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PERMIAN EVAPORITE

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CAUTION

This report describes research carried out by staff members of the Bureau of Economic Geology that addresses the feasibility of the Permian Basin for isolation of high-level nuclear wastes. The report describes the progress and current status of research and tentative conclusions reached. Interpretations and conclusions are based on available data and state-of-the-art concepts, and hence, may be modified by more information and further application of the involved sciences.

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HYDROLOGY OF AN EVAPORITE AQUITARD: PERMIAN EVAPORITE
STRATA, PALO DURO BASIN, TEXAS

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ABSTRACT

Permian evaporite strata of the Palo Duro Basin, Texas Panhandle, are considered to have low permeabilities, but are capable of some leakage. Cross-sectional and areal modeling of regional ground-water flow in the Palo Duro Basin indicates leakage from the Upper aquifer through the Evaporite aquitard that contributes approximately 30 percent of the water in the underlying Deep-Basin Brine aquifer. The regional estimate of permeability of the aquitard ($\approx 10^{-4}$ md) is in general agreement with laboratory testing of salt permeability. Oxygen and hydrogen isotopic composition and the Cl/Br ratio of the deep-basin brines also suggest leakage through the aquitard. Isotopically nonequilibrated meteoric waters are found to depths of 8,000 ft.

Leakage through the aquitard may occur as either matrix or fracture flow. According to petrographic and geochemical studies of the evaporite strata, halite dissolution and minor recrystallization occurred predominantly during Permian time. Post-Permian waters may have altered the top of the Evaporite aquitard. Fracturing is prevalent in the rock strata in the basin; the greatest density of fracturing occurs in areas with tectonic deformation. The potential for leakage may be greatest in these areas.

INTRODUCTION

Many hydrogeologic concepts are based on the idea that low-permeability rocks function as aquitards rather than as aquicludes; that is, given enough time, rocks with low permeabilities have the capability of transmitting fluids. Recharge and discharge through low-permeability argillaceous aquitards are important parts of the hydrodynamics of many sedimentary basins as well as of individual aquifers (see Bredehoeft and others, 1982, for an excellent summary). Implicit in the phenomenon of cross-formational flow in the Red Earth Region (Toth, 1978) is the significance of leakage across low-permeability aquitards.

The concept of regional fluid flow through evaporites has received less attention. Hydrologic studies of evaporites have been site specific and generally have assumed rock salt to be impermeable. The early decision to consider salt as an appropriate medium for the disposal of high-level nuclear waste was based on the impermeable nature of rock salt. The Committee on Waste Disposal of the Division of Earth Sciences of the National Research Council (1957) concluded that "...no water can pass through the salt. Fractures are self healing" (p. 4) and "the salt is impervious to the passage of water because of its plasticity and crystalline structure, so that the mined-out space is very dry." (p. 135). Roedder (1984) commented on the 1955 National Research Council Meeting, "Several of the attendees indicated that salt mines were the driest of all mines that they knew and that obviously salt deposits must have been dry for a long time in the past; if not, they would have dissolved in the intervening hundreds of millions of years." Since then studies have concentrated on determining the permeability of salt rocks being considered for storage of natural and refined products or for the disposal of chemical and nuclear wastes. These studies have also examined migration of fluid inclusions (internal fluids) through salt under thermal gradients. A distinction is made between external and internal fluids; external fluids are those that originate outside the rock unit and move into and through the rock by some external force; internal fluids are those waters that have been trapped as inclusions during deposition or early diagenesis of minerals such as halite. These internal fluids

migrate under thermal gradients and provide a possible source of water for corrosion of cannisters being used for waste storage (see Roedder, 1984, for a detailed discussion of internal fluids).

This paper deals with the potential for regional migration of external waters through evaporite strata and the effects of leakage on the hydrodynamics and hydrochemistry of the waters in a sedimentary basin. The Permian evaporite section of the Palo Duro Basin, Texas Panhandle, provides a case study of the possible mechanisms and effects of fluid movement through a thick evaporite section. The paper presents 1) a review of permeability studies of salt, 2) a summary of the regional hydrogeologic setting of the Palo Duro Basin, 3) the hydrologic data that indicate that the evaporites function as an aquitard, 4) the hydrogeochemical data that support the concept of leakage through the salt section, 5) the petrographic and geochemical data that suggest how leakage occurs, and 6) the integration of the hydrologic, geochemical, and petrographic evidence regarding flow through evaporite strata. The research is part of an ongoing characterization of the hydrogeology of the Palo Duro Basin, which is being considered by the U.S. Department of Energy as a potential site for a high-level nuclear waste repository.

