Throws of en echelon faults east of Mt. Bonnell fault are generally less than 50 ft (15 m) in the northwestern part of the zone; these faults increase in displacement to the south toward Hays County. Total displacement across the fault zone is about 1,000 ft (300 m) (Rodda and others, 1970) in the north, decreasing to 520 ft (160 m) in Comal County (Abbott, 1975).

The exposed geologic units are mainly of Cretaceous age (table 1). Limestones and dolomites of the Glen Rose Formation are the oldest rocks that crop out in the area (fig. 3). The upper two members of the Glen Rose Formation as well as the overlying Walnut Formation exist only in small outcrops along the Mt. Bonnell fault in the southwestern part of the study area. Rudist limestones and dolomites of the Edwards Formation are the most abundant rocks in the area. The Edwards Formation and the overlying Georgetown Formation are considered to be hydrologically connected in the Austin area (Baker and others, in press) and constitute the Edwards aquifer. Above the Georgetown Formation, the Del Rio Clay and the Buda Formation conclude the Lower Cretaceous Comanche Series. The Upper Cretaceous Gulf Series is composed of the Eagle Ford Formation, the Austin Group, and the Taylor Group, all of which crop out in the eastern part of the fault zone. Quaternary deposits are terraces and alluvium along the Colorado River and along area creeks (fig. 3). Table 1 summarizes the stratigraphy of geologic units in the Austin area.

Glen Rose Formation

The Glen Rose Formation consists of alternating strata of marl, dolomite, and limestone. Five informal members have been recognized and defined by Rodda and others (1970), primarily on the basis of the relative abundance of thin dolomitic beds. Members 1, 2, and 4 consist dominantly of interbedded limestone, nodular limestone, and marl; the thickness ranges from about 120 ft (37 m) to 250 ft (76 m). Member 3 consists of fine-grained, porous dolomite and dolomitic limestone and is about 70 ft (21 m) thick. Member 5 contains more thinly bedded dolomite and dolomitic limestone and is approximately 100 ft (30 m) thick. In the Mt. Bonnell area, many beds in member 5 contain pockets of celestite (Rodda and others, 1970, p. 3). The dolomitic members of the Glen Rose Formation are minor aquifers that locally supply small amounts of water containing relatively high sulfate concentrations.

Walnut Formation

South of Austin, the Walnut Formation is subdivided into two members, the Bull Creek Member and the Bee Cave Member. They have contrasting lithologies and have been mapped separately. The Bull Creek Member consists of about 35 ft (10.5 m) of hard, fine-grained to coarse-grained fossiliferous limestone. The Bee Cave Member consists of nodular marl and limestone and has a total thickness of about 30 ft (9 m).

Edwards Formation

The Edwards Formation consists of rudist limestone, dolomite, nodular chert, and solution collapse breccias

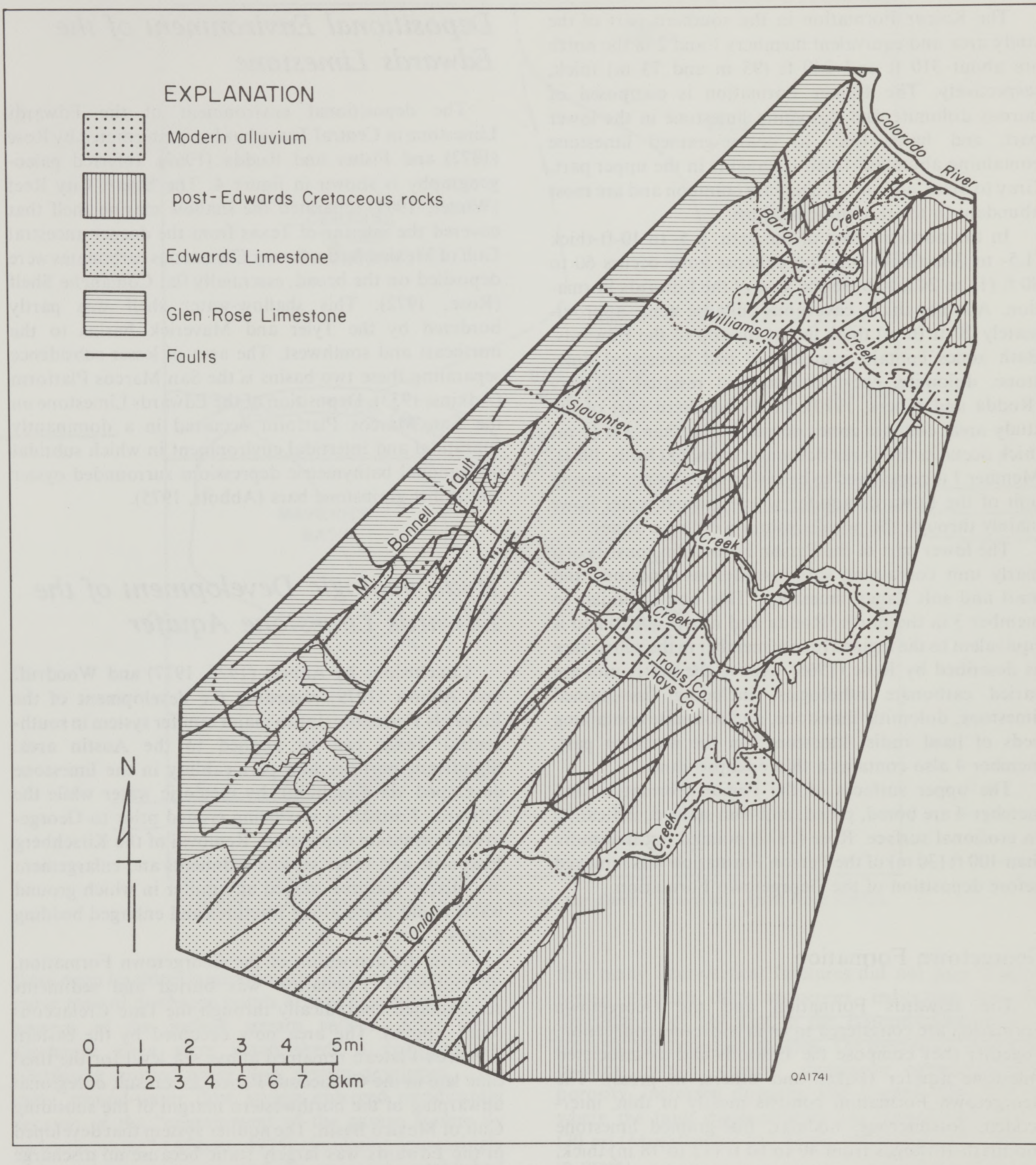

FIGURE 3. Geologic map of the study area, after Rodda and others (1970), Smith (1978), Kolb (1981), and Garner (unpublished data).

(Rodda and others, 1966; Fisher and Rodda, 1969). In the Austin West quadrangle and in the northern part of the Oak Hill quadrangle, the Edwards has been subdivided into four informal members on the basis of lithology (Rodda and others, 1970). In contrast, Smith (1978) and Kolb (1981) used the terminology of Rose (1972), who elevated the Edwards to group status and named two new formations, the Kainer and the Person Formations (table 1). In the northern part of the area, the Edwards is about 300 ft (91 m) thick. The formation generally thickens downdip toward Hays County, where it is about 400 ft (122 m) thick (Smith, 1978).

The Kainer Formation in the southern part of the study area and equivalent members 1 and 2 in the north are about 310 ft and 240 ft (95 m and 73 m) thick, respectively. The Kainer Formation is composed of porous dolomite and dolomitic limestone in the lower part, and hard, fine- to coarse-grained limestone containing abundant fossil fragments in the upper part. Gray to black nodules of chert are common and are most abundant in the dolomitic sections.

In the northern part of the area, a 5- to 10-ft-thick (1.5- to 3-m-thick) solution collapse zone occurs 60 to 80 ft (18 to 24 m) above the base of the Edwards Formation. Another thick, cavernous collapse zone, approximately 20 ft (6 m) thick lies at the top of the Edwards. Both zones contain iron-stained and brecciated limestone, dolomite, chert, calcite, and residual red clay (Rodda and others, 1970). In the southern part of the study area, collapse zones normally less than 3 ft (1 m) thick occur in the lower dolomitic member (Kolb, 1981). Member 1 is considered to be the principal water-bearing unit of the Edwards aquifer where ground water flows mainly through the porous solution collapse zones.

The lower part of the Person Formation consists of a marly unit containing soft fossiliferous limestone and marl and soft flaggy limestone. This unit is similar to member 3 in the north (Rodda and others, 1970) and is equivalent to the regional dense member in the subsurface as described by Rose (1968). Above the marly unit are varied carbonate lithologies, including fine-grained limestone, dolomitic limestone, and dolomite containing beds of hard rudist limestone. In the northern part, member 4 also contains a thin collapse zone.

The upper surfaces of the Person Formation and member 4 are bored, pitted, and iron stained, indicating an erosional surface. Rose (1972) pointed out that more than 100 ft (30 m) of the Person Formation was removed before deposition of the Georgetown Formation.

Georgetown Formation

The Edwards Formation and the Georgetown Formation are considered to be in hydraulic connection. Together they compose the Edwards and its associated limestone aquifer (Baker and others, in press). The Georgetown Formation consists mostly of thin, interbedded, fossiliferous, nodular, fine-grained limestone and marl. It ranges from 40 to 60 ft (12 to 18 m) thick.

Del Rio Clay

The Del Rio Clay, a selenitic, calcareous, pyritic, fossiliferous clay and marl, is about 75 ft (23 m) thick. The Del Rio Clay is the confining stratum for the Edwards aquifer. It crops out in the eastern part of the Balcones Fault Zone.

Depositional Environment of the Edwards Limestone

The depositional environment of the Edwards Limestone in Central Texas has been interpreted by Rose (1972) and Fisher and Rodda (1969). Inferred paleogeography is shown in figure 4. The Stuart City Reef (Winter, 1961) separated the shallow marine shelf that covered the interior of Texas from the deeper ancestral Gulf of Mexico Basin. Lower Cretaceous carbonates were deposited on the broad, essentially flat Comanche Shelf (Rose, 1972). This shallow-water shelf was partly bordered by the Tyler and Maverick Basins to the northeast and southwest. The area of lesser subsidence separating these two basins is the San Marcos Platform (Adkins, 1933). Deposition of the Edwards Limestone on the San Marcos Platform occurred in a dominantly supratidal and intertidal environment in which subtidal or lagoonal bathymetric depressions surrounded oyster and rudist grainstone bars (Abbott, 1975).

Hydrogeologic Development of the Edwards Limestone Aquifer

Conclusions by Abbott (1975, 1977) and Woodruff and Abbott (1979) regarding the development of the Edwards Formation into a major aquifer system in south-central Texas can be applied to the Austin area. Significant porosity and permeability in the limestone developed via dissolution by meteoric water while the Edwards Formation was being eroded prior to Georgetown Formation deposition. Removal of the Kirschberg Evaporite and other sabkha sediments and enlargement of collapse features created an aquifer in which ground water could move along fractures and enlarged bedding planes.

With the deposition of the Georgetown Formation, the Edwards Formation was buried and sediments accumulated sporadically through the Late Cretaceous (Gulf Epoch). The area now occupied by the eastern Edwards Plateau remained above sea level for the final time late in the Cretaceous Period as a result of regional upwarping of the northwestern margin of the subsiding Gulf of Mexico Basin. The aquifer system that developed in the Edwards was largely static because no discharge points existed that allowed a through-flowing ground-water system (Abbott, 1975).

The Edwards aquifer system as it now exists was greatly affected by Balcones faulting in the middle Miocene. Balcones faulting created significant topographic relief and caused incision of streams in response to a change in local base level. In addition, Balcones faulting produced a system of fractures and faults, many perpendicular to the dip of the Cretaceous strata. Along

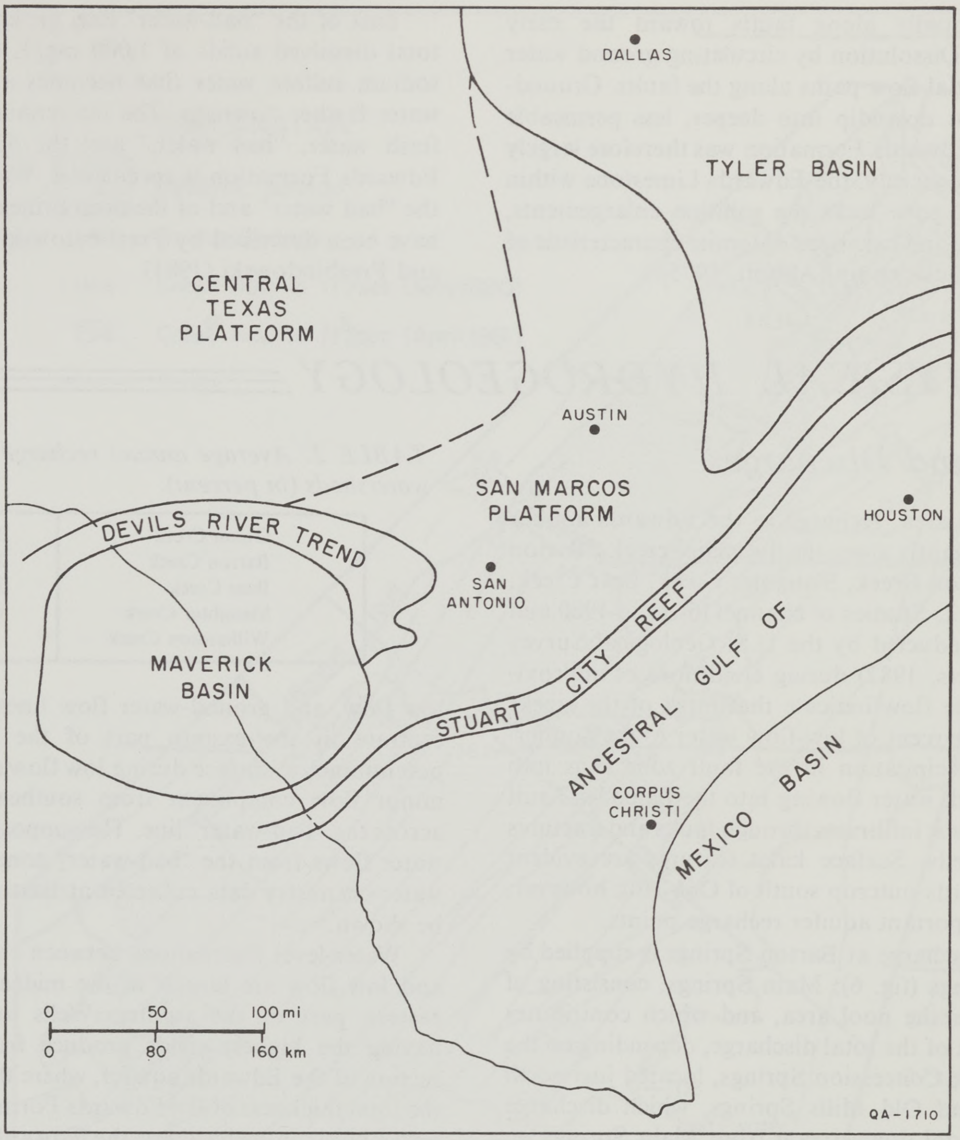

FIGURE 4. Regional elements of Texas during the Early Cretaceous. After Rose (1972).

these open fractures large amounts of ground water could move toward discharge points at lower elevations at the bottom of the incised stream valleys. After some discharge sites were established, a continuously circulating ground-water flow system developed. This early ground-water flow system enlarged significantly because of the "engrainment of the cavern system by meteoric water" (Abbott, 1975) circulating increasingly more recharge water toward the previously established discharge points.

Cavernous porosity was created not only along vertical fractures but also along bedding planes. St. Clair (1978) pointed out that most of the faults in the northwestern part of the area show displacements of less than 20 ft (6 m) and that this particular faulting probably resulted from collapse of rocks overlying the evaporitic beds of the Edwards Limestone. Abbott (1975) observed that many near-vertical fractures did not pass uninterrupted through thick sequences, indicating that the distribution of porosity in the Edwards Limestone is strongly controlled by bedding. The intensity of Balcones faulting, which created significant vertical-fracture porosity, increased downdip and toward Hays County where the fault displacements are greater than 100 ft (30 m) (Muehlberger and Kurie, 1956; Slade and others, in press).

The eastern boundary of fresh water of the Edwards aquifer is the "bad-water" line. Although the "bad-water" line is roughly parallel to the trend of the Balcones Fault Zone, it actually crosses faults and facies boundaries. Abbott (1975) interpreted the "bad-water" line to be a boundary that was not crossed by circulating ground water moving under structural or hydrologic controls. After creation of the Balcones Fault Zone, ground water moved preferentially along faults toward the early discharge sites. Dissolution by circulating ground water enlarged the initial flow paths along the faults. Ground-water movement downdip into deeper, less permeable sections of the Edwards Formation was therefore largely restricted. Consequently, the Edwards Limestone within the "bad-water" zone lacks the solution enlargements, recrystallization, and calcitized dolomite characteristic of the equivalent rocks updip (Abbott, 1975).

East of the "bad-water" line, ground water contains total dissolved solids of 1,000 mg/L or more and is a sodium sulfate water that becomes a sodium chloride water farther downdip. The interconnection among the fresh water, "bad water," and the deep brines in the Edwards Formation is speculative. Water chemistries of the "bad water" and of the deep brines in Central Texas have been described by Prezbindowski (1981) and Land and Prezbindowski (1981).