PERMEABILITY OF ROCK SALT

Previous investigations of the permeability of salt have been laboratory studies and in situ tests of domal or bedded salt deposits. Gloyne and Reynolds (1961) found that permeabilities for dome salt ranged widely from 0 to 10^2 millidarcies (md). Gilpatrick and others (1982) interpreted these "0" values as less than 0.5×10^{-3} md; permeability of bedded salt was lower and ranged from 0 to 10^{-2} md. Aufricht and Howard (1961) observed that average gas permeabilities for salt from four mines in Gulf Coast salt domes ranged from 2×10^{-3} to 631 md with no overburden pressures. Permeability with respect to a flowing brine was reduced to 0 when a confining pressure of 800 psi was applied. In a more recent study of Avery Island

salt dome, Blankenship and Stickney (1983) measured in situ gas permeabilities that ranged from 0.6×10^{-6} to 5.6×10^{-3} md with an average value of $\approx 10^{-4}$ md.

Permeability studies of bedded salts have concentrated primarily on the Permian Salado Formation at the Waste Isolation Pilot Project (WIPP), Carlsbad, New Mexico. In situ measurements conducted by Peterson and others (1981) indicated gas permeabilities ranging from 3×10^{-3} md to 2×10^{-2} md. Liquid permeability tests indicated an approximate value of 10^{-3} md. Laboratory tests of salt samples indicated lower k values for similar types of salt with a range of 10^{-3} to 10^{-4} md (Sutherland and Cave, 1980).

Fluid flow is along crystal boundaries rather than through the crystals. Gilpatrick and others (1982) estimated the permeability along halite grain boundaries by measuring fluid flow between stressed halite crystals. Permeabilities range from 1.4×10^{-4} to 1.4×10^{-2} md for crystal sizes that varied from .01 to 1 cm and a thickness of 0.2 mm for the opening between crystal boundaries. Gas permeability through a single salt crystal was estimated to be approximately 10^{-9} md (Sutherland and Cave, 1980), or 10^5 times lower than permeability along grain boundaries, and therefore is not an important pathway for fluid migration. Recrystallized textures are evidence that fluids saturated with NaCl move through halite. In a study of the textures and fabrics of halite in Oakwood salt dome, East Texas, Dix and Jackson (1982) observed recrystallization of the halite and an increased density of fluid inclusions down to 2 m below the caprock/salt contact. The original anhydrite banding was still present in the altered halite, indicating that solid-state recrystallization (micro solution and reprecipitation) and not large-scale solution and reprecipitation had occurred. Kreidler and Dutton (1983) suggested that these inclusions represent deep basinal waters that caused the last dissolution event of the salt dome. Recrystallization of halite and the associated fluid inclusions must have resulted from the migration of external fluids into the salt.

Fracture permeability may be an important aspect of fluid flow in evaporite aquitards. Based on differences between laboratory and in situ permeabilities, Peterson and others (1981)

concluded that the higher permeabilities for the in situ test resulted from flow through fractures. Measured pressures matched predicted build-up curves for fracture flow. Gilpatrick and others (1982) showed that permeability at the salt crystal boundary increases by several orders of magnitude as the spacing between crystals increases by an order of magnitude. In a related study, Chaturvedi (1984) observed open fractures in the Salado Formation that were associated with gas blowouts in potash mines. The fractures were considered to be the reservoir containing the high-pressured gas; the fractures resulted from geologic processes and not from the blowout. The presence of these fractures argues against the generally accepted concept that fractures in salt heal themselves.

The reported range of salt permeability of 10^{-3} to 10^{-5} md is low but within the range of permeabilities determined for other aquitards. On the basis of laboratory tests by Peterson (1954), Young and others (1964), and Davis (1969), Freeze and Cherry (1979) concluded the average hydraulic conductivity of shale ranged from 10^{-10} to 10^{-12} m/s ($\approx 10^{-2}$ to 10^{-4} md). The potential for evaporites to transmit fluids appears similar to that of shales; therefore, evaporites as aquitards are important elements in a regional hydrologic system.

THE EVAPORITE AQUITARD, PALO DURO BASIN, TEXAS PANHANDLE

Regional Hydrogeologic Setting

The hydrogeologic setting of the Palo Duro Basin is made up of three heterogeneous units, which are informally referred to in this paper as the Upper aquifer, the Evaporite aquitard, and the Deep-Basin Brine aquifer (table 1) (Bassett and Bentley, 1983). The Upper aquifer includes fluvial deposits of the Tertiary Ogallala Formation as well as the fluvial, lacustrine, and deltaic deposits of the Triassic Dockum Group and Cretaceous formations. The Evaporite aquitard consists of Permian (Leonardian, Guadalupian, and Ochoan) deposits of red mudstone and siltstone, limestone, dolostone, anhydrite, and halite (table 1). The Deep-Basin Brine aquifer is

composed of Wolfcampian and upper Pennsylvanian shelf-margin carbonates, Paleozoic arkosic fan-delta deposits (granite wash), lower Paleozoic carbonate rocks, and Cambrian sandstones.