PHYSICAL HYDROGEOLOGY

Recharge and Discharge

In the study area, recharge to the Edwards aquifer occurs predominantly along the five major creeks: Barton Creek, Williamson Creek, Slaughter Creek, Bear Creek, and Onion Creek. Studies of channel losses in 1980 and 1981 (fig. 5) conducted by the U.S. Geological Survey (Slade and others, 1982) during conditions of approximate steady-state flow indicate that most of the creeks lose up to 100 percent of low-flow water to the aquifer. Most of the precipitation in the fault zone runs into the creeks. Creek water flowing into the Balcones Fault Zone from the west infiltrates through faults and fractures in the streambeds. Surface karst features are evident along the Edwards outcrop south of Oak Hill; however, they are not important aquifer recharge points.

The total discharge at Barton Springs is supplied by five major springs (fig. 6); Main Springs, consisting of three springs in the pool area, and which contributes 75 to 83 percent of the total discharge, depending on the amount of flow; Concession Springs, located just north of the pool; and Old Mills Springs, which discharge from a small pool downstream from Main Springs on the south bank of Barton Creek.

Slade and others (in press) estimated the total recharge to the aquifer (fig. 7). The contribution of each watershed is shown in table 2. Spring discharge and average annual pumpage (about 5 ft^3/sec) from the aquifer balance total recharge that occurs along the five major creeks (Slade and others, 1982).

TABLE 2. Average annual recharge from different watersheds (in percent).

Onion Creek	34
Barton Creek	28
Bear Creek	20
Slaughter Creek	12
Williamson Creek	6

Ground-Water Flow in the Aquifer

The pattern of ground-water flow can be inferred from the distribution of hydraulic head in the aquifer. Figures 8 and 9 show the potentiometric surfaces during high and low flow according to water-level measurements made during 1979 and 1981, and during 1978, respectively. Flow patterns inferred from the hydraulic head distribution suggest that during high flow the dominant flow direction is southwest to northeast toward Barton Springs. In contrast, the main flow component shifts to a south-to-north direction during conditions of low flow, and ground-water flow lines appear to concentrate in the eastern part of the fault zone. The potentiometric surface during low flow also documents a minor flow component from southeast to northwest across the "bad-water" line. The supposition that ground water flows from the "bad-water" zone is supported by water-chemistry data collected at Barton Springs,as will be shown.

Water-level fluctuations between conditions of high and low flow are largest in the mideastern and northeastern part of the aquifer. Wells in the study area having the highest yields produce from the confined section of the Edwards aquifer, where the wells penetrate the total thickness of the Edwards Formation. In general, water levels in wells along the Edwards outcrop to the west are relatively deep. Large yields are not obtained near the updip boundary of the aquifer (Smith, 1978). The Mt. Bonnell fault apparently is a barrier boundary marking the western limit of the aquifer.

Interaction between Aquifer and Springs

The change in potentiometric surface between high and low flow conditions is documented by individual water-level hydrographs from wells in the area (fig. 10). The Texas Department of Water Resources well numbering system was used in this report. Figure 11 shows that wells in the confined section of the aquifer display water-level fluctuations up to 90 ft (33 m). Moreover, these changes in water level correlate with changes in discharge of Barton Springs, suggesting an aquifer system with good hydrologic interconnection to Barton Springs. However, there are some exceptions: well 58-42-810,

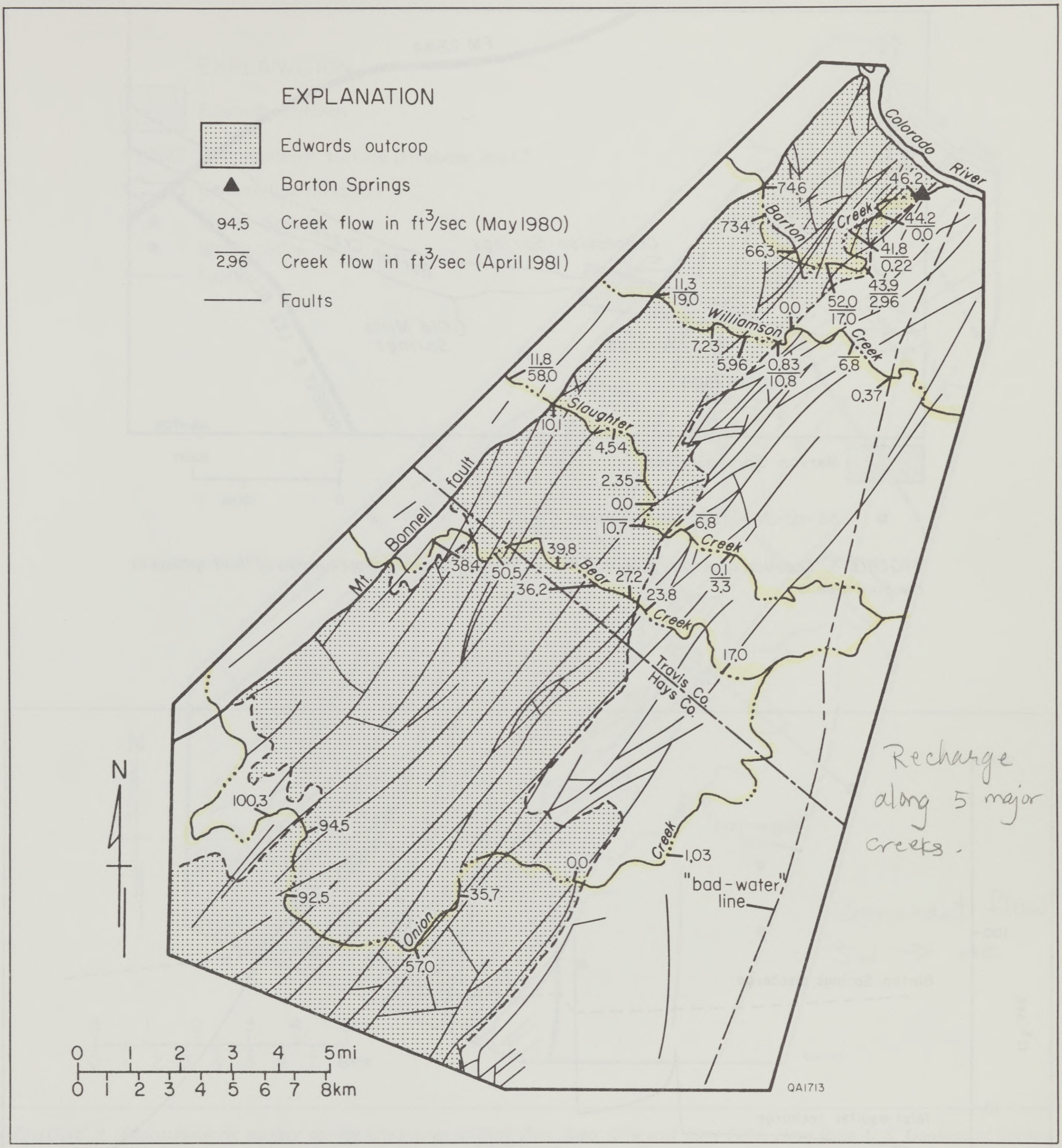

FIGURE 5. Measurements of stream flow showing channel losses in the Balcones Fault Zone. After Slade and others (1982).

which is located in the Rollingwood residential area to the west of the springs, shows no significant water-level variation and no correlation with changes in spring discharge. Also, water levels in well 58-42-913 did not show any significant changes during 1982 (Senger, 1983). This indicates that the main hydrologic connection within the aquifer is from the south and southwest to the northeast toward Barton Springs.

Well 58-50-301 is of interest because of its water-level fluctuation. This well is located just east of the "bad-water" line, where water has more than 1,000 mg/L total dissolved solids. Water-level variations exhibited by this

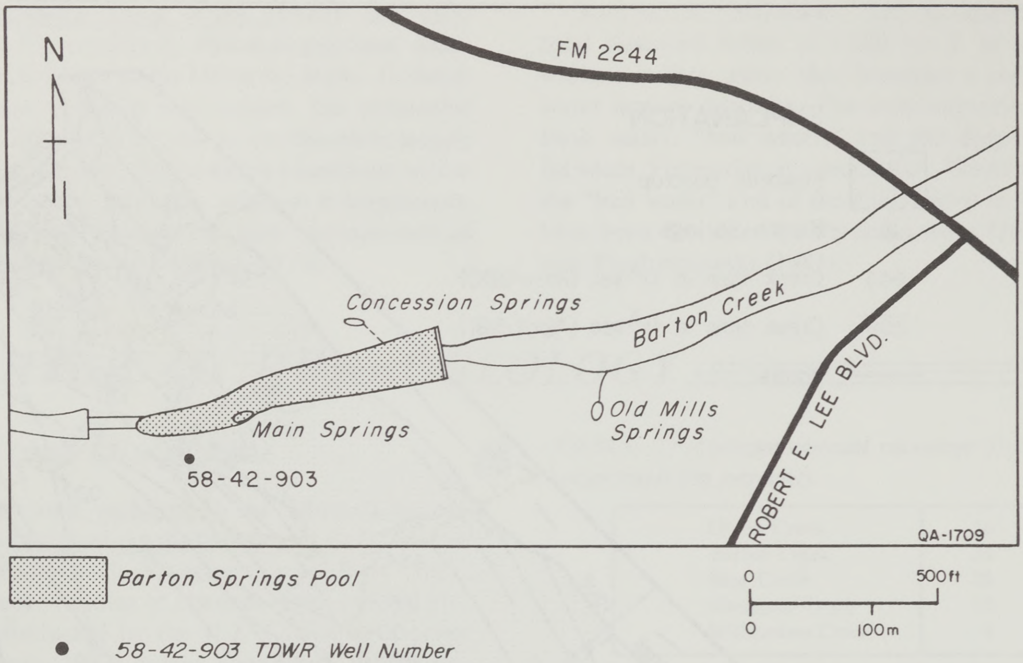

FIGURE 6. Location of major springs of Barton Springs. Main Springs consists of three springs in the pool area.

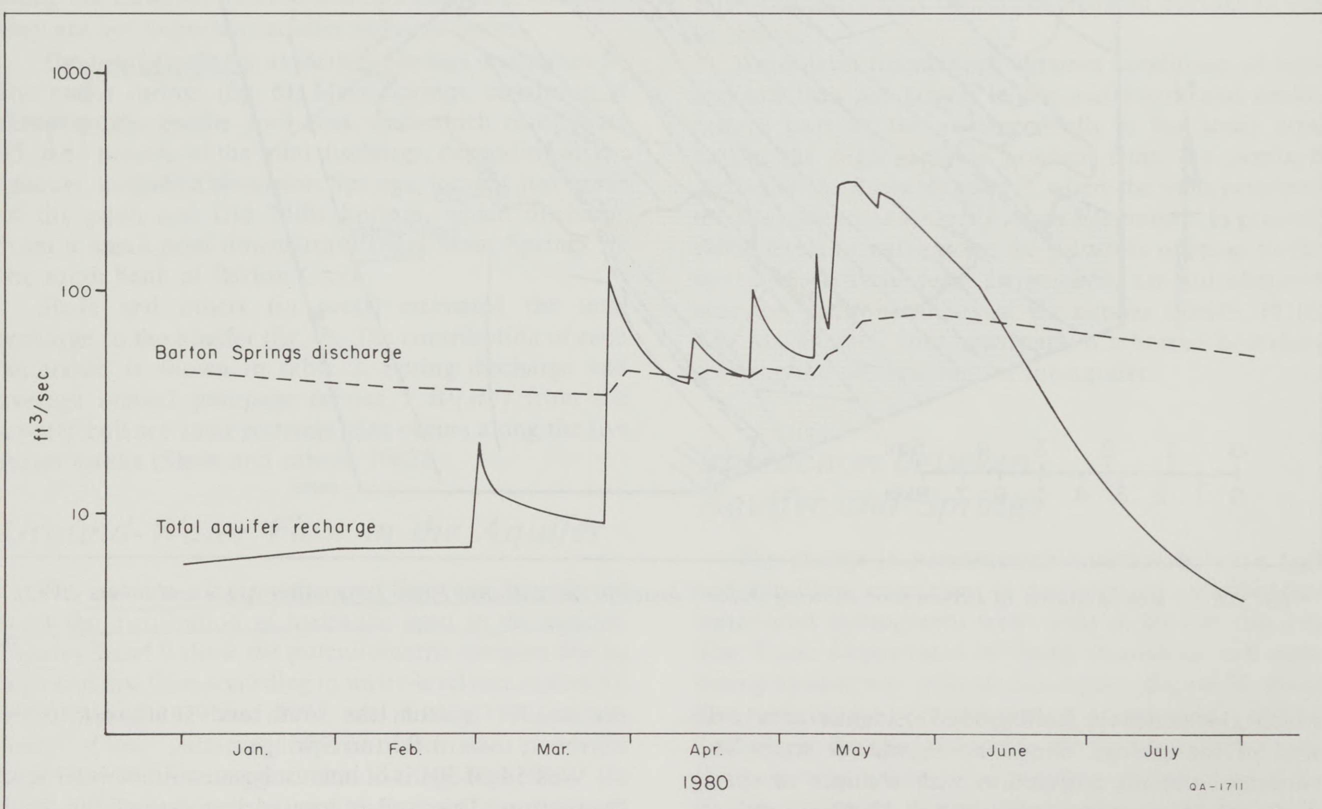

FIGURE 7. Total recharge to the aquifer compared with total discharge in Barton Springs. Data from Slade and others (in press). Note that flow rates on y-axis are in logarithmic scale.

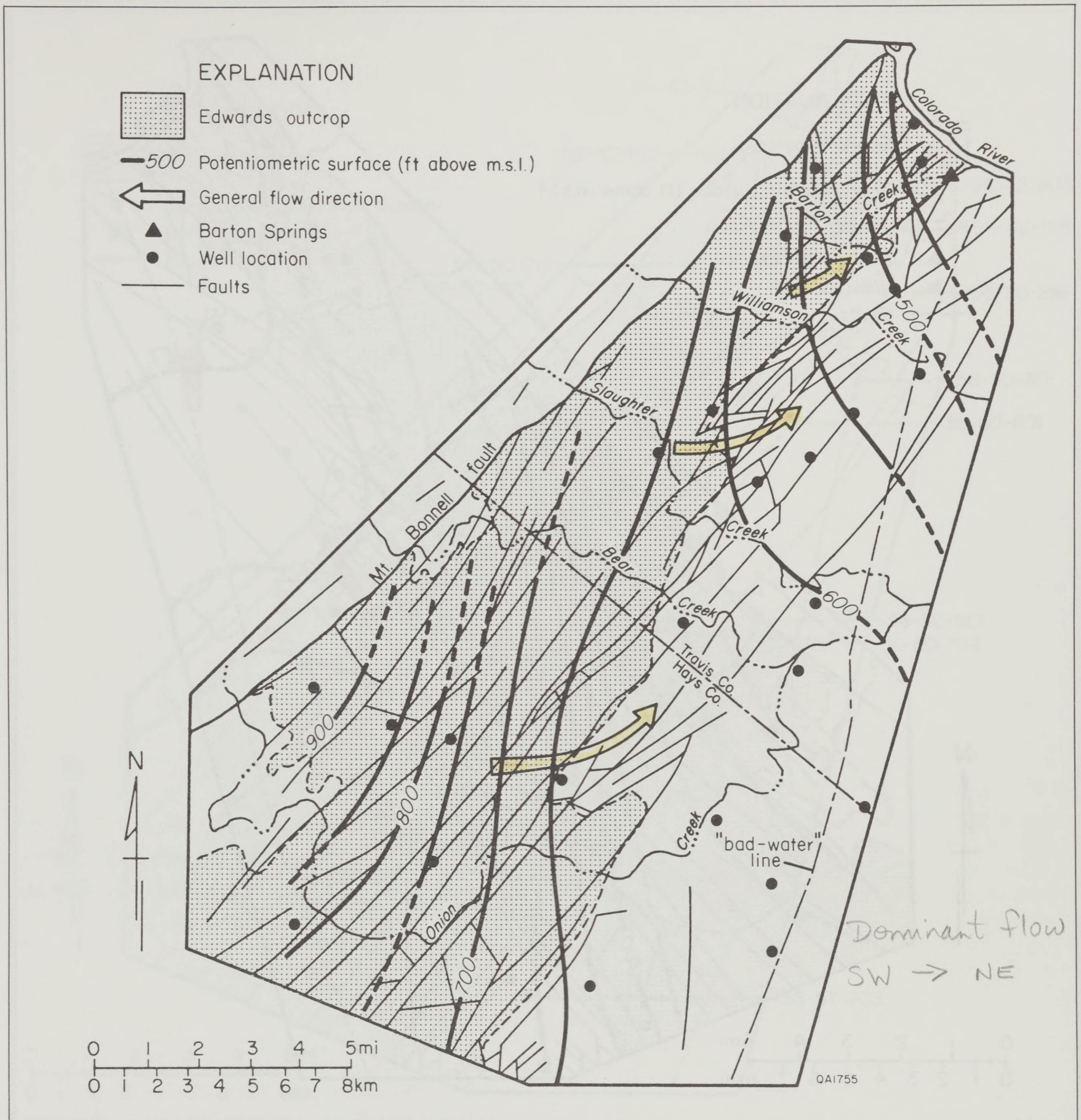

FIGURE 8. Potentiometric surface during conditions of high flow, June 1979 and June 1981. Data from U.S. Geological Survey, Austin.

well were as high as 50 ft (15 m) during 1979 and 1980. Moreover, changes in water level in this well correlate with changes in spring discharge. This indicates that a hydraulic connection exists between the "bad-water" zone and the main fresh-water aquifer.

Because of the close correlation between water-level changes in the aquifer and changes in spring discharge, the water level in well 58-42-903, located 200 ft (70 m) from the main spring outlet, has been monitored continuously with an automatic water-level recorder to measure the total discharge at Barton Springs.

Figure 12 shows a good correlation between the water level in well 58-42-903 and the total discharge from Barton Springs. Changes in water level in the well

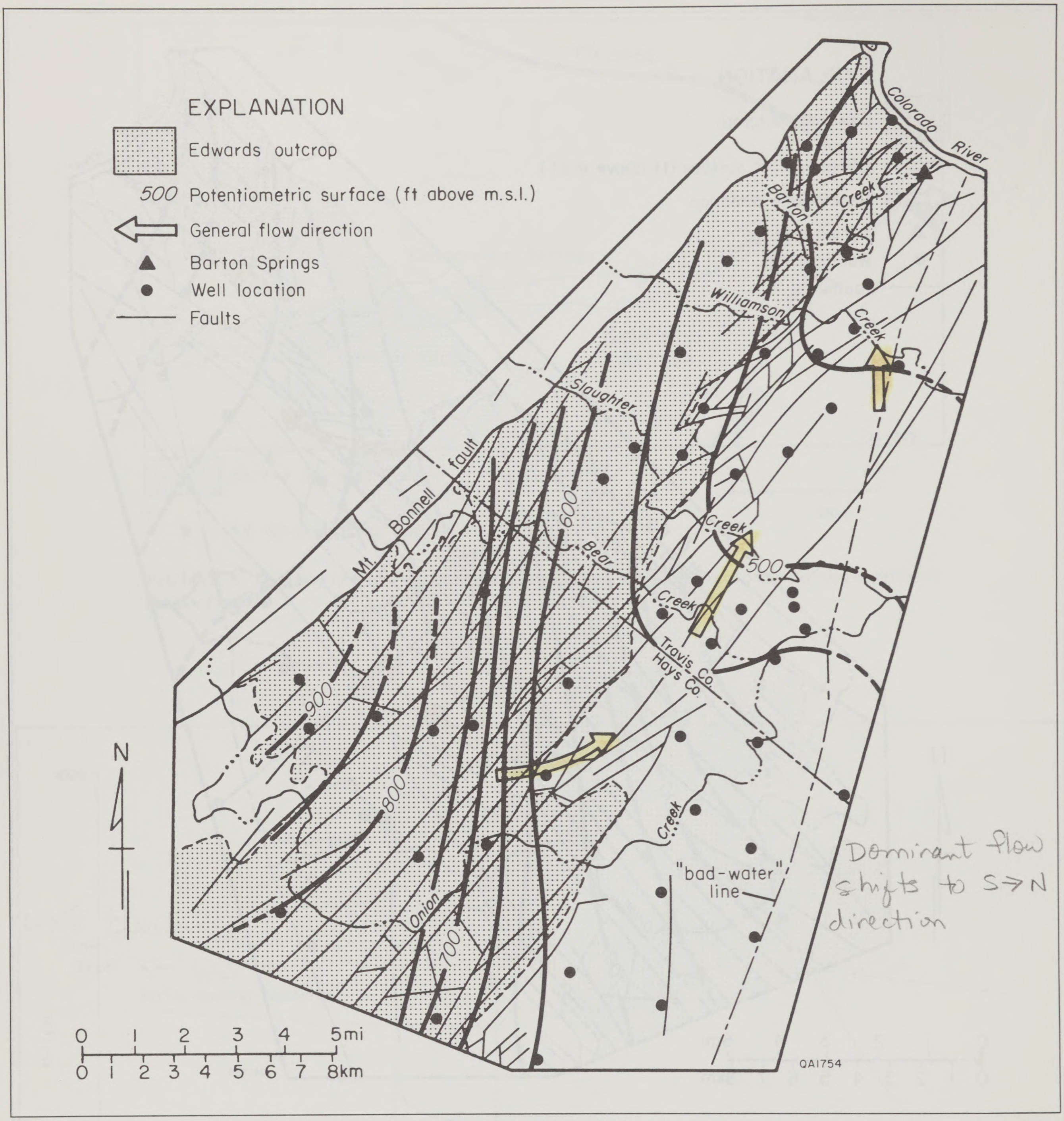

FIGURE 9. Potentiometric surface during low-flow conditions, August 1978. Data from U.S. Geological Survey, Austin.

correspond to changes in water level in the pool. Water-level changes also correlate with the total spring flow. The limited discharge measurements indicate that discharge is higher when the pool is drained than when the pool is filled.

The water-level decline in the aquifer caused by draining the pool can also be observed in well 58-42-915. This well is located about 1 mi (1.7 km) southwest of Barton Springs. Figure 13 shows a sharp response in water level approximately 30 min after the drain gates of the pool are opened; it takes about 30 min to drain the pool. The water level in the pool drops to between 3 and 4 ft (about 1 m) when the pool is drained.

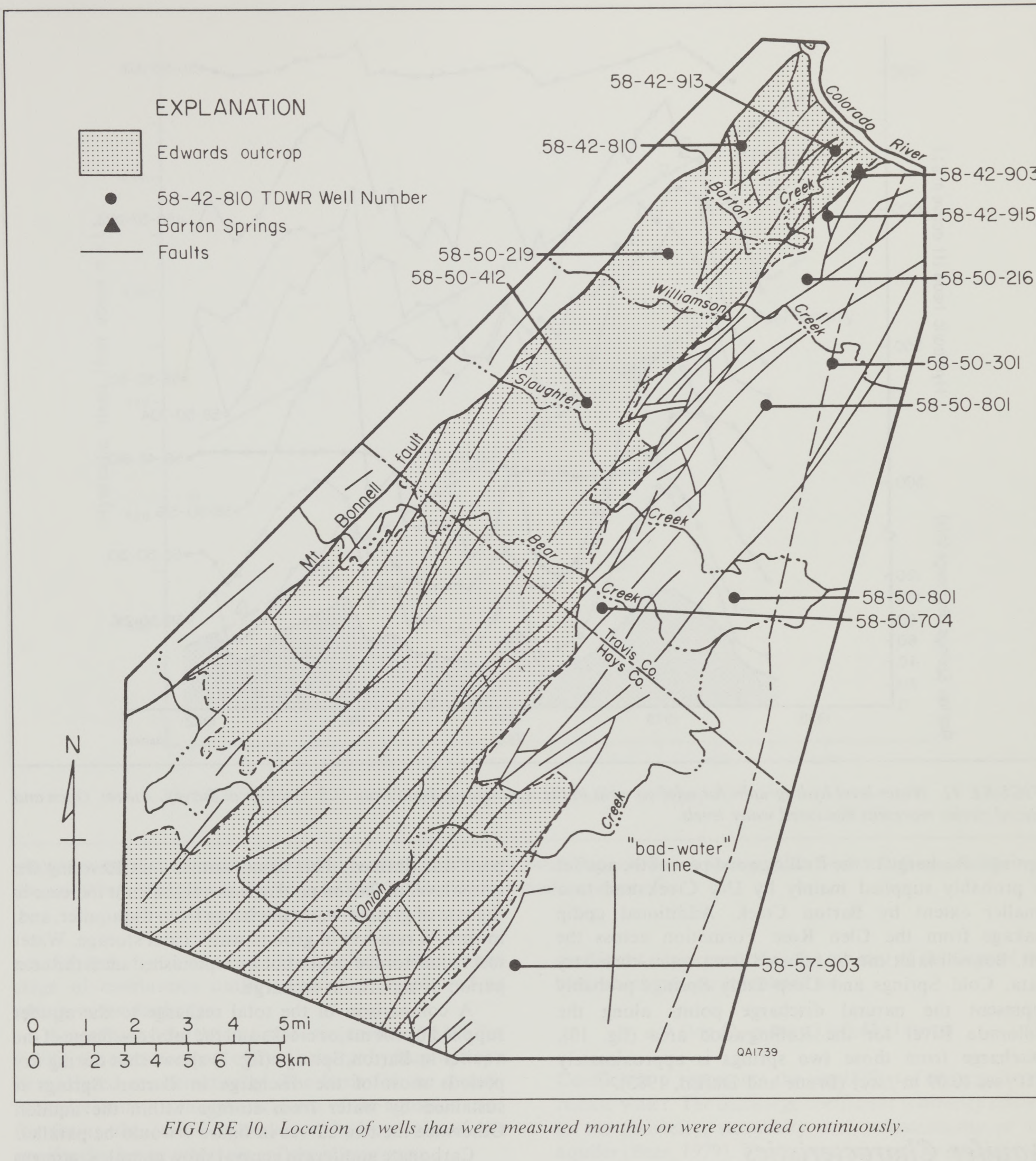

FIGURE 10. Location of wells that were measured monthly or were recorded continuously.

During conditions of relatively low flow, the water-level decline in the aquifer caused by draining the pool can be recognized in water-level records of well 58-50-216, which is located about 2.7 mi (4.5 km) southwest of Barton Springs (fig. 14). In contrast, the water level in well 58-42-913, located about 0.6 mi (1 km) northwest of the springs in the Rollingwood area (fig. 10), shows no response to pool draining. These contrary water-level responses indicate that the dominant hydrologic connection between the springs and the aquifer is south and southwest of Barton Springs, corresponding to the general direction of the Balcones faulting.

The part of the Edwards aquifer that is in the Rollingwood area appears to be isolated from Barton

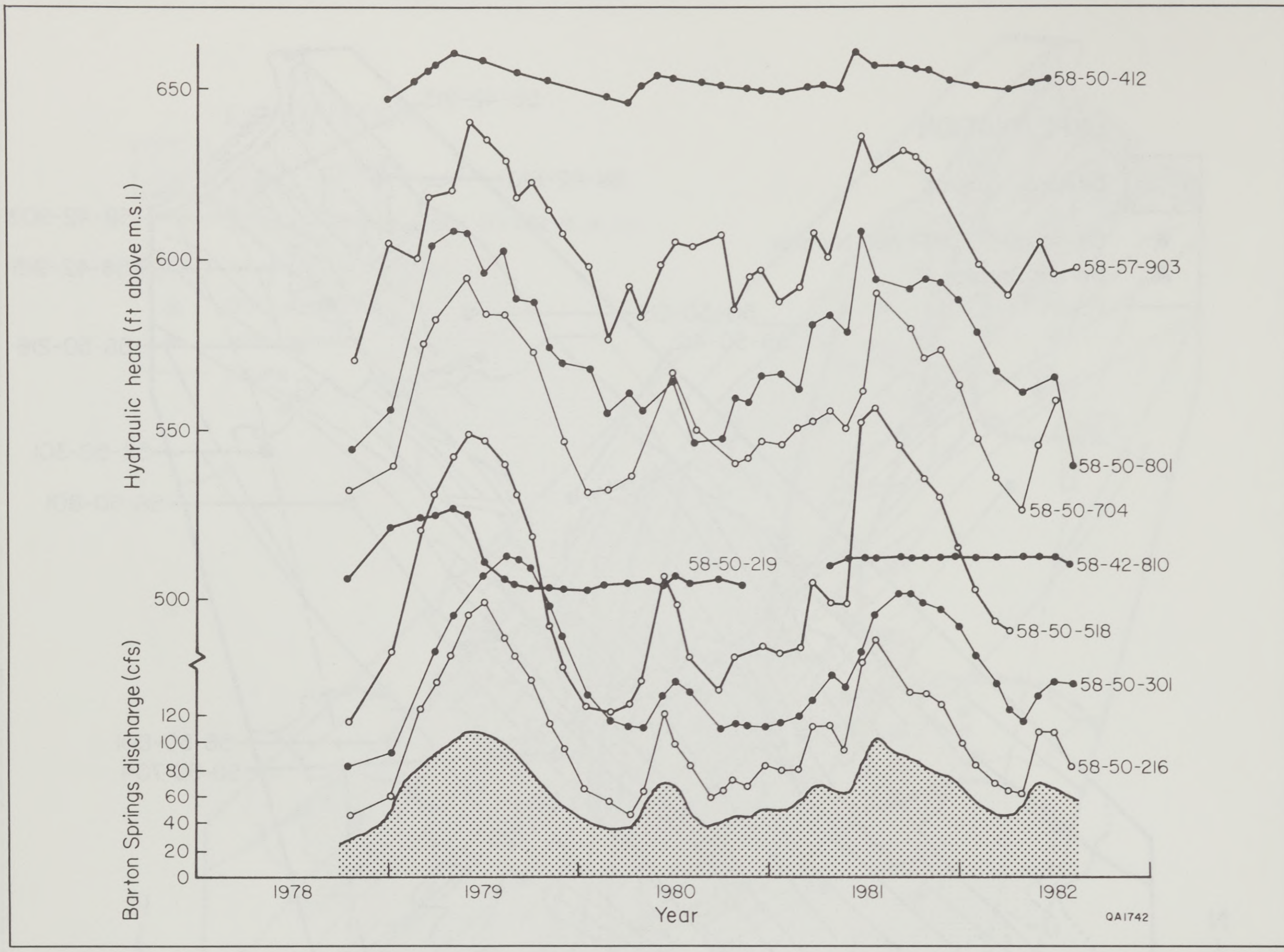

FIGURE 11. Water-level hydrographs for selected wells in the Austin area. Data from U.S. Geological Survey, Austin. Open and closed circles represent measured water levels.

Springs. Recharge to the Rollingwood part of the aquifer is probably supplied mainly by Dry Creek and to a smaller extent by Barton Creek. Additional updip leakage from the Glen Rose Formation across the Mt. Bonnell fault can be inferred from water-chemistry data. Cold Springs and Deep Eddy Springs probably represent the natural discharge points along the Colorado River for the Rollingwood area (fig. 10). Discharge from those two springs is approximately 3 ft^3/sec (0.09 m^3/sec) (Brune and Duffin, 1983).

Aquifer Characteristics

The water-level response in wells 58-42-915 and 58-50-216, as shown in figures 13 and 14, reflects an interesting characteristic of the aquifer. After Barton Springs pool was refilled, the water level in well 58-42-915 did not recover to the expected higher water level that existed before the pool was drained. A similar response occurred at well 58-50-216 during low-flow conditions, where the water level decreased more rapidly when the pool was drained. This demonstrates that lowering the water level in Barton pool causes a significant increase in the rate of ground-water discharge from the aquifer, and, in turn, a removal of ground water from storage. Water lost from storage might not be replenished until the next period of significant recharge.

A comparison of the total recharge to the aquifer supplied by the major creeks and the total discharge of the aquifer at Barton Springs (fig. 7) shows that during dry periods most of the discharge in Barton Springs is sustained by water from storage within the aquifer. Otherwise the two curves in figure 7 would be parallel.

Carbonate aquifers in general show complex patterns of ground-water flow because of their heterogeneity and anisotropy. It is difficult to assign hydrologic parameters to a karst aquifer on the basis of limited results of pumping tests. However, with regard to the aquifer characteristics described previously, the recession-curve analysis of discharge variation and water-level declines can be used to obtain certain quantitative information about the aquifer. The most suitable time for discharge measurement is during relatively dry periods when

ϕ variations

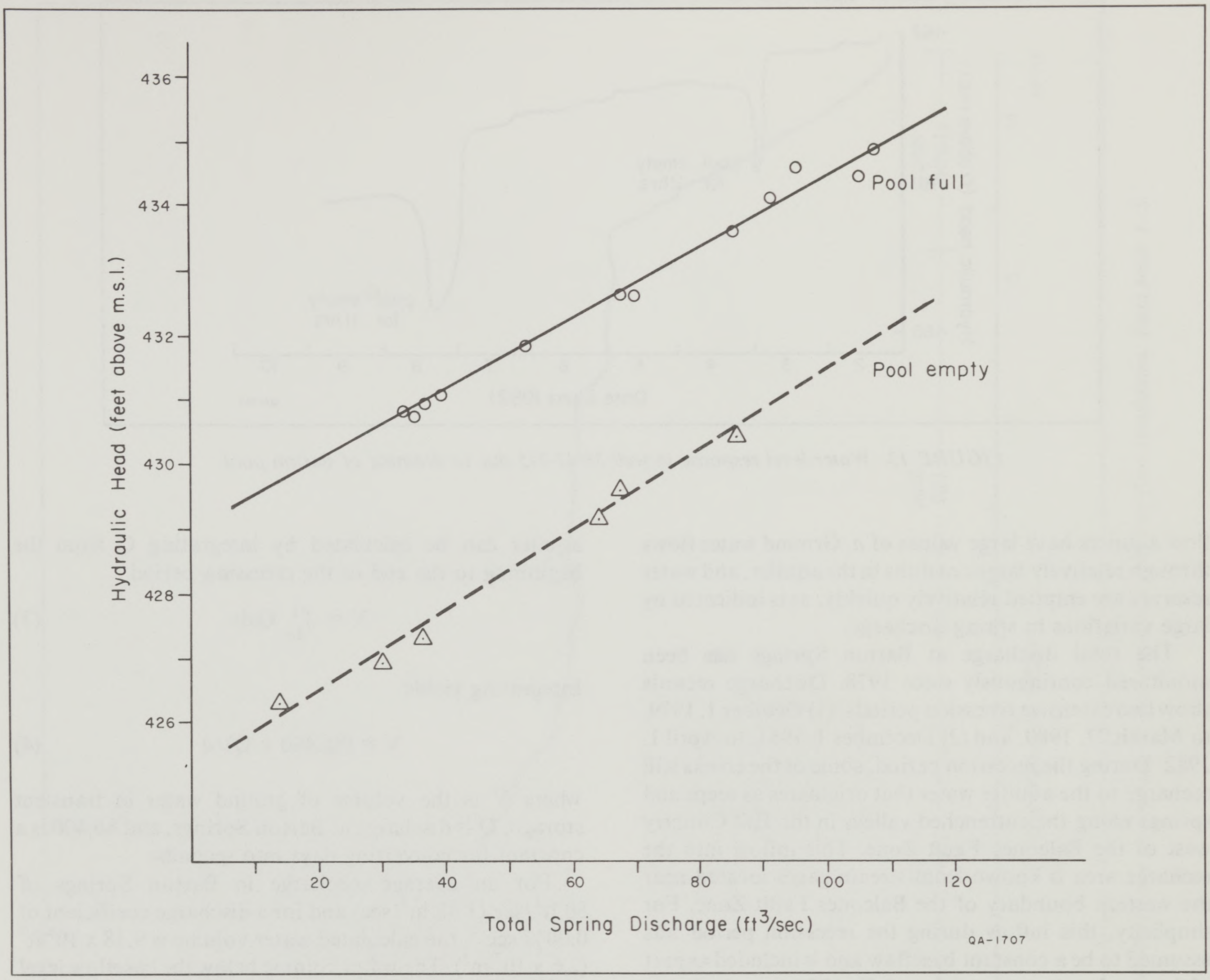

FIGURE 12. Correlation of water levels in well 58-42-903 and total discharge in Barton Springs. Data from U.S. Geological Survey, Austin. Water level measured when pool was full ○ and when pool was empty △.

aquifer recharge is minimal. The aquifer is then in the stage of continuous outflow, which is monitored as spring-water discharge.