Geohydrologic units of the Palo Duro Basin are bounded to the west by the Sacramento and Sierra Grande Uplifts, the eastern limit of Laramide deformation, and to the east by the Hardeman Basin (figs. 1 and 2). The Amarillo Uplift and the Oldham Nose are considered as a generally impermeable boundary on the north (fig. 1). Some flow may cross the western part of the uplift where the Wolfcamp is continuous over the uplift (Wirojanagud and others, 1984). The Matador Arch to the south is considered to be a no-flow boundary of the Deep-Basin Brine aquifer. During the Laramide Orogeny, the basin was uplifted and tilted eastward. Because of this structural relief and the distribution of outcrops, recharge areas are to the west of the basin and discharge areas are to the east. Ground-water flow is generally eastward across the Palo Duro Basin.

The main sources of water in the Deep-Basin Brine aquifer are thought to be recharge in New Mexico and downward leakage through the Evaporite aquitard (figs. 2 and 3). Whereas most ground water recharging in the Pecos River Basin is discharged to the Pecos River, some regional underflow recharges the deep saline aquifers (Summers, 1981; Senger and Fogg, 1984). Ground-water flow in the Deep-Basin Brine aquifer is generally west-to-east and northeast (Wirojanagud and others, 1984). Ground-water flow in the Deep-Basin Brine aquifer is controlled by permeability distribution within the deep section. Ground water flows from the basin center northeast toward the thick, permeable granite wash deposits that flank the Amarillo Uplift. The uplift may be a barrier to flow between the Palo Duro and Anadarko Basins; at the uplift, flow is diverted to the southeast along the strike of the granite-wash (arkosic fan) facies. Potentiometric levels in the Deep-Basin Brine aquifer are significantly below those in the High Plains aquifer because 1) fluid pressure is controlled by land-surface elevation in the recharge and discharge zones and not by depth below the Upper aquifer and 2) the aquifer in the granite wash may act as a hydrologic sink lowering pressures throughout

the rest of the basin (Senger and Fogg, 1984). Because of these lower pressures, there is a potential for ground water to flow from the Upper aquifer to the Deep-Basin Brine aquifer (Wirojanagud and others, 1984; Senger and Fogg, 1984). Transit time across the basin within the deep aquifers is expected to be 2-4 million years (Wirojanagud and others, 1984). Discharge points appear to be in a dispersed zone across North-Central Texas and southern Oklahoma and not at the discrete saline springs just to the east of the Caprock Escarpment.

Geologic Characterization of the Evaporite Aquitard

The Evaporite aquitard is composed of Middle and Upper Permian evaporite-bearing strata extending from the top of the Wolfcamp strata to the dissolution zone at the top of the Permian evaporite strata. Strata consist of four major lithofacies, halite, anhydrite, dolomite, and red argillite (Handford, 1980). A core study of the evaporite section from the DOE Gruy Federal No. 1 test well in the central part of the Palo Duro Basin indicates 58 percent halite and anhydrite, 32 percent red beds, and 10 percent dolomite. The total thickness of the section varies from 650 to 1550 m, thickening toward the Midland Basin. The evaporite section is composed of five major genetic units: 1) the Wichita-Red Cave Formations, 2) the lower Clear Fork and Tubb Formations, 3) the Upper Clear Fork and Glorieta Formations, 4) the San Andres Formation, and 5) the post-San Andres Guadalupian and Ochoan strata. Each genetic unit represents a regressive-transgressive sequence (Dutton and others, 1982). A typical cycle within a unit such as the San Andres Formation is dark anhydritic mudstone, limestone and dolomite, anhydrite, halite, and terrigenous red beds. Each rock type represents a different depositional facies associated with different salinities. The cycle begins with an influx of marine waters that dissolve some of the underlying halite from the previous cycle, leaving a residuum of mud and anhydrite. Normal marine conditions return with the deposition of limestone. With intense evaporation and increasing salinities, dolomitization of limestone, gypsum/anhydrite precipitation, and finally halite precipitation occur. Terrigenous red-bed

clastics may then be deposited over the salt (Fracasso and Hovorka, 1984). Halite beds rarely are more than 1-3 m thick.

Hydrologic Evidence of Leakage through the Evaporite Aquitard

The Evaporite aquitard is an important element in the hydrology of the Palo Duro Basin. Without the aquitard, fluid pressures in the Deep-Basin Brine aquifer would equilibrate with the Upper aquifer and not be subhydrostatic as observed. Hydraulic heads in the Upper aquifer are 250 to 650 m higher than fresh-water equivalent heads in the saline Wolfcamp aquifer (fig. 3), indicating a potential for downward ground-water flow. If the evaporite strata have some finite permeability, there should be leakage through the aquitard to the deep saline aquifers. Two numerical modeling studies of the Palo Duro Basin - one an areal model and the second a cross-sectional model (Wirojanagud and others, 1984, Senger and Fogg, 1984, respectively) - indicate leakage is downward through the Evaporite aquitard. The areal model simulates the observed potentiometric surface of the Wolfcamp aquifer. The model incorporates the permeability of the different lithologies in the aquifer as well as an average permeability for the overlying aquitard that allows leakage from the Upper aquifer to the deeper section. Initial estimate for vertical permeability of the aquitard was 2.0×10^{-4} md, which was derived from the harmonic means of permeabilities for two cross sections through the evaporites, using typical or measured values for each substratum. Numerical simulations using this vertical permeability resulted in a computed head distribution that differs significantly from the Wolfcamp potentiometric surface. When permeability was lowered to 8×10^{-5} md, a reasonable agreement was found. Computed fluxes through the aquitard per unit area are very small; however, because of the large surface area, leakage through the Evaporite aquitard accounts for 50 percent of the ground water flowing through the Wolfcamp aquifer and 33 percent of the flow through the whole Deep-Basin Brine aquifer.