According to the theoretical basis provided by Maillet (1905), the recession part of the discharge hydrograph can be analyzed mathematically. The basis for the quantitative treatment is the general form of the equation used for fitting the recession curves of hydrographs of the aquifer discharge when the inflow is near zero

$$Q(t) = Q_0 e^{-a (t-t_0)} \quad (1)$$

where, according to figure 15, Q(t) is the spring discharge (m^3/sec) during the period t_0 to t, Q_0 is the spring discharge at the initial time t_0, and a is the discharge coefficient (sec^{-1}).

The recession part of the hydrograph curve is a straight line on a semilogarithmic scale. The discharge coefficient a is expressed as the tangent of the slope of the line, which can be calculated as

$$a = (\log Q_0 - \log Q_t)/0.4343\ (t-t_0) \quad (2)$$

Coefficient a represents the capability of the aquifer to release water. The discharge coefficient is directly related to the geometry, storativity, and transmissivity of the aquifer (Bear, 1979). These properties can be investigated by analyzing the hydrographs of spring discharge. In general, the value of a decreases as the underground resistance to flow increases. In a carbonate aquifer with a small value of a, ground water flows through small, interconnected solution openings, fractures, and intergranular pores. A narrow range of spring-flow variations indicates that the water reserves are emptied slowly, a phenomenon which is referred to as a diffuse-flow aquifer (Thrailkill, 1978). In contrast, concentrated-

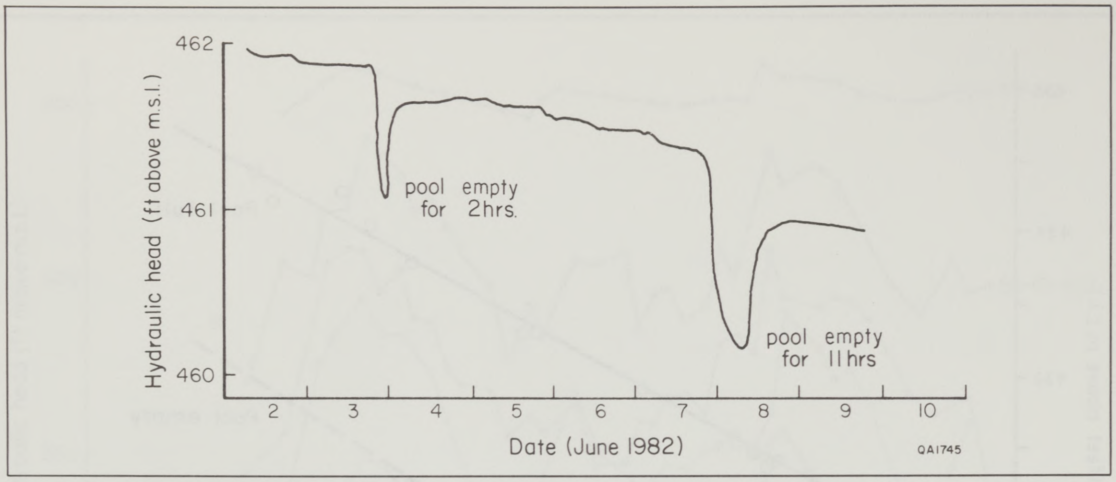

FIGURE 13. Water-level response in well 58-42-915 due to draining of Barton pool.

flow aquifers have large values of *a*. Ground water flows through relatively large conduits in the aquifer, and water reserves are emptied relatively quickly, as is indicated by large variations in spring discharge.

The total discharge at Barton Springs has been monitored continuously since 1978. Discharge records show two extensive recession periods: (1) October 1, 1979, to March 27, 1980, and (2) December 1, 1981, to April 1, 1982. During the recession period, some of the creeks still recharge to the aquifer water that originates as seeps and springs along the entrenched valleys in the Hill Country west of the Balcones Fault Zone. This inflow into the recharge area is known from stream gages located near the western boundary of the Balcones Fault Zone. For simplicity, this inflow during the recession period was assumed to be a constant baseflow and is included as part of the regulated reserve of the aquifer. The mean recharge during the first and second recession periods was about 7 ft^3/sec and 23 ft^3/sec (0.2 m^3/sec and 0.65 m^3/sec), respectively (R. M. Slade, personal communication, 1983). Compared with the average annual pumpage from the aquifer of about 5 ft^3/sec (0.14 m^3/sec) (Brune and Duffin, 1983), baseflow recharge during the first recession period appears to be negligible.

The discharge coefficient *a* for the first recession period yielded a value of 0.0047 sec^{-1}, suggesting a diffuse-flow aquifer with overall Darcian ground-water flow velocities (that is, velocity is proportional to the hydraulic gradient). Near Barton Springs, however, the flow lines converge and the flow velocities increase, eventually approaching inertial flow conditions (Senger, 1983). In this case, the Darcian velocity is no longer directly proportional to the hydraulic gradient.

Water Volume

According to Torbarov (1978), the volume of ground water in transient storage (above the baseflow level) in the aquifer can be calculated by integrating Q from the beginning to the end of the recession period

$$V = \int_{t_0}^{t} Q_t dt \qquad (3)$$

Integrating yields

$$V = (86{,}400 \times Q)/a \qquad (4)$$

where V is the volume of ground water in transient storage, Q is discharge at Barton Springs, and 86,400 is a constant for converting days into seconds.

For an average discharge in Barton Springs of 50 ft^3/sec (1.42 m^3/sec) and for a discharge coefficient of 0.0047 sec^{-1}, the calculated water volume is 9.18×10^8 ft^3 (2.6×10^7 m^3). The water volume below the baseflow level is substantial, as is indicated by a minimum discharge of 34 ft^3/sec (0.96 m^3/sec) at Barton Springs at the end of the first reces-sion period. In comparison, Slade and others (in press) estimated the saturated volume in the Edwards aquifer above the base-level elevation of Barton Springs, based on the potentiometric surface during average flow conditions, as amounting to nearly 8.83×10^{11} ft^3 (2.5×10^{10} m^3). Given the average storativity of 0.0075 (Senger, 1983) and Slade's data on saturated volume of limestone, the total volume of water in the aquifer amounts to 4.59×10^8 ft^3 (1.3×10^8 m^3). With respect to available water resources, this estimate is probably too high given the wide variation in average annual discharge. After the prolonged drought, the annual spring flow in 1956 was less than 15 ft^3/sec (0.43 m^3/sec) at Barton Springs, compared with the long-term average annual spring flow of about 50 ft^3/sec (1.42 m^3/sec).

The volume of saturated limestone between water levels in the aquifer, occurring when Barton Springs flow is 50 ft^3/sec and 34 ft^3/sec, is only about 4 percent of the total volume of saturated limestone between the elevation

FIGURE 14. Water-level response in well 58-50-216 due to draining of Barton pool during low-flow conditions. Data from U.S. Geological Survey, Austin.

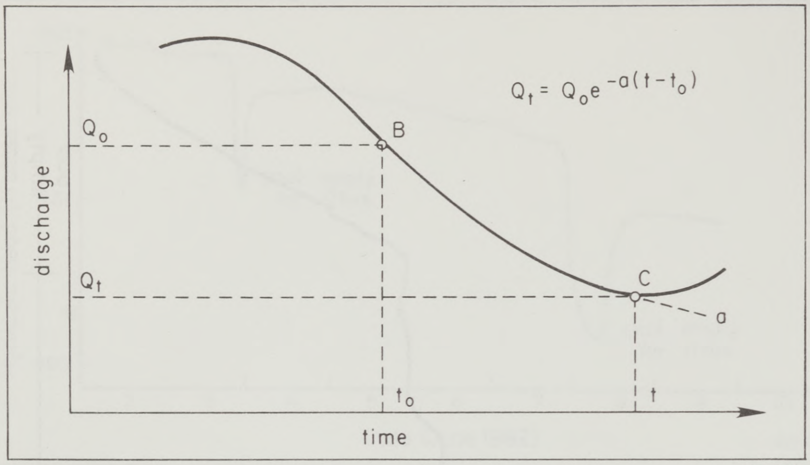

FIGURE 15. Part of the hydrograph with the recession curve that is analyzed. After Milanovic (1981).

of Barton Springs pool (435 ft, 131 m) and the water-level elevation in the aquifer during average flow conditions when Barton Springs flows at about 50 ft^3/sec (R. M. Slade, personal communication, 1983). The volume of ground water in transient storage (2.6 x 10^7 m^3) when spring discharge is 50 ft^3/sec composes about 10 percent of the total volume of ground water in the aquifer (2.5 x 10^8 m^3), assuming an overall storativity of 0.0075 (Senger, 1983). The difference between the fraction of water volume and the fraction of saturated rock volume above the baseflow level suggests that storativity below the baseflow level is probably much lower than the average storativity value of 0.0075. The method for calculating storativities (Senger, 1983) assumes that they are representative only for the part of the aquifer above the baseflow level.

Aquifer Parameters

Two important hydrogeological characteristics of aquifers are transmissivity and storativity. In porous aquifers, these parameters are obtained by pumping tests. In carbonate aquifers having dominantly fracture flow, hydrogeologic data obtained from pumping tests are in general unrepresentative of large parts of the aquifer. Brune and Duffin (1983) reported transmissivities for the Edwards aquifer that were based on pumping test results as ranging from 400 to 300,000 gal/d/ft (5.8 x 10^{-5} to 4 x 10^{-2} m^2/sec).

In this study, overall aquifer parameters were estimated using two different approaches. First, transmissivities and storativities were computed by analyzing the hydrographs of selected wells during the recession period. This is similar to analyzing water-level declines in monitoring wells during a pumping test, whereby the springs act as the pumped well. Similarly, an average storativity of the aquifer was obtained by comparing the outflow from the aquifer and the average decrease of water levels during the recession period (Torbarov, 1978; Milanovic, 1981). Second, the hydrologic parameters of the aquifer near the springs were estimated by calibrating a two-dimensional transient ground-water flow model, implemented with the computer program FLUMP (Narasimhan and others, 1978). Input data for the model included information about water-level fluctuations and discharge at Barton Springs. Calculated transmissivities and storativities using the recession-curve analysis range from 0.1 m^2/sec to 0.4 m^2/sec and 0.001 to 0.023, respectively (table 3). The numerical simulation of the transient water-level response in well 58-42-915 yielded a transmissivity of 0.2 m^2/sec (fig. 16). Storativity in the model, however, was 0.00075, which is one order of magnitude lower than the estimated average storativity (0.0075) of the entire aquifer (Senger, 1983).

The application and results of both methods are discussed in more detail by Senger (1983). The major assumption of the ground-water flow model—that Darcian flow conditions predominate in the aquifer—is

TABLE 3. Transmissivities and storativities obtained from recession-curve analysis of water-level declines in selected wells during 1979 and 1980.

WELL	TRANSMISSIVITY (m^2/sec)	STORATIVITY
58-50-216	0.17	0.023
58-50-219	0.40	0.001
58-50-301	0.10	0.012
58-50-518	0.14	0.003
58-50-704	0.14	0.001
58-50-801	0.14	0.003

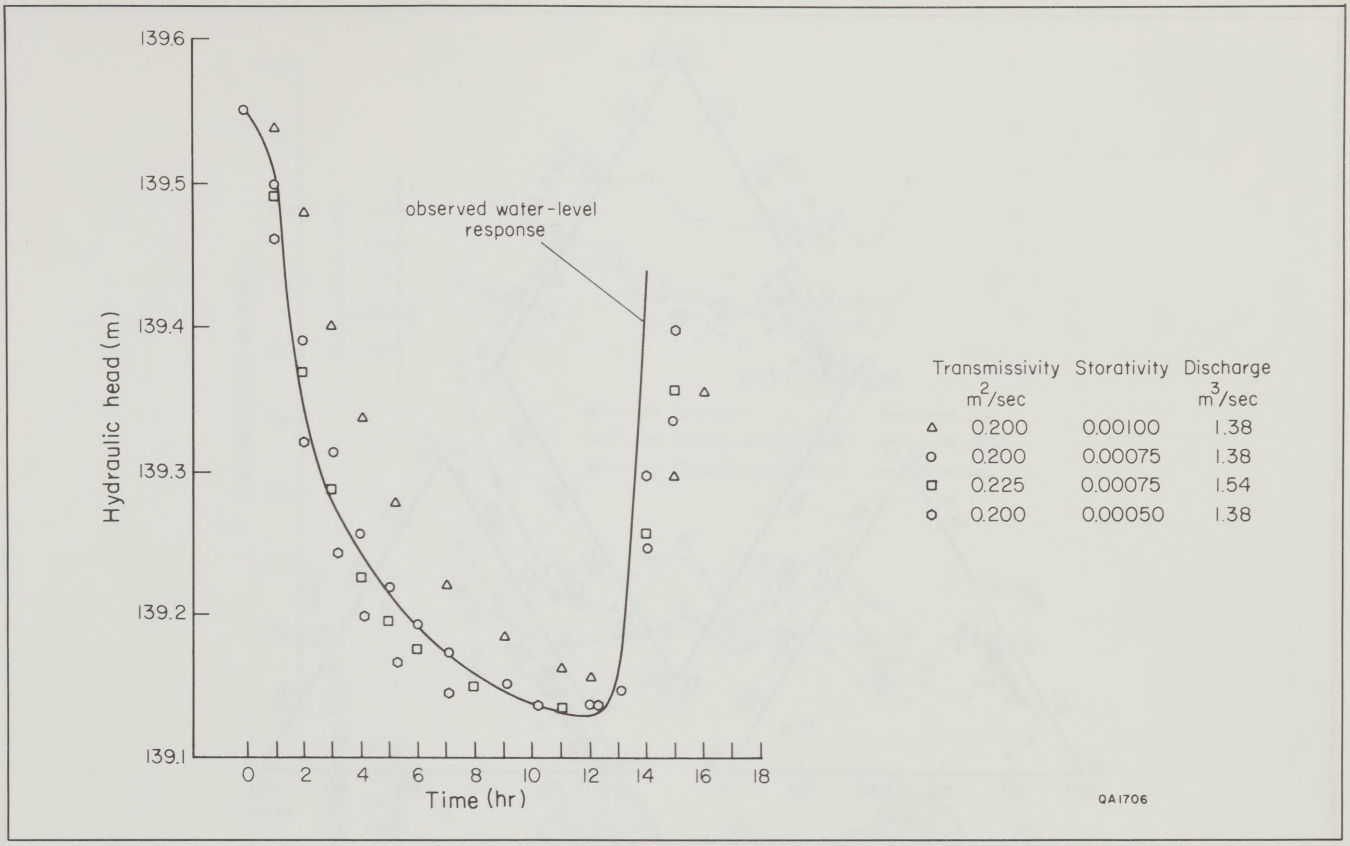

FIGURE 16. Computed hydraulic heads for well 58-42-915 (isotropic conditions) compared with the observed water level.

supported by the low discharge coefficient. Furthermore, the model reproduced with acceptable accuracy the observed discharge at Barton Springs and the transient water-level response in well 58-42-915 (fig. 16), and it verified the overall transmissivities obtained from the water-level declines in various wells in the aquifer (table 3). Storativity in the model, however, was lower by approximately one order of magnitude. This could be attributed to the simplicity of the model, as discussed by Senger (1983).

HYDROCHEMISTRY

The Edwards Limestone aquifer contains calcium bicarbonate water, and in some areas calcium magnesium bicarbonate water, that becomes sodium sulfate water downdip. Farther downdip, Edwards ground water becomes sodium chloride water (fig. 17).