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Cross sectional modeling by Senger and Fogg (1984) (fig. 3) yielded similar results using the average permeability values for the different hydrologic units, according to Wirojanagud and others (1984). The initial estimate of 2.0×10^{-4} md for the permeability of the Evaporite aquitard resulted in a reasonable simulation of the underpressured conditions of the Deep-Basin Brine aquifer. Increasing permeability to 2.0×10^{-3} md caused too much leakage and computed heads in the deep basin aquifers approached hydrostatic conditions. Senger and Fogg (1984) concluded that a value of 2.0×10^{-4} md represented an upper limit of aquitard permeability. According to the cross-sectional model, 20 percent of the ground water in the Deep-Basin Brine aquifer (Wolfcamp plus deeper saline aquifers) could result from leakage through the Evaporite aquitard. Leakage rate (Darcian flow) through the aquitard is 6×10^{-8} m/day and average linear velocity is 6×10^{-5} m/day, assuming a porosity of 0.1 percent.

The modeling studies suggest that an average permeability value for the evaporite section is approximately 10^{-4} md, a value which agrees with the laboratory matrix permeabilities from the Salado Formation (Sutherland and Cave, 1980) but is less than the in situ values obtained by Peterson and others (1981), which may have measured fracture permeability. These permeability values for the Palo Duro represent regional and not site specific estimates and, therefore, incorporate fracture permeability as well as grain boundary (matrix) permeability. These regional values, however, correlate better with the range for Sutherland and Cave's (1980) matrix permeability (10^{-3} to 10^{-4} md) than with the range for Peterson and others' (1981) fracture permeability, suggesting that fracture permeability in the Evaporite aquitard may not be significant or that the matrix permeability for the aquitard may be lower than is indicated by laboratory measurements of salt. Fracture permeability in the Evaporite aquitard of the Palo Duro Basin, however, is considered important (and will be discussed in a later section). The permeability of the halite in the section may not be the limiting factor; the permeabilities of shales and red beds may be lower than the permeability of salt and may lower the estimated overall permeability of the aquitard. Bredehoeft and others (1982) report in-situ shale hydraulic

conductivities of 5×10^{-11} cm/s ($\approx 10^{-5}$ md) for the Pierre Shale, which is one order of magnitude lower than the estimated permeability of the Evaporite aquitard.

Vertical leakage through the Evaporite aquitard is complicated by the presence of interbedded carbonates and sandstones that have permeabilities higher than those of the evaporites. For example, permeability in the 15 to 35 m thick carbonate strata in the lower part of the San Andres Formation ranges from 6×10^{-3} to 2×10^{-1} md. Potentiometric surface mapping of the San Andres carbonates in the Evaporite aquitard of the Palo Duro Basin and the regions surrounding the basin suggests regional, horizontal flow southeastward through the San Andres carbonates. Flow lines from vertical leakage would be deflected 2 to 50 km by this horizontal flow (Dutton and Orr, in preparation).

Using this average value of 10^{-4} md for the permeability of the Evaporite aquitard, recharge to the Deep-Basin Brine aquifer via leakage appears to be an important hydrogeologic process. These estimates of amount of leakage, however, are based on a limited amount of data for a large geographic area. Additional methods, such as identification of the leakage waters by geochemical analyses and identification of pathways of leakage through the aquitard, are needed to complement the modeling results.

Isotopic and Chemical Evidence of Leakage through the Evaporite Aquitard

If fluids are leaking through the Evaporite aquitard, the chemical and isotopic composition of brines in the Wolfcamp carbonates should verify this leakage process. Representative samples of formation waters were collected in the most porous zones in four deep hydrologic test wells (Stone and Webster Engineering Corporation No. 1-Mansfield, Stone and Webster Engineering Corporation No. 1-J. Friemel, Stone and Webster Engineering Corporation No. 1-Zeeck, and Stone and Webster Engineering Corporation No. 1-Sawyer, fig. 1). The deepest zone penetrated the Ordovician Ellenburger Formation. Most of the sampling, however, was within Pennsylvanian carbonates, Pennsylvanian granite wash (arkosic alluvial fan sand-

stones), and Permian Wolfcamp carbonates. Samples were analyzed for major, minor, and trace elements, isotopic composition of carbon, hydrogen, oxygen, and sulfur, and gas content (table 2) (Fisher, 1984). Interpretation of the hydrochemistry of the deep saline aquifers is based on these analyses. Other investigators have measured additional chemical and isotopic species in the water (e.g., Hubbard and others, 1984).

The isotopic compositions of the brines from the subsalt aquifers vary across the basin. Brines from two wells (No. 1-Zeeck and No. 1-Sawyer) in the interior and eastern part of the basin have $\delta^{18}\text{O}$ values that range from -1.7 to +2.2 ‰ and approach isotopic equilibrium with dolomite for the sampled depths and temperatures (fig. 4). In contrast, the brines in the northern part of the basin (No. 1-Mansfield) are not isotopically equilibrated with dolomite and have oxygen isotope compositions (-5.1 to 5.8 ‰) similar to the isotopic composition of waters in the Upper aquifer. In the north-central part of the basin (No. 1-J. Friemel), brines are not equilibrated with respect to dolomite and have oxygen isotope compositions (-2.0 to -3.7 ‰) that appear to indicate a mixture of deep basinal and shallow meteoric waters. This lack of isotopic equilibration is unexpected considering the suggested old ages (60 - 260 million years) of the brines (Ziakowski and others, 1984) and the long residence times (2-5 million years) that have been calculated from numerical modeling (Wirojanagud and others, 1984; Senger and Fogg, 1984).

The kinetics of the isotopic exchange do not appear to be a factor controlling the isotopic disequilibrium. Several of the brines at lower formation temperatures approach or are in isotopic equilibria with dolomite whereas the deeper and hotter brines show the greater disequilibrium (fig. 4). In addition, the range of deep-basin brine temperatures (35° - 56°C) would not result in significantly different water-rock exchange rates between samples.

The lack of equilibration appears to result from mixing two different waters, 1) a deep-basin brine that has equilibrated with the carbonate rocks, and 2) a nonequilibrated brine. A plot of δD vs $\delta^{18}\text{O}$ (fig. 5) can be interpreted as a mixing curve that extends from the isotopic

composition of meteoric water in the region to the isotopic composition of equilibrated deep-basin brines. Brines from the northern part of the basin (No. 1-Mansfield well) have isotopic compositions similar to the isotopic values of meteoric water from the Upper aquifer. Brines from the central and eastern parts of the basin (No. 1-Zeeck and No. 1-Sawyer wells) appear to be equilibrated with the aquifer matrix. Isotopic compositions of the waters from the north central part of the basin (No. 1-J. Friemel well) are intermediate between the other two water types and appear to be a mix between deep basinal and meteoric waters. The volume percent of isotopically nonequilibrated water in the brine can be estimated from simple mixing calculations. For the case of Wolfcamp brine from No. 1-J. Friemel (table 2), the oxygen isotopic composition is ≈ -3.7 ‰. The predicted isotopic composition for a brine in equilibrium with dolomite at formation temperature is ≈ 0 ‰. Assuming that the isotopically depleted water had an original isotopic composition similar to modern meteoric water for the region (≈ -5 ‰), then the Wolfcamp brine from No. 1-J. Friemel contains approximately 75 percent meteoric water. A mixing of meteoric water with the deep basinal brines explains the lack of isotopic equilibration for some of the samples.

Two sources for the meteoric waters should be considered. First, water from the Upper aquifer could leak downward through the matrix or along fractures in the Evaporite aquitard. As described previously, fluid flow through the salt section appears possible; there is finite permeability and a hydraulic gradient that would permit downward flow. Conversely, a second possible source, suggested by Knauth and Hubbard (1984), is that water could be recharged into permeable granite-wash deposits, which have been shed off granite uplifts, such as the Amarillo Uplift or the Oldham Nose. Water in the granite wash would travel down the flanks of the basin and then into the carbonates. The process is assumed to be so recent that isotopic equilibration has not occurred.

Chloride/bromide ratios of the deep-basin brines suggest that the waters in the deep-basin aquifers result partly from leakage through the evaporite section. Brines with high Cl/Br ratios

(2,000-3,000) typically are associated with dissolution of halite by meteoric ground water (Whittemore and Pollock, 1979, Kreitler and Bassett, 1983, Richter, 1984). Whittemore and Pollock (1979) reported Cl/Br ratios of 2,400 to 3,200 for commercial brines where fresh waters are injected into salt beds, halite is dissolved, and the brine is collected. Brines with lower Cl/Br ratios (160-250) are typical of deep-basin brines and may result from multiple cycles of solution and recrystallization of halite where the brine becomes enriched in Br and the recrystallized halite is depleted (Land and Prezbindowski, 1980). Experimental studies have documented this reaction as well as the effects of additional complexities, such as surface area, Mg content, and solid solution ratio (Wilson and Long, 1984). A plot of Cl/Br vs $\delta^{18}\text{O}$ (fig. 6) for the brines from the Deep-Basin Brine aquifer shows the Cl/Br ratio increases as the $\delta^{18}\text{O}$ values decrease. From this inverse relationship between $\delta^{18}\text{O}$ and Cl/Br, it appears that meteoric water from the Upper aquifer is dissolving halite without multiple cycles of solution and reprecipitation and then leaking into the Deep-Basin Brine aquifer.