Barton Springs

Chemical analyses of waters from Barton Springs show conspicuous variation in values under varying flow conditions. Figure 18 shows the results of chemical analyses for the period 1978 to 1981 and indicates an increase in sodium, chloride, sulfate, and magnesium with decreasing spring flow. Sodium and chloride exhibit the largest fluctuation and increase exponentially during low flow (fig. 19). St. Clair (1978) attributed this increase of sodium and chloride during conditions of low flow to an influx of Lake Austin water having relatively high concentrations of sodium and chloride. As mentioned previously, there is no hydraulic connection between Barton Springs and the Rollingwood area (or Lake Austin) to the west. Although water-chemistry data display a trend of Barton Springs water toward that of

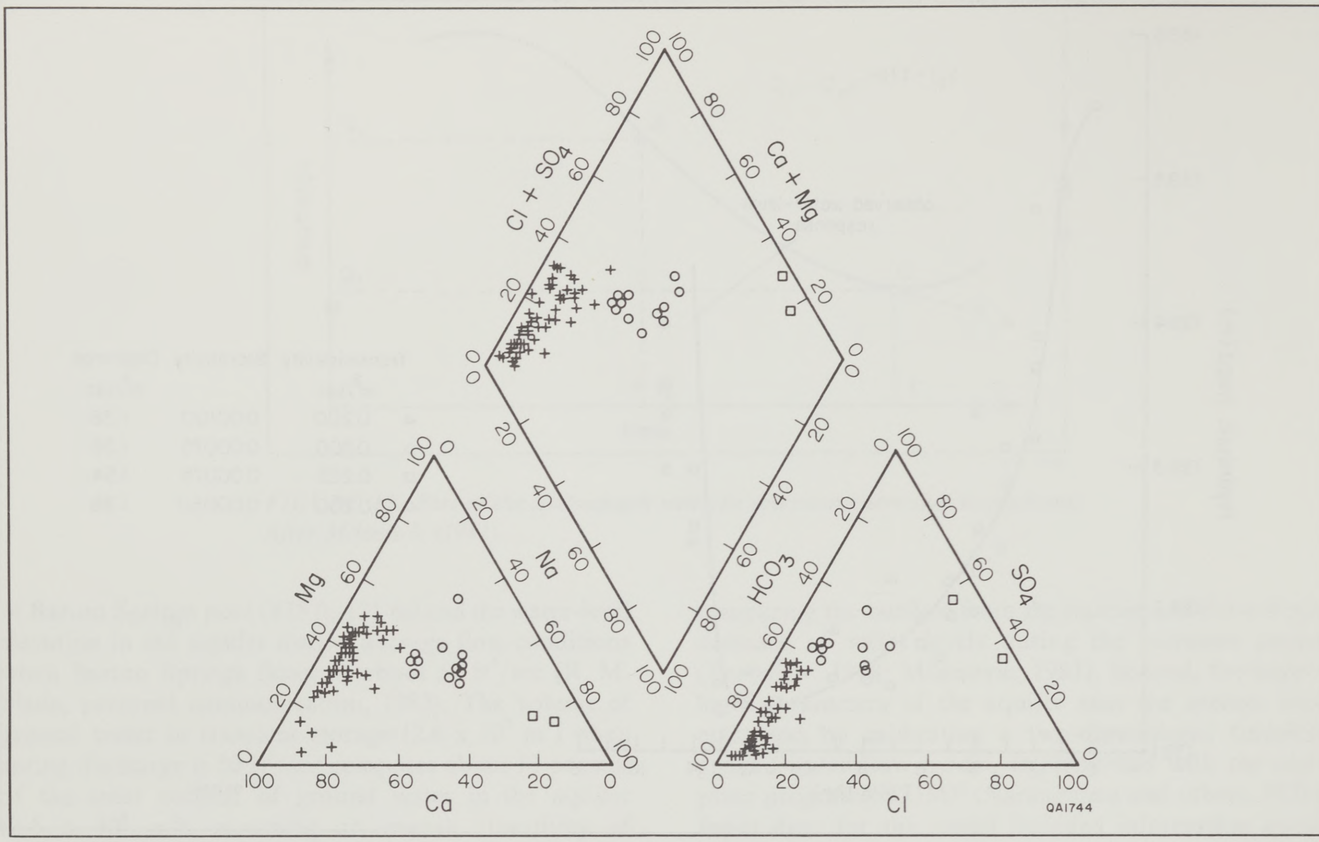

FIGURE 17. Trilinear diagram of water-chemistry analyses from the Edwards aquifer. Data from U.S. Geological Survey, Austin. Water samples collected from wells located updip the "bad-water" line: + ; in the vicinity of the "bad-water" line: ○ ; and downdip from the "bad-water" line: □ .

Lake Austin water (fig. 20), this trend can be explained by influx of water from the "bad-water" zone, as indicated by the potentiometric surface during low flow.

Strontium concentrations in samples collected in the summer of 1982 during decreasing flow conditions are high in Barton Springs waters compared with concentrations in Lake Austin. Figure 21 shows a distinct trend of the water chemistry in Barton Springs from a composition typical of ground water in the Edwards outcrop area toward a composition similar to "deep" Edwards aquifer water from wells near the "bad-water" line. Water from Lake Austin and ground water in the Rollingwood area contain lower strontium concentrations; consequently, these waters cannot account for the relatively high strontium concentrations in the spring waters.

In general, most wells in the study area show little variation in water chemistry through time, except for well 58-50-216 located about 2.8 mi (4.5 km) southwest of Barton Springs (fig. 10). A water sample from well 58-50-216 collected after a relatively dry summer in 1982 showed a water composition similar to that of "bad water": TDS content was more than 1,000 mg/L (Senger, 1983). This composition suggests not only that there is a hydraulic connection between the fresh-water aquifer system and the "bad-water" zone, but also that during low flow there is a significant encroachment of high-TDS water into the main flow of the aquifer supplying Barton Springs. "Bad water" supplied to Barton Springs was estimated to be about 5 to 10 percent when the springs flowed at 20 ft^3/sec (0.6 m^3/sec) during the relatively dry period in the summer of 1978 (Senger, 1983).

The "Bad-Water" Zone

Hydraulic head distribution in the aquifer during conditions of low flow (fig. 9) indicates a minor hydraulic gradient from southeast to northwest across the "bad-water" line. The interconnection between the "bad-water" zone and the main aquifer body is also suggested by the water-level fluctuations of well 58-50-301 located just east of the "bad-water" line (fig. 11).

Chemical data from the "bad-water" zone are limited, and no information exists about seasonal variations in the chemistry of these waters. Chemical data on waters from well 58-50-301 in 1948 and 1949 show large differences in TDS that may be related to differences in the overall flow conditions of the aquifer. A water sample from well 58-50-301 in October 1948 contained 8,870 mg/L of total dissolved solids, whereas Barton Springs discharge was

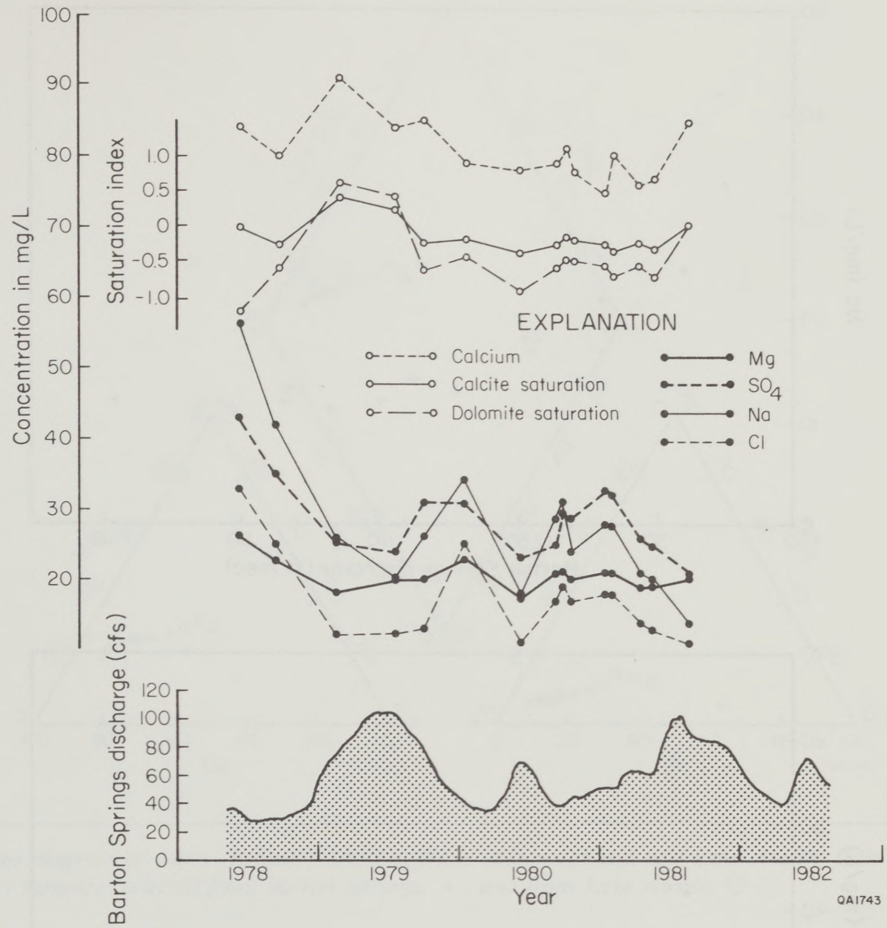

FIGURE 18. Chemical composition of water from Barton Springs during varying discharge.

about 20 ft^3/sec (0.6 m^3/sec). In July 1949, when the springs discharged at about 50 ft^3/sec (1.42 m^3/sec), TDS decreased to 1,479 mg/L. This relationship suggests that there are large fluctuations in water levels as well as significant variations in water chemistry at locations in the "bad-water" zone.

In the San Antonio area, water from the "bad-water" zone has a highly variable TDS content (1,150 to 4,300 ppm). Prezbindowski (1981) explained the water chemistry as being controlled by two processes: (1) mixing of fresh water from the Edwards aquifer moving downdip into the basin with deep saline waters moving up and out of the basin; and (2) dissolution of the Edwards Limestone by undersaturated ground water moving downdip. The stable isotope composition of "bad water" (Prezbindowski, 1981) indicates that water from the zone is predominantly meteoric and originated as recharge updip. The chemical composition is probably controlled by the lithology of the rocks and perhaps by mixing with deep brines (Longman and Mench, 1978).

Edwards Aquifer

The chemical composition of ground water in the aquifer grades downdip from a calcium bicarbonate and calcium magnesium water in the recharge area to a sodium sulfate water and finally to a sodium chloride water deep within the basin (fig. 17). However, relatively high sulfate concentrations also exist in the updip part of the aquifer (fig. 22). In addition, the plot of strontium concentrations versus sodium concentrations in samples from the area (fig. 21) outlines three different types of waters: (1) water with low concentrations of strontium

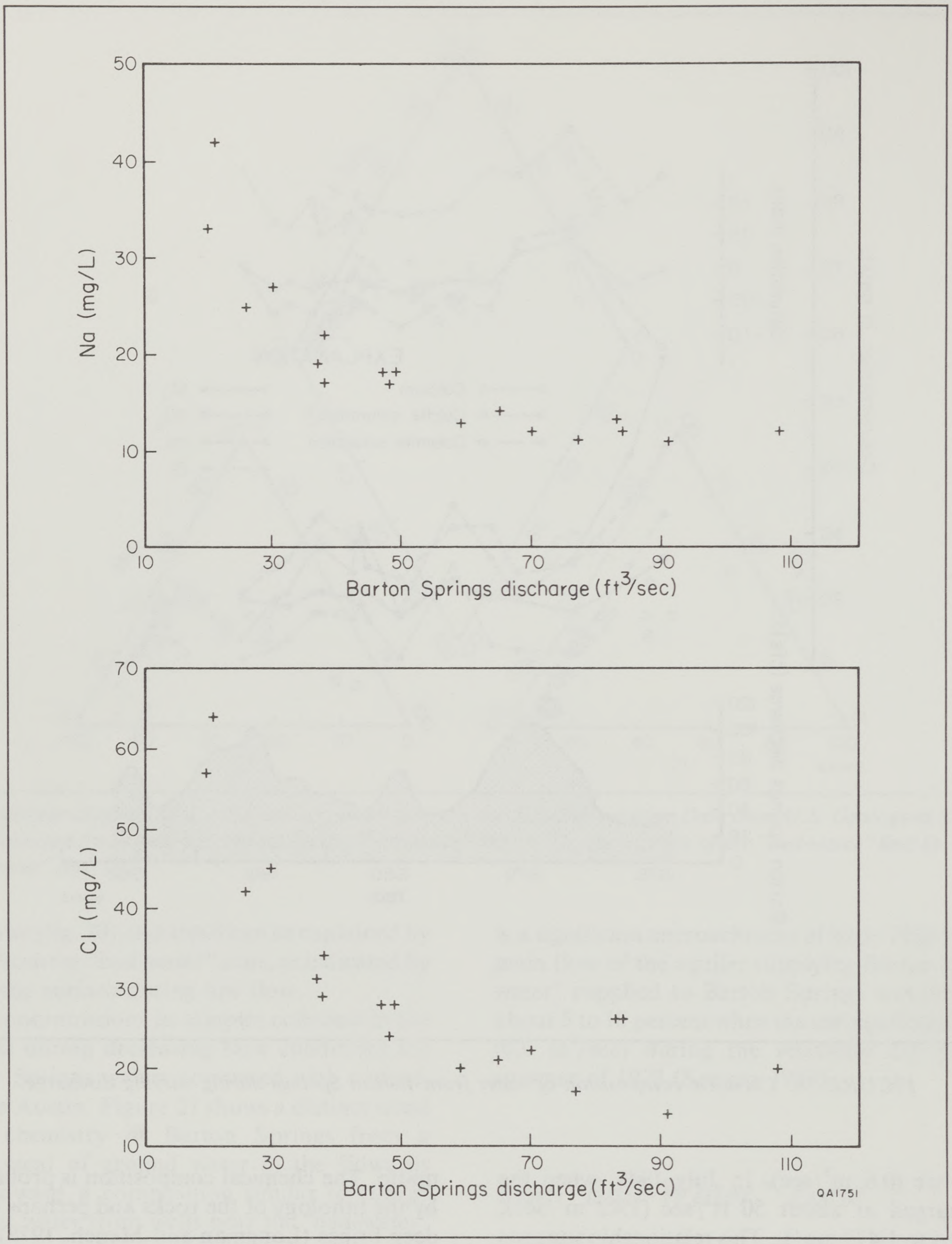

FIGURE 19. Increase of sodium and chloride concentrations in Barton Springs water during decreasing discharge.

and sodium (recharge water), (2) water with high concentrations of strontium and sodium in "deep" Edwards water, and (3) water with high strontium concentrations but relatively low sodium concentrations.

Leakage of water from the Glen Rose Formation may control the presence of water with high strontium-sodium ratios. Water chemistry of the Glen Rose Formation is shown diagrammatically in figure 23. Most samples, however, are from wells west of the Mt. Bonnell fault and outside the fault zone. No data on strontium concentrations were obtained from Glen Rose wells within the Balcones Fault Zone. However, celestite ($SrSO_4$) nodules in the Glen Rose Formation (Rodda and others, 1970) indicate that a high strontium content in ground water there is likely. Figure 24 shows the location of the wells in the area where waters were sampled. The areal distribution of strontium concentrations suggests the following:

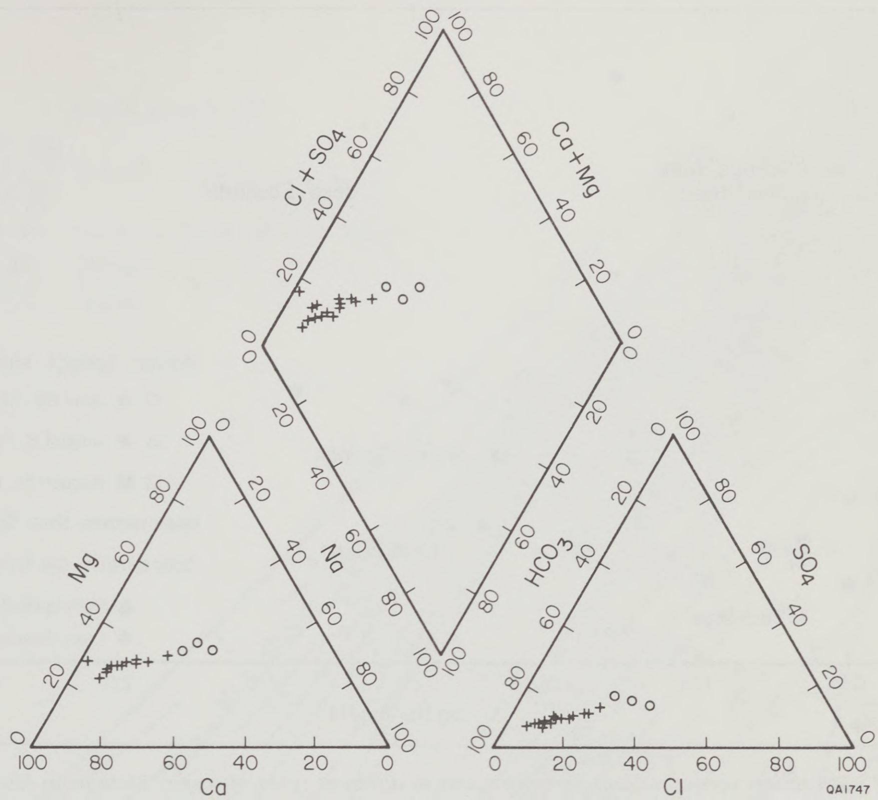

FIGURE 20. Trilinear diagram of water chemistry analyses from Barton Springs and Lake Austin. Data from U.S. Geological Survey, Austin. Water samples collected from Barton Springs; +, and from Lake Austin; ○.