The Na/Cl weight ratio for brines in the Wolfcamp carbonates suggests that the source of chloride initially is from salt dissolution and that the brines are not flowing from the granite wash to the Wolfcamp carbonates. The Na/Cl weight ratio for brine derived solely from salt dissolution is approximately 0.65; ratios less than 0.65 indicate loss of Na by geochemical reactions such as ion exchange or albitization. Halite dissolution is considered the source of high concentrations of chlorides in the Palo Duro brines (Bassett and Bentley, 1983). At each well, Wolfcamp samples have Na/Cl ratios of approximately 0.57 and are closer to this halite dissolution ratio than are any of the deeper, stratigraphically lower samples (granite wash, Mississippian and Pennsylvanian carbonates) (fig. 7), which suggests that the water chemistry of the Wolfcamp brines is more closely controlled by the process of salt dissolution than are the brines in the deeper units. The observed decrease in the Na/Cl ratio suggests some movement of brines from the Wolfcamp to the deeper units. Vertical hydraulic gradients in the Deep-Basin Brine aquifer also indicate a component of downward flow (Senger and Fogg, 1984; Orr

and others, in press). If water flows from the granite wash to Wolfcamp carbonates, a geochemical reaction is needed to explain an increase in Na/Cl ratios. We are not aware of a documented geochemical reaction other than salt dissolution that would result in increasing the Na/Cl ratio.

Isotopic and geochemical evidence in the Deep-Basin Brine aquifer suggests that there is leakage from the Upper aquifer through some parts of the Evaporite aquitard into the Deep-Basin Brine aquifer. Leakage does not appear uniform across the basin. Deep-Basin brines from No. 1-Sawyer and No. 1-Zeeck wells are isotopically equilibrated. At No. 1-J. Friemel, the water may represent a mixture of equilibrated deep-basin brine and leakage waters. At the No. 1-Mansfield well, most of the brines originated as leakage waters.

Lithologic Evidence of Aquitard Leakage

The flow of water through an evaporite aquitard potentially can occur along three different pathways: 1) intracrystalline (through the crystal), 2) intercrystalline (along grain boundaries), and 3) along major discontinuities such as fractures and faults. Intracrystalline flow may be occurring as evidenced by recrystallization textures in halite as observed by Dix and Jackson (1982), but is not considered important to overall aquitard leakage because the permeability of a crystal is estimated to be several orders of magnitude less than intercrystalline or fracture permeability. A better understanding of the latter two pathways can be gained from petrographic and geochemical investigations of core from Middle and Upper Permian evaporite rocks in the Evaporite aquitard and from structure studies of the tectonic deformation of the Palo Duro Basin.

Within a halite bed, characteristic minor facies represent a subcycle of halite precipitation and subsequent dissolution. A subcycle in halite exhibits a vertical sequence of bedded anhydritic halite at the base and altered halite and mudstone interbeds at the top. Each facies exhibits some evidence of salt dissolution and possible recrystallization. The bedded halite

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contains individual salt crystals having original chevron banding that are truncated by dissolution surfaces. Recrystallization occurs along the grain boundaries. Diagenetic anhydrite and dolomite are found along the grain boundaries as well as within the crystals. The interbedded mudstones were introduced during episodes of subaerial exposure. Karstic features, such as pits, pipes, and vertical fractures, extend downward from the mud-rich zones into the anhydritic halite (Hovorka, 1983a; Fisher and Hovorka, 1984).

These pervasive salt dissolution and possible recrystallization features are predominantly syndepositional or early diagenetic in their timing. The dissolution features are restricted to individual evaporite cycles. Karstic dissolution events did not disrupt overlying beds (Hovorka, 1983a). Geochemical parameters, such as Br, show similar cyclic changes in concentration from evaporite cycle to evaporite cycle, suggesting that Br variations in the San Andres are related to syndepositional processes and not to an influx of postdepositional meteoric water (Fisher and Hovorka, 1984).

Post-Permian meteoric waters have penetrated the top of the Evaporite aquitard. A variety of relict fabrics and collapse features indicate post-Permian dissolution of the shallowest evaporite strata (Hovorka, 1983b). Recrystallized halite occurs locally as much as 30 m beneath the post-Permian dissolution zone, suggesting that meteoric waters have penetrated that distance into the Evaporite aquitard.

Recrystallization at the top of the salt section suggests postdepositional movement of ground water into the salts. Given our present understanding of the petrography and geochemistry of salt, dissolution and recrystallization elsewhere in the section appear to result from syndepositional or early diagenetic flow of water. The critical unanswered question is what geochemical and textural changes would be observed if post-Permian waters were moving through the evaporites.

Fluid movement through the Evaporite aquitard may be occurring predominantly along fractures rather than along grain boundaries. Based on outcrop study and core and fracture log

analyses from test wells, Collins (1984) identified oriented joint sets in nearly all the strata tested or analyzed within the Palo Duro Basin. Jointing is present in the Permian evaporites as well as the Mesozoic strata above and Paleozoic strata below the evaporites. Fractures have also been identified in the Permian Salado Formation in the Delaware Basin (Chaturvedi, 1984). Joints in the Tertiary Ogallala Formation are not common. Joint systems probably result from the structural downwarping in the Paleozoic and subsequent uplift during the Cretaceous. The degree of fracturing is greatest toward the Amarillo Uplift, where structural deformation is also the greatest. Analyses of fracture logs and core show the density of fracturing decreases from the No. 1-Mansfield well, to the No. 1-J. Friemel well and to the No. 1-Zeeck well, which has the lowest density of fractures (Collins, 1984). The No. 1-Mansfield well is adjacent to the Amarillo Uplift, the No. 1-J. Friemel well is along a structural trend off the Amarillo Uplift, and the No. 1-Zeeck well is farthest from the Uplift and major structural deformation (fig. 1). Even though the No. 1 Sawyer well is close to the Amarillo Uplift, minimal fracturing is observed in the core and the fracture identification logs.

The brines from the No. 1-Mansfield well have the largest component of meteoric water, No. 1-J. Friemel the next, and No. 1-Zeeck the least. This coincidence of fracture density with the amount of meteoric water in the deep-basin brines suggests that meteoric waters are leaking down fractures. In the region of the No. 1-J. Friemel well, Gustavson and Budnik (1984) observed a coincidence between fracture orientation and a trough of enhanced salt dissolution at the top of the salt strata. They concluded that a thinning of salts may have resulted from more rapid dissolution in the northeast-trending fracture zone. Fluid flow down fractures in the evaporites may be the dominant process for moving waters through the aquitard. Bredehoeft and others (1982) reached this same conclusion in their discussion of fluid flow through the Pierre Shale; fracture flow is the primary pathway for fluid flow through aquitards.

CONCLUSIONS

Several conclusions can be drawn from this study.

1. Numerical modeling indicates that the Permian evaporite strata in the Palo Duro Basin can function as a leaky aquitard. The aquitard effectively separates the shallow Upper aquifer from the Deep-Basin Brine aquifer, and fluid pressures in the Deep-Basin Brine aquifer are not in hydrostatic equilibrium with the Upper aquifer. Leakage through the aquitard may contribute significant quantities of water to the deep aquifers because of the large lateral extent of the aquitard. Average linear velocities are estimated to be 6×10^{-5} m/day. The permeability range for the Evaporite aquitard calibrated from the numerical ground-water models (8×10^{-5} to 2.8×10^{-4} md) is in general agreement with laboratory investigations of bedded salt permeability (10^{-4} md).
2. The chemical and isotopic compositions of the deep-basin waters suggest leakage through the evaporite strata. Leakage through the Evaporite aquitard is considered the most logical explanation of the presence of meteoric water or a component of meteoric water at depths down to 8,000 feet.
3. Many of the primary textures of the halite have been altered by dissolution and minor recrystallization. Alteration probably occurred syndepositionally or in early postdepositional time. Post-Permian waters may have altered only the upper part of the Evaporite aquitard. However, the extent of an overprint of post-Permian dissolution and recrystallization is not yet understood.
4. Fluid flow through the aquitard may be along fractures rather than as diffuse flow along grain boundaries. Areas having the greatest tectonic deformation therefore should have the greatest potential to allow leakage. This may explain why some deep-basin brines are in isotopic equilibrium with the host rock whereas others are not.

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Table 1. Hydrogeologic units in the Palo Duro Basin (from Bassett and Bentley, 1983).

System	Series	Group	General lithology and depositional setting	Hydrogeologic element	Hydrogeologic unit
Quaternary			Fluvial and lacustrine clastics	Ogallala aquifer	Upper aquifer
Tertiary					
Cretaceous					
Triassic		Oockum	Fluvial deltaic and lacustrine clastics and limestones	Oockum aquifer	
Permian	Ochoa		Salt, anhydrite, red beds and peritidal dolomite	Evaporite aquitard	Evaporite aquitard
	Guadalupe	Artesia			
		Pease River			
	Leonard	Clear Fork			
		Wichita			
	Wolfcamp		Shelf and platform carbonates, basin shale and deltaic sandstones	Wolfcamp carbonate aquifer	Deep-basin brine aquifer
Pennsylvanian				Pennsylvanian carbonate aquifer	
				Basin shale aquitard	
				Upper Paleozoic granite wash aquifer	
Mississippian			Shelf limestone and chert	Lower Paleozoic carbonate aquifer	Basin shale aquitard
Ordovician		Ellenburger	Shelf dolomite		
Cambrian			Shallow marine (?) sandstone		
Precambrian			Igneous and metamorphic	Basement aquiclude	Basement aquiclude

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Table 2. Chemical and isotopic composition of waters collected from Department of Energy - Stone and Webster Engineering Corporation test wells in the Deep-Basin Brine aquifer, Palo Duro Basin. Chemical analyses performed by Mineral Studies Laboratory, Bureau of Economic Geology, The University of Texas at Austin. Isotopic analyses performed by Global Geochemistry, Inc. Additional chemical analyses are in Fisher (1984).