- Ground water with low concentrations of strontium and sodium exists in the Edwards outcrop area.
- Ground water with high concentrations of strontium and sodium exists in wells nearest the "bad-water" line and represents "deep" Edwards water.
- Edwards ground water that is affected by leakage from the Glen Rose Formation is high in strontium but low in sodium and exists in wells mainly in the southeastern part of the area. A well with the highest strontium concentration is located in the Rollingwood area just east of the Mt. Bonnell fault.

This last group of waters affected by leakage from the Glen Rose Formation contains relatively high sulfate concentrations, which is also more typical of Glen Rose water. Further evidence of leakage from the Glen Rose Formation into the Edwards aquifer is demonstrated by figure 25, wherein the molar sulfate to chloride ratio is plotted versus sulfate concentrations. The plot also delineates the different types of waters in the study area as follows:

- Typical "recharge" water has low concentrations of chloride and sulfate.
- "Deep" Edwards water is characterized by a sodium sulfate water that becomes a sodium chloride water farther downdip.
- Glen Rose water contains high concentrations of sulfate. Sulfate to chloride ratios increase with increasing sulfate concentrations.
- Chemistry of ground water affected by leakage from the Glen Rose Formation is similar to the water chemistry that plots at the intersection of the composition trends displayed by the previously discussed types of waters (fig. 25).

Waters affected by leakage from the Glen Rose Formation are mainly from wells located in the eastern part of the fault zone, but some wells are on the Edwards outcrop just east of the Mt. Bonnell fault.

Leakage from the Glen Rose Formation is probably not upward through the Walnut Formation into the Edwards Formation but instead is lateral across fault surfaces (fig. 26). Where there are large fault displacements, the Edwards Formation is in contact with updip Glen Rose strata. In general, the largest displacements occur along the Mt. Bonnell fault and in the eastern part of the study area in Hays County and southeastern Travis County. Edwards ground water east of the Mt. Bonnell

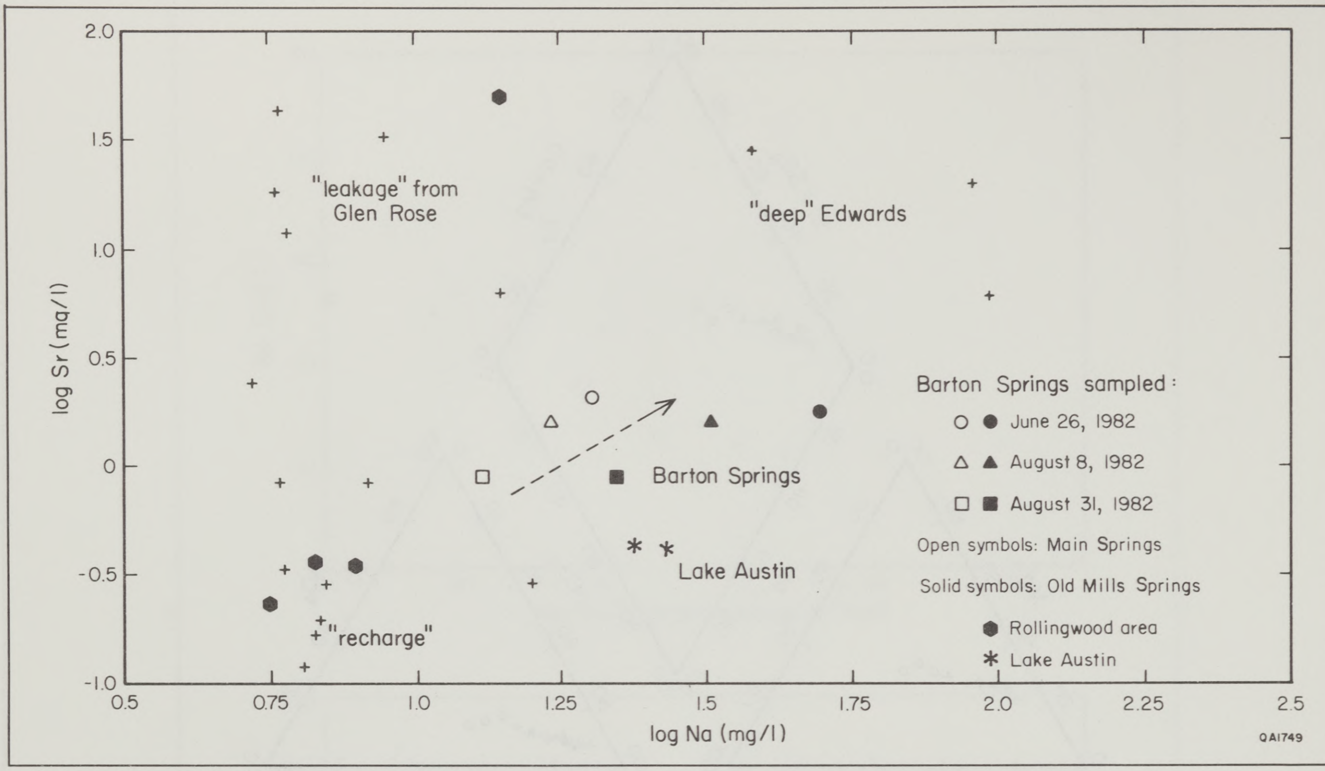

FIGURE 21. Strontium versus sodium concentrations in different types of water. Data from Senger (1983).

fault and in the southeastern part of the study area also displays typically high concentrations of sulfate and strontium. In the eastern part of the study area, however, the water chemistry is more complicated because of proximity to the "bad-water" zone farther to the east, where high concentrations of strontium and sulfate also exist. But in contrast, "deep" Edwards water contains more sodium than does the updip area.

Carbonate Equilibria

Development of a carbonate aquifer depends on the geologic setting of the host rock and on the saturation state of the ground water with respect to minerals within the carbonate rocks. The saturation state of ground water indicates if limestone dissolution is possible and would increase the porosity of the aquifer. The dominant minerals in limestones of the Edwards Formation are calcite and dolomite.

Carbonate equilibria of various water samples from the area investigated were calculated by the computer program SOLMNEQ developed by Kharaka and Barnes (1973). The program is designed to calculate solution speciation and saturation states of the aqueous phase with respect to various mineral phases, given analytical concentrations of the elements, pH, and temperature. The program computes the equilibrium distribution of various chemical species in a solution and compares the activity products of various combinations of these dissolved species with the theoretical equilibrium constants that would exist were the waters in equilibrium with various solid mineral phases. The saturation state of a particular water is given as

$$SI = \log (AP/KT) \qquad (5)$$

where SI is the saturation index, AP is the activity product of the solution, and KT is the equilibrium solubility product of the species at the temperature of the water.

Chemical analyses of waters to be used for carbonate equilibrium calculations were restricted to those samples for which pH, temperature, and alkalinity were measured in the field. Further, samples with questionable pH measurements or other doubtful results were eliminated. The accuracy of the saturation states is affected largely by the accuracy of the pH measurement; an error of 0.1 pH unit translates into an error of 0.1 unit in saturation index for calcite (Pearson and Rettman, 1976).

Ground water from the Edwards aquifer shows a wide variation in carbonate saturation (fig. 27). Saturation index values for calcite range from −0.724 to +0.560 with an arithmetic mean of −0.101. Dolomite saturation varies from −1.462 to +0.950 with a mean of −0.166.

Most samples from the Edwards aquifer were collected in the summers of 1978 through 1981; thus, the

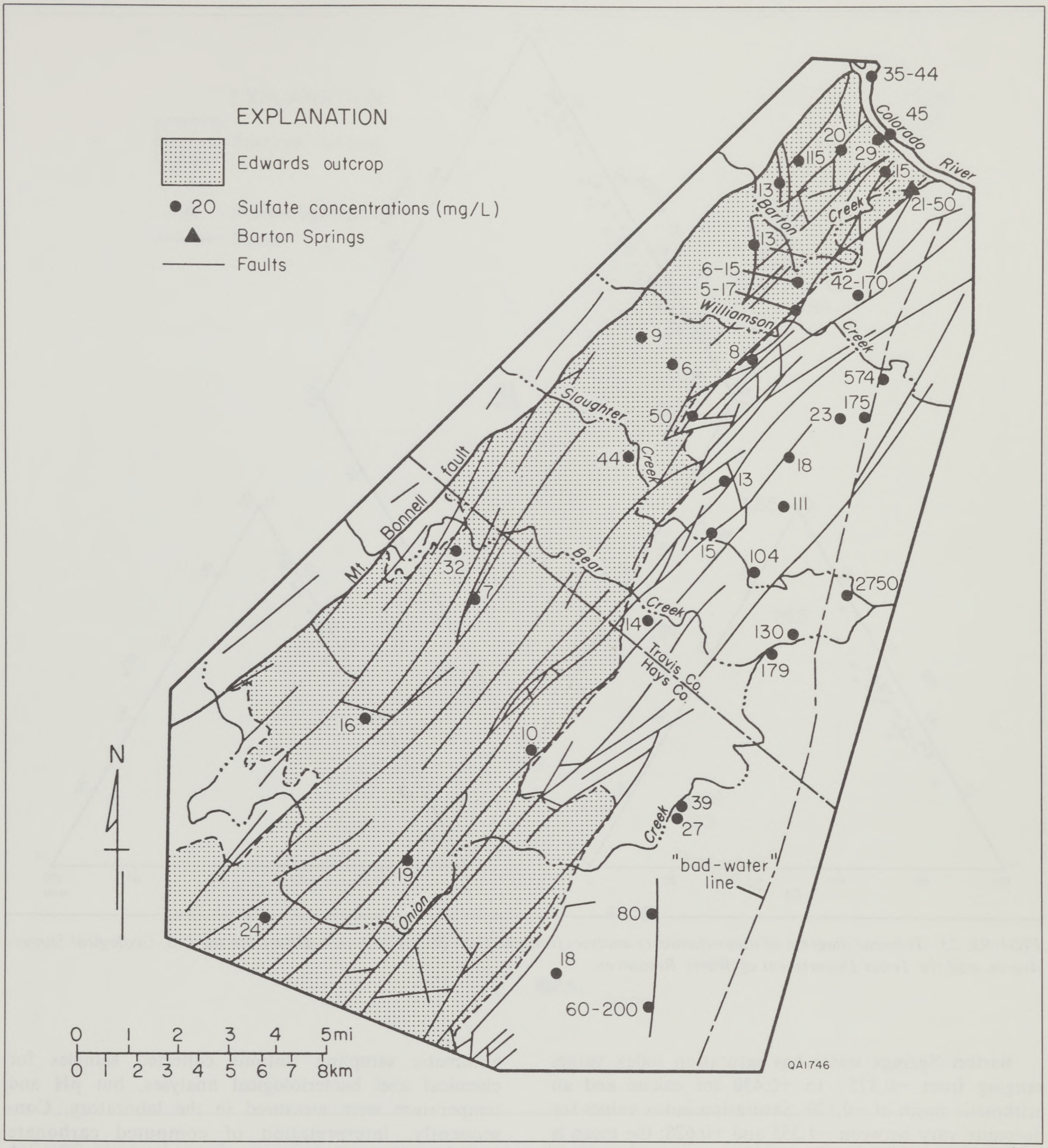

FIGURE 22. Areal distribution of sulfate concentration in the Edwards aquifer.

wide range of carbonate equilibrium values cannot be explained by seasonal variations. Comparing saturation states of waters collected during high flow (1979, 1981) and low flow (1978, 1980) also indicates no significant correlation. The areal distribution of saturation indices shows that ground water along the Edwards outcrop is predominantly undersaturated with respect to calcite and dolomite. Saturation indices of ground water in the confined part of the aquifer do not show a trend of varying saturation indices with flow direction.

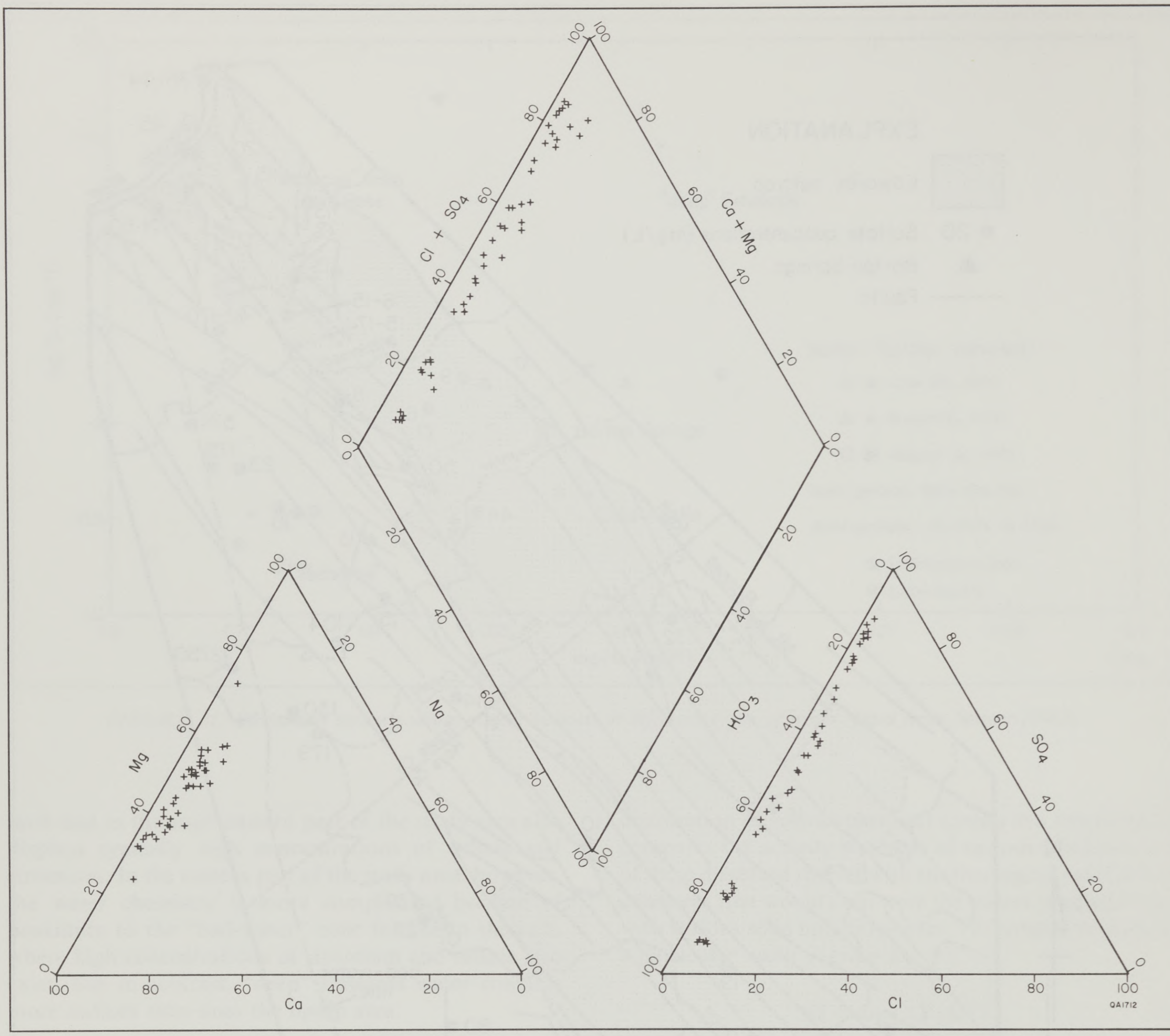

FIGURE 23. Trilinear diagram of water-chemistry analyses from the Glen Rose Formation. Data from the U.S. Geological Survey, Austin, and the Texas Department of Water Resources.

Barton Springs water has saturation index values ranging from −0.375 to +0.430 for calcite and an arithmetic mean of −0.136. Saturation index values for dolomite vary between −1.355 and +0.628; the mean is −0.459 (fig. 28). Barton Springs water is predominantly undersaturated with respect to calcite and dolomite. The data do not suggest seasonal variations, but saturated water in the springs occurs at times of highest discharge from the springs (1979 and 1981).

Carbonate equilibrium values for water samples from creeks indicate that saturation exists with respect to calcite and dolomite. Saturation index values for calcite and dolomite range from 0.192 to 1.088 and from 0.229 to 2.066, respectively (fig. 29). During floods in the creeks, automatic sampling stations collected samples for chemical and bacteriological analyses, but pH and temperature were measured in the laboratory. Consequently, interpretation of computed carbonate equilibria of these samples is limited. The chemical composition of most of these samples, however, indicates that flood waters are undersaturated as computed by SOLMNEQ (Senger, 1983). Barton Creek, which has the highest flow rate of all the creeks during floods, still contains saturated water. The overall water chemistry of Barton Creek during floods differs from the chemistry of recharge water in the other creeks sampled where the creeks flow into the fault zone. Flood water from Barton Creek flowing into the recharge zone east of

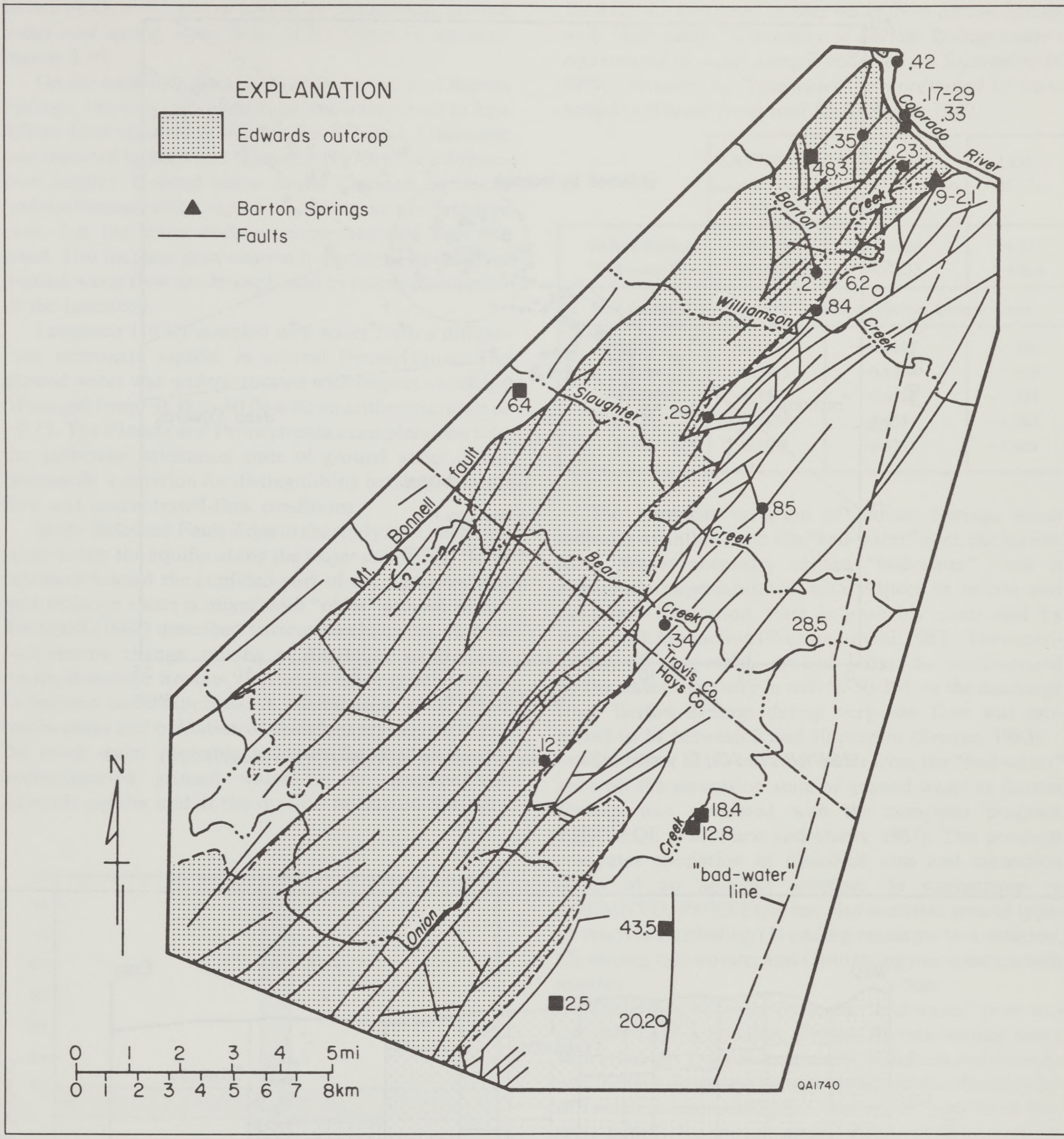

FIGURE 24. Areal distribution of strontium concentrations in the Edwards aquifer. Strontium concentrations controlled by 1. recharge water: ●, 2. "bad water": ○, and 3. leakage from Glen Rose Formation: +.

the Mt. Bonnell fault is relatively high in calcium and bicarbonate, suggesting saturation with respect to calcite and dolomite as computed by SOLMNEQ.

It is noteworthy that the concentration of dissolved carbon dioxide (pCO_2) calculated for water samples from surface streams (pCO_2 range from 0.001 to 0.004 atm) is higher and in most cases substantially higher than pCO_2 of water in equilibrium with normal atmosphere, which is about 0.0003 atm. This high content of dissolved carbon dioxide is probably due to the activity of organisms, oxidation of organic carbon, and interaction with soil water draining into the streams. Such soil water may

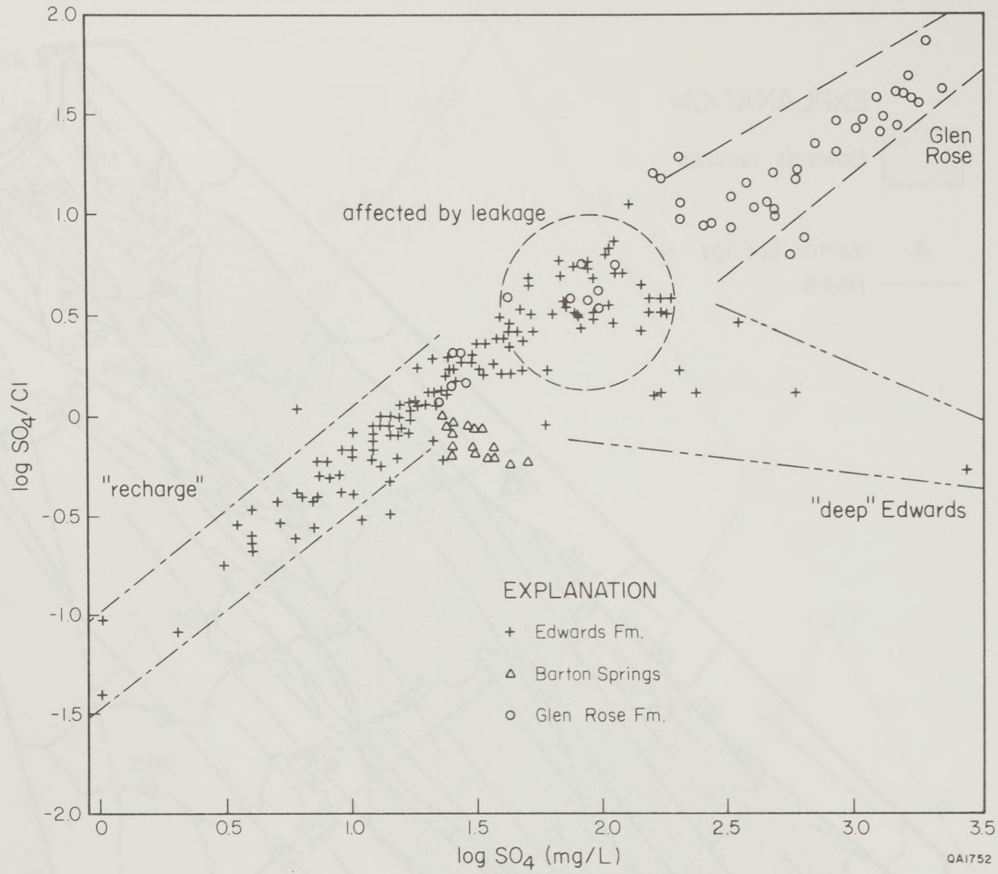

FIGURE 25. SO_4/Cl versus sulfate concentration for different types of water.

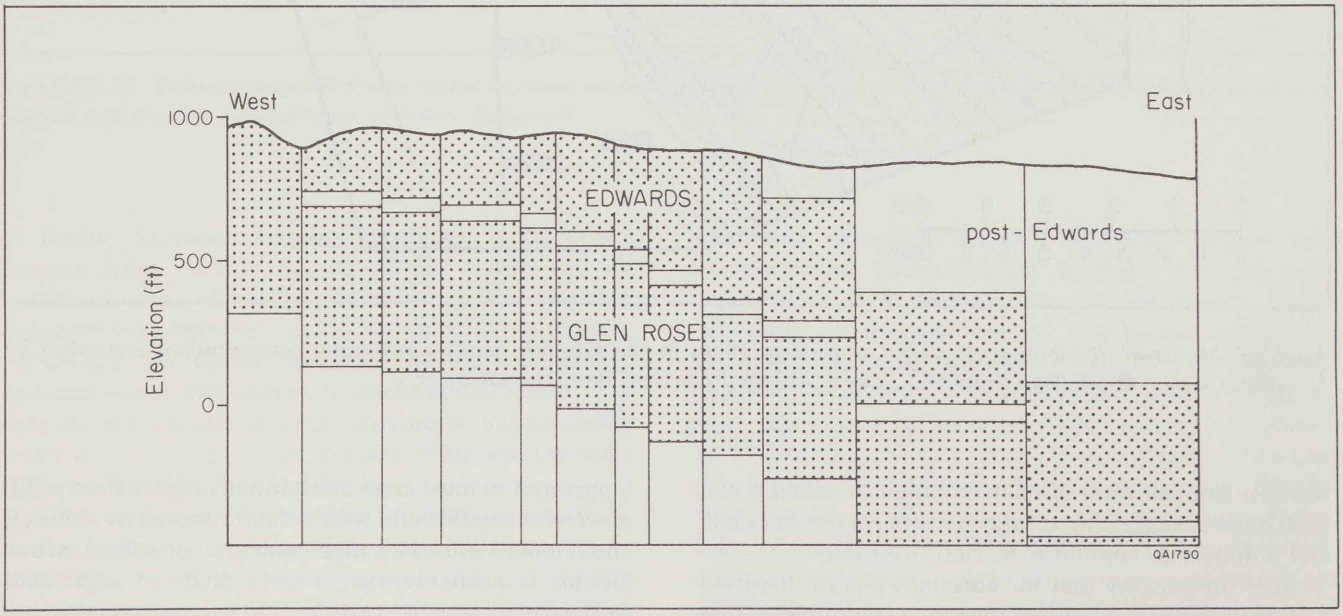

FIGURE 26. Schematic cross section across the Balcones Fault Zone, modified from Smith (1978). Vertical exaggeration = 7X.

have a pCO_2 of 0.1 atm or higher. In comparison, ground water and spring water have pCO_2 values of approximately 0.01.

On the basis of hydrological characteristics of Barton Springs, the Edwards Limestone was interpreted to be a diffuse-flow aquifer. Similarly, the Floridan Limestone was reported by Back and Hanshaw (1970) to be a diffuse-flow aquifer. Ground water in the Floridan aquifer is undersaturated with respect to calcite in the recharge area, but the water becomes supersaturated near the coast. This increase in saturation index in the direction of ground-water flow can be explained by calcite dissolution of the limestone.

Langmuir (1971) sampled well water from a diffuse-flow carbonate aquifer in central Pennsylvania. The ground water was undersaturated with respect to calcite; SI ranged from −0.38 to +0.04 with an arithmetic mean of −0.15. The Florida and Pennsylvania examples show that the carbonate saturation state of ground water is not necessarily a criterion for distinguishing between diffuse-flow and concentrated-flow conditions.

In the Balcones Fault Zone in the study area, recharge water enters the aquifer along the major creeks and flows eastward toward the confined part of the aquifer where new recharge water is mixed with "older" ground water. Thrailkill (1968) described various mechanisms, such as temperature change, mixing of dissimilar waters, and floods in surface streams, that could cause ground water to become undersaturated. A combination of all these mechanisms and oxidation of abundant organic matter in the creek water probably accounts for predominantly undersaturated ground water in the semiconfined Edwards aquifer and at the outflow at Barton Springs.

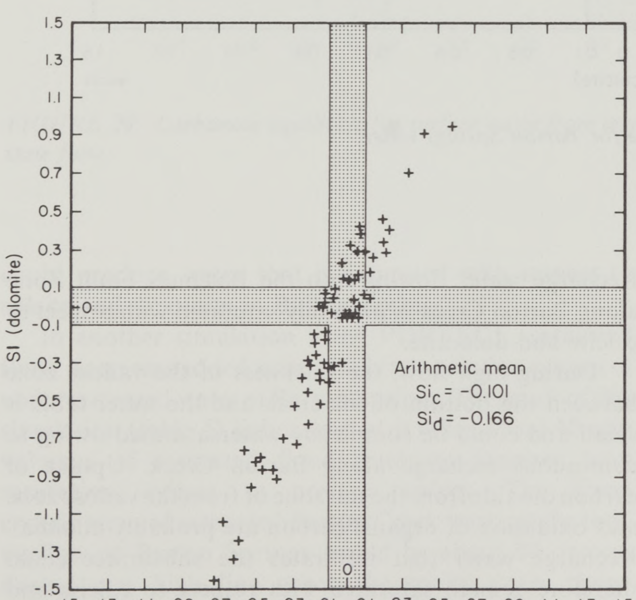

FIGURE 27. Carbonate equilibria for Edwards aquifer water.

TABLE 4. Results of mixing water from Barton Springs with "bad water." Chemistry of Barton Springs water is represented by water sample collected on September 19, 1979; chemistry of "bad-water" is represented by water sample collected from well 58-50-301 in 1949.

	SATURATION INDICES		LOG pCO_2
	Calcite	Dolomite	
Barton Springs	−0.1124	−0.6006	−1.537
"bad water"	+0.7673	+1.7233	−4.029
Unit volumes of "bad water" added to Barton Springs water:			
0.050	−0.1274	−0.6155	−1.549
0.075	−0.1415	−0.6293	−1.560
0.100	−0.1547	−0.6421	−1.571
0.125	−0.1670	−0.6541	−1.583
0.150	−0.1895	−0.6755	−1.604

The chemical variation of Barton Springs water indicates an influx from the "bad-water" zone during low flow. Water chemistry in the "bad-water" zone is apparently supersaturated with respect to calcite and dolomite. Saturation state is probably controlled by water-rock interaction (Prezbindowski, 1981). The contribution of nonpotable ground water, as characterized by the water chemistry in well 58-50-301, to the discharge from Barton Springs during very low flow was estimated to be between 5 and 10 percent (Senger, 1983).

The effect of the influx of water from the "bad-water" zone on the saturation state of ground water at Barton Springs was simulated with the computer program PHREEQE (Parkhurst and others, 1981). This program computes speciation of dissolved ions and saturation states of an aqueous solution. In comparison to SOLMNEQ, PHREEQE can also simulate several types of reactions including (1) adding reactants to a solution, (2) mixing two waters, and (3) titrating one solution with another.

The influx of water from the "bad-water" zone was simulated like a titration. Typical Barton Springs water with relatively low concentrations of sodium and chloride was the initial aqueous solution to which specific amounts of a solution representing the "bad water" (well 58-50-301) were added. Saturation states of the resulting aqueous solutions are shown in table 4. Addition of water from the "bad-water" zone does not increase the saturation indices of the resulting aqueous solution. Calcite and dolomite saturation indices actually decrease slightly despite the influx of highly saturated water.

The decrease in saturation indices owing to influx of "bad water" could permit enhanced carbonate dissolution at the interface between fresh water and "bad water." Back and others (1979) investigated the effect of mixing fresh ground water discharging into a lagoon with saline ocean water, both saturated with respect to calcite. At

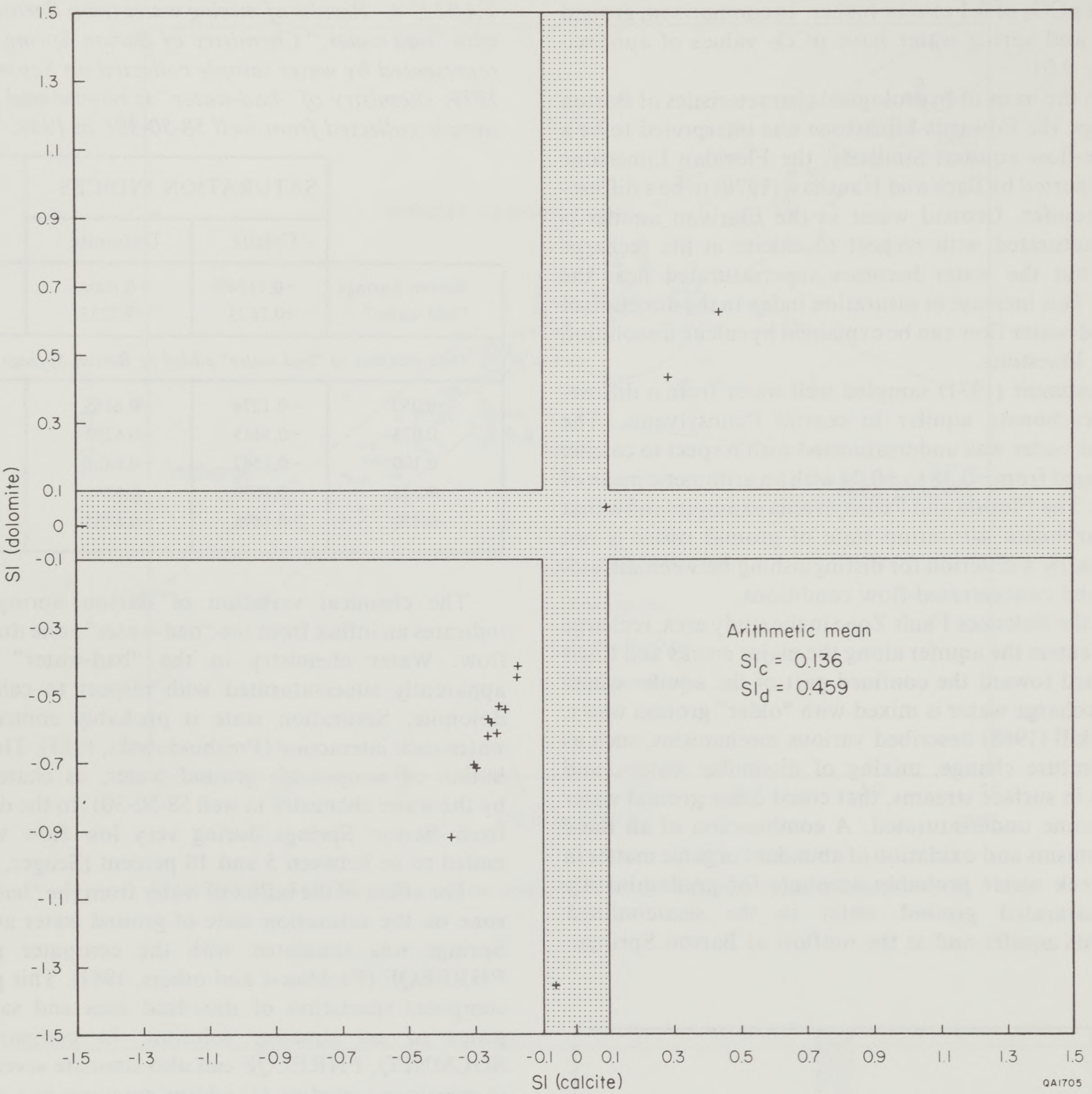

FIGURE 28. Carbonate equilibria for Barton Springs water.

the interface of the two solutions existed a brackish dispersion zone that was undersaturated with respect to calcite. Calcite dissolution in this zone was considered to be an important geomorphic process in forming the beaches along the east coast of the Yucatan Peninsula.

As mentioned earlier, the only water samples from Barton Springs that are saturated with respect to calcite and dolomite coincide with the highest discharge at Barton Springs in 1979 and 1981 (fig. 18). Saturation of Barton Springs water with respect to calcite and dolomite could occur when spring discharge is sustained primarily by recharge to the aquifer from Barton Creek. During floods, Barton Creek exhibits by far the highest flow rates compared with all the other creeks in the recharge area. Recharge water flowing into the Balcones Fault Zone along Barton Creek is probably saturated with respect to calcite and dolomite.

During high flow, the thickness of the vadose zone between the bottom of the creek and the water table is small and could be completely watersaturated owing to continuous recharge along Barton Creek. Uptake of carbon dioxide from the soil zone or from the vadose zone and oxidation of organic carbon are probably minimal. Recharge water that infiltrates the subsurface could therefore remain saturated with respect to calcite and dolomite. The contribution of essentially saturated water from Barton Creek may be large enough to overcome the mixing effect with undersaturated ground water. This

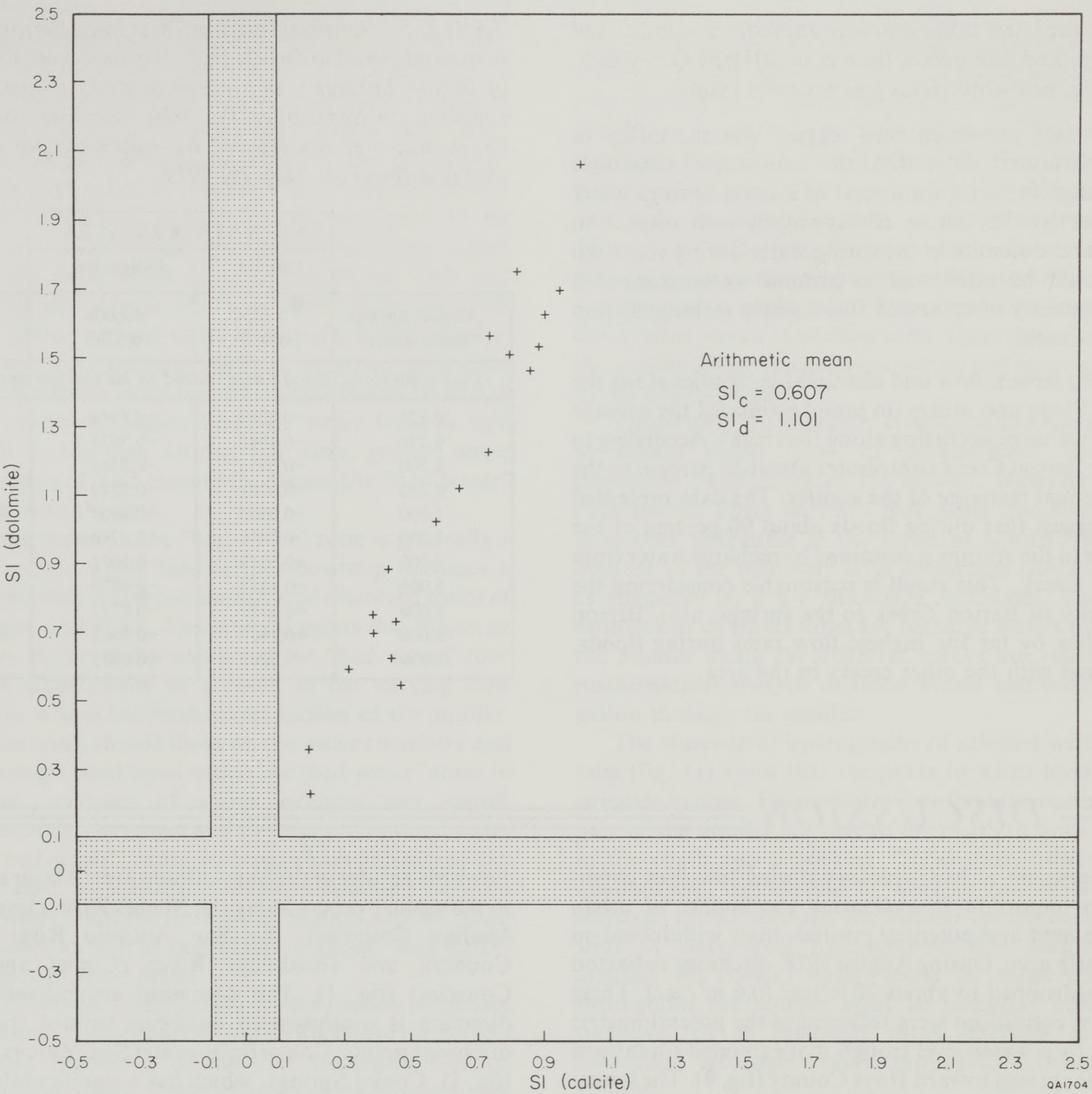

FIGURE 29. Carbonate equilibria for surface water from creeks; water samples collected during conditions of approximate steady-state flow.

could produce water that is saturated with respect to calcite and dolomite in Barton Springs.

In another simulation using PHREEQE, saturated flood water was added successively to a solution typical of undersaturated water at Barton Springs. The result of this simulation (table 5) indicates that it takes about 10 unit volumes of saturated flood water to increase both saturation indices to positive values. The direct contribution of saturated Barton Creek water to the total outflow at Barton Springs would be about 90 percent during very high flow when the spring water shows saturation with respect to calcite and dolomite.

There are limitations in interpreting these results. The simulation of the effect of mixing different solutions on the saturation state of the resulting solution is probably oversimplified. Chemical reactions that probably occur during the mixing of the solutions were not taken into account. Instead, the water chemistry of Barton Springs was assumed to represent the product of chemical processes occurring during the flow of ground water. Carbonate equilibrium values from flood-water samples are probably inaccurate because pH and temperature were measured in the laboratory and not in the field. This is also true for carbonate equilibria determination for water from well 58-50-301, which was assumed to represent the water chemistry of the "bad-water" zone. Values of saturation indices computed by the program PHREEQE differ from those obtained by SOLMNEQ because the latter takes into account more possibilities of ion pairs and complexes than does PHREEQE, which, therefore, probably yields less accurate results.

The data presented here suggest that the influx of highly saturated water from the "bad-water" zone does not change the saturation state of Barton Springs water significantly. The cause of saturation with respect to calcite and dolomite in the spring water during very high flow could be attributed to ground water composed predominantly of saturated flood water recharged from Barton Creek.

Using stream flow and channel-loss studies along the creeks, Slade and others (in press) estimated the average annual recharge occurring along the creeks. According to table 2, Barton Creek contributes about 28 percent to the total annual recharge of the aquifer. The data presented here suggest that during floods about 90 percent of the outflow in the springs is sustained by recharge water from Barton Creek. This result is reasonable considering the proximity of Barton Creek to the springs; also, Barton Creek has by far the highest flow rates during floods, compared with the other creeks in the area.

TABLE 5. Results of mixing water from Barton Springs with saturated flood water from Barton Creek. Chemistry of Barton Springs water is represented by water sample collected on September 19, 1979; chemistry of Barton Creek water is represented by water sample collected during a flood on May 29, 1979.

	SATURATION INDICES		LOG
	Calcite	Dolomite	pCO_2
Barton Springs	−0.1224	−0.6006	−1.537
"flood water"	+0.3679	+0.3707	−2.495
Unit volumes of "flood water" added to Barton Springs water:			
0.125	−0.1094	−0.5940	−1.583
0.250	−0.1045	−0.5835	−1.624
0.500	−0.0913	−0.5561	−1.693
0.750	−0.0761	−0.5249	−1.750
1.000	−0.0603	−0.4927	−1.799
2.000	−0.0018	−0.3740	−1.937
4.000	+0.0813	−0.2061	−2.091
6.000	+0.1355	−0.0970	−2.177
8.000	+0.1727	−0.0221	−2.232
10.000	+0.1997	+0.0323	−2.271
12.000	+0.2202	+0.0735	−2.300

DISCUSSION

Assessment of hydrogeology during low-flow conditions is important in evaluating the impact of urban development and potential ground-water withdrawal on the Austin area. During August 1978, discharge in Barton Springs dropped to about 20 ft^3/sec (0.6 m^3/sec). These low-flow conditions were reflected in the potentiometric surface by a water-level trough that extended southward from the springs toward Hays County (fig. 9). The lowest discharge ever recorded at Barton Springs (since 1894) was about 10 ft^3/sec (0.3 m^3/sec) at the end of a prolonged drought in 1956.

The ground-water flow divide between the Edwards aquifer, Austin region, and Edwards aquifer, San Antonio region, is defined by a potentiometric high in northern Hays County. The altitude at Barton Springs of 440 ft (134 m) and the altitude at San Marcos Springs of 670 ft (204 m) represent the lowest water level in each section of the aquifer. A significant drop in water level (below 670 ft) caused by a prolonged drought or by excessive ground-water withdrawal in the area of the ground-water flow divide could create a hydraulic gradient between San Marcos Springs and Barton Springs. Thus, ground water from relatively higher potentials in the Edwards aquifer, San Antonio region, would flow toward Barton Springs (Guyton and Associates, 1958).

The possibility of inflow of ground water from the south is supported by the hydrologic setting of the Edwards aquifer in the San Antonio area. Major recharge to the aquifer occurs along the Nueces River (Uvalde and Medina Counties), the San Antonio River (Bexar County), and Guadalupe River (Comal and Hays Counties) (fig. 1). The dominant ground-water flow direction is southwest to northeast toward the major discharge points: Comal Springs and San Marcos Springs (fig. 1). Comal Springs, which has a significantly higher mean annual discharge rate of 254 ft^3/sec (7.2 m^3/sec) compared to San Marcos Springs with 144.4 ft^3/sec (4.09 m^3/sec) and Barton Springs with 50 ft^3/sec (1.42 m^3/sec), went dry during the drought in 1956, while San Marcos Springs and especially Barton Springs yielded significant discharge. Therefore, during extremely low flow, ground water from the southern part of the Edwards Underground Reservoir apparently could move into the Austin area.

Water chemistry in most of the wells in the aquifer does not change significantly during variations in flow. Barton Springs, however, shows significant increases in chloride and sodium with decreasing discharge. This increase is related to the influx of "bad water." Possible encroachment of "bad water" into the major flow path of the aquifer during low flow indicates that the "bad-water" line is not a stationary boundary. Excessive ground-water withdrawal from the confined section of the aquifer could eventually cause inflow of nonpotable water into the fresh-water zone.

The "bad-water" zone is lithologically characterized by low permeability with intergranular porosity (Abbott, 1975). Fluid movement in this zone can be expected to be relatively slow. Comparing the water-level variations in well 58-50-301 with the water level of the other wells (fig. 11) suggests that low permeability in the "bad-water" zone causes the delay of the water-level response of well 58-50-301. Therefore, during the recession periods the water level in well 58-50-301 remains relatively higher than do the water levels in wells west of this well, and nonpotable water from the "bad-water" zone can move into the major ground-water flow path of the aquifer. During the high recharge periods in spring and early summer, the water levels in wells 58-50-216 and 58-50-518 become relatively higher than the water level in well 58-50-301 to the east. During that time, ground water moves eastward and essentially causes the "bad-water" line to shift to the east.

Ground water in the "bad-water" zone is generally a sodium sulfate water that farther downdip becomes a sodium chloride water containing total dissolved solids of 1,000 mg/L and more. This study suggests that the water chemistry in the updip section of the "bad-water" zone can vary significantly as a result of the varying flow conditions within the fresh-water section of the aquifer.

Future work should focus on the water chemistry and hydrogeologic conditions within the "bad-water" zone. In particular, analyses of stable isotopes can supply information on the origin and evolution of ground water in the "bad-water" zone. Isotopic characterization can also be used to identify possible interaction between ground water from the fresh-water section and the "bad-water" zone, as well as interaction between "bad water" and deep formation brines.

The effects of urbanization on the quality of surface and subsurface waters in the Austin area were the subject of a study conducted by the U.S. Geological Survey, Austin district office. Channel losses along Barton Creek vary along its course downstream from Loop 360 (fig. 2). During conditions of relatively high ground-water flow, Barton Creek gains water, whereas during conditions of low ground-water flow, the lower reaches of Barton Creek lose surface water to the aquifer. This potential recharge area near dense urban development makes Barton Creek very sensitive to pollution, which in turn would rapidly affect the water quality in nearby Barton Springs. In fact, high bacteria counts in recent years indicate human and animal sources of pollution in Barton Springs water, mostly after heavy rainfall. Determining the hydrodynamic and dispersive characteristics within the aquifer is important in evaluating pollutant transport in the ground water.

Tracer tests are generally conducted in karst aquifers to obtain travel times of ground-water flow and dispersion coefficients of the aquifer. However, tracer experiments along Barton Creek using fluorescent dyes have been unreliable in the Edwards aquifer, probably because of the relatively large dispersion characteristics as suggested by a low discharge coefficient. Therefore, estimations of travel times for recharge waters that enter the aquifer along the different creeks are restricted to mathematical analysis of flood events and their propagation through the aquifer.

The water-level hydrographs of selected wells in the area (fig. 11) show that the peaks in water level do not coincide in time. The peaks from wells in the northeastern part of the aquifer lag behind the peaks in wells located in the south and southwest. These data, however, are qualitatively insufficient to analyze time-of-travel of recharge events through the aquifer on the basis of the shift in water-level hydrographs. For mathematical analysis of floods, water levels in numerous wells throughout the aquifer must be measured either continuously or in relatively short intervals to detect individual recharge events propagating through the aquifer.

SUMMARY

The Edwards aquifer in the Austin region is characterized by large water-level fluctuations and good hydrologic interconnection with its major discharge site, Barton Springs. Recharge to the aquifer occurs predominantly along the five major creeks within the Balcones Fault Zone. Creek water flows into the Balcones Fault Zone from the west and infiltrates through faults and fractures into the aquifer along the creek beds, losing up to 100 percent of the stream flow to the aquifer.

The potentiometric surface of ground water in the aquifer changes significantly between periods of high and low flow. During high flow, the main ground-water flow component is from the southwest toward Barton Springs. Ground-water flow lines during low flow are concentrated in the eastern part of the Balcones Fault Zone. The largest water-level fluctuations occur in the northeastern part of the aquifer, and changes in water levels of wells correlate well with changes in discharge. Water levels of wells in the city of Rollingwood, however, show no correlation with flow in Barton Springs.

Aquifer parameters were evaluated quantitatively using the recession curves of the outflow at Barton Springs and water-level decline data from observation wells in the area of investigation. In addition, the water-level response in well 58-42-915 was simulated using a transient ground-water flow model. Calculated values of transmissivities and storativities based on the recession-curve analysis range from 0.1 m^2/sec to 0.4 m^2/sec and from 0.001 to 0.023, respectively. Using a transmissivity of 0.2 m^2/sec, the ground-water flow model reproduced with acceptable accuracy the observed discharge at Barton Springs and the transient response of water level in well 58-42-915. Storativity in the model was 0.00075, which is one order of magnitude lower than the average storativity of 0.0075 estimated for the whole aquifer.

The chemistry of water at Barton Springs varies with varying flow. The increase of sodium, chlorine, sulfate, and especially strontium with decreasing discharge indicates influx from the "bad-water" zone. The interaction between the "bad-water" zone and the fresh-water aquifer is indicated in the water-level fluctuations of well 58-50-301, located in the "bad-water" zone, and the good correlation of changes in water level with changes in spring flow.

Water chemistry in the Edwards aquifer generally remains constant. The aquifer contains calcium bicarbonate water that becomes a sodium sulfate water and then farther downdip, a sodium chloride water. In some locations, however, leakage from the Glen Rose Formation increases the sulfate and strontium concentrations. Leakage is associated with large displacements of faults, which bring the Edwards Formation into contact with the Glen Rose Formation updip.

Carbonate equilibrium values of water samples from the area exhibit a wide range of saturation indices and indicate undersaturation with respect to calcite and dolomite for the ground water and Barton Springs water. Various mechanisms such as temperature change, mixing of dissimilar ground waters, and floods in surface streams are likely to significantly affect the saturation state of the ground water. In particular, the influx of highly saturated "bad water" into the fresh-water aquifer theoretically results in a decrease in saturation indices. This mechanism would enhance carbonate dissolution at the interface between fresh water and "bad water," thereby creating increased permeabilities in the interface section of the aquifer.

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