Well Zone	Sawyer #1 Ellenburger Carbonate-Sand.	Sawyer #1 U. Ellenburger Carbonate	Sawyer #1 Mississippian Limestone	Sawyer #1 Pennsylvanian Granite Wash	Sawyer #1 Permian Wolfcamp	Mansfield #1 Permian Wolfcamp
Depth (ft)	4,716	4,604	4,500	4,228	3,172	4,313
T (°C)	39.0	39.0	39.0	39.0	35.0	38.0
pH	5.35	7.5	5.4	4.4	6.1	5.6
Ca (mg/l)	22,900	21,600	19,400	18,300	6,760	5,960
Mg (mg/l)	2,600	2,810	2,310	2,180	2,490	1,320
Na (mg/l)	60,200	64,000	61,000	63,400	43,450	78,400
K (mg/l)	459	420	337	332	118	369
Sr (mg/l)	660	710	574	563	113	99
HCO ₃ (mg/l)	560	40	106	26	144	131
SO ₄ (mg/l)	169	230	237	336	2,010	1,220
Cl (mg/l)	139,000	133,000	132,000	136,000	86,300	138,000
Cl/Br (g/g)	296	271	227	237	261	1,290
Na/Cl (g/g)	.43	.48	.46	.46	.50	.57
δ ¹⁸ O (SMOW)	-1.73	-0.83	-0.61	-1.19	+0.72	-5.78
δD (SMOW)	-27.0	-13.0	-18.0	-15.0	-23.0	-43.0
δ ³⁴ S			+12.86	+13.77	+10.37	+17.35
δ ¹³ C (POB)	+5.14	-2.25	+3.42	+0.12	+1.83	+2.63

Well Zone	Mansfield #1 Permian Wolfcamp	Zeeck #1 Pennsylvanian Carbonate	Zeeck #1 Permian Wolfcamp	Zeeck #1 Permian Wolfcamp	J. Friemel #1 Granite Wash	J. Friemel #1 Granite Wash
Depth (ft)	4,514	7,140	5,603	5,474	8,168	8,122
T (°C)	41.0	56.0	38.0	41.0	55.0	55.0
pH	4.8	5.8	6.9	6.25	5.9	5.9
Ca (mg/l)	6,020	11,950	6,780	6,240	15,700	15,400
Mg (mg/l)	1,600	1,550	1,290	1,320	2,110	2,130
Na (mg/l)	76,100	76,350	61,700	69,300	77,930	78,230
K (mg/l)	380	197	314	339	620	610
Sr (mg/l)	90	1,380	250	110	444	420
HCO ₃ (mg/l)	161	31	113	94	35	69
SO ₄ (mg/l)	1,320	81	977	1,650	440	487
Cl (mg/l)	131,000	151,500	105,000	120,000	148,500	148,100
Cl/Br (g/g)	1,110	188	231	234	922	1,028
Na/Cl (g/g)	.58	.50	.59	.58	.52	.53
δ ¹⁸ O (SMOW)	-5.12	+0.73	+0.80	+2.17	-2.88	-3.53
δD (SMOW)	-42.0	-14.0	-23.0	-23.0	-28.0	-35.0
δ ³⁴ S	+16.6	+12.58		+13.3	+9.76	+9.77
δ ¹³ C (POB)	+1.83	+3.31		+2.76		-6.92

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Table 2. (continued)

Well Zone	J. Friemel #1 Granite Wash	J. Friemel #1 Granite Wash	J. Friemel #1 Granite Wash	J. Friemel #1 Pennsylvanian Carbonate	J. Friemel #1 Wolfcamp
Depth (ft)	8,040	7,895	7,707	7,300	5,825
T (°C)	55.0	55.0	53.0	52.0	38.0
pH	6.0	6.2	5.2	6.4	4.2
Ca (mg/l)	15,050	17,120	19,330	12,790	8,390
Mg (mg/l)	2,180	2,560	3,020	2,560	2,160
Na (mg/l)	77,300	77,670	87,580	78,460	72,470
K (mg/l)	636	578	551	530	589
Sr (mg/l)	410	451	571	363	150
HCO ₃ (mg/l)	71	21	43	73	205
SO ₄ (mg/l)	492	519	378	631	1,373
Cl (mg/l)	147,420	156,900	178,500	150,800	129,770
Cl/Br (g/g)	1,053	581	649	973	882
Na/Cl (g/g)	.52	.50	.49	.52	.56
δ ¹⁸ O (SMOW)	-3.46		-2.03	-2.7	-3.7
δD (SMOW)	-34.0		-18.	-31.0	-36.0
δ ³⁴ S	+9.86	+10.12	14.7	+13.9	+14.5
δ ¹³ C (PDB)	-6.51	-6.65	-5.7	+2.4	+3.0

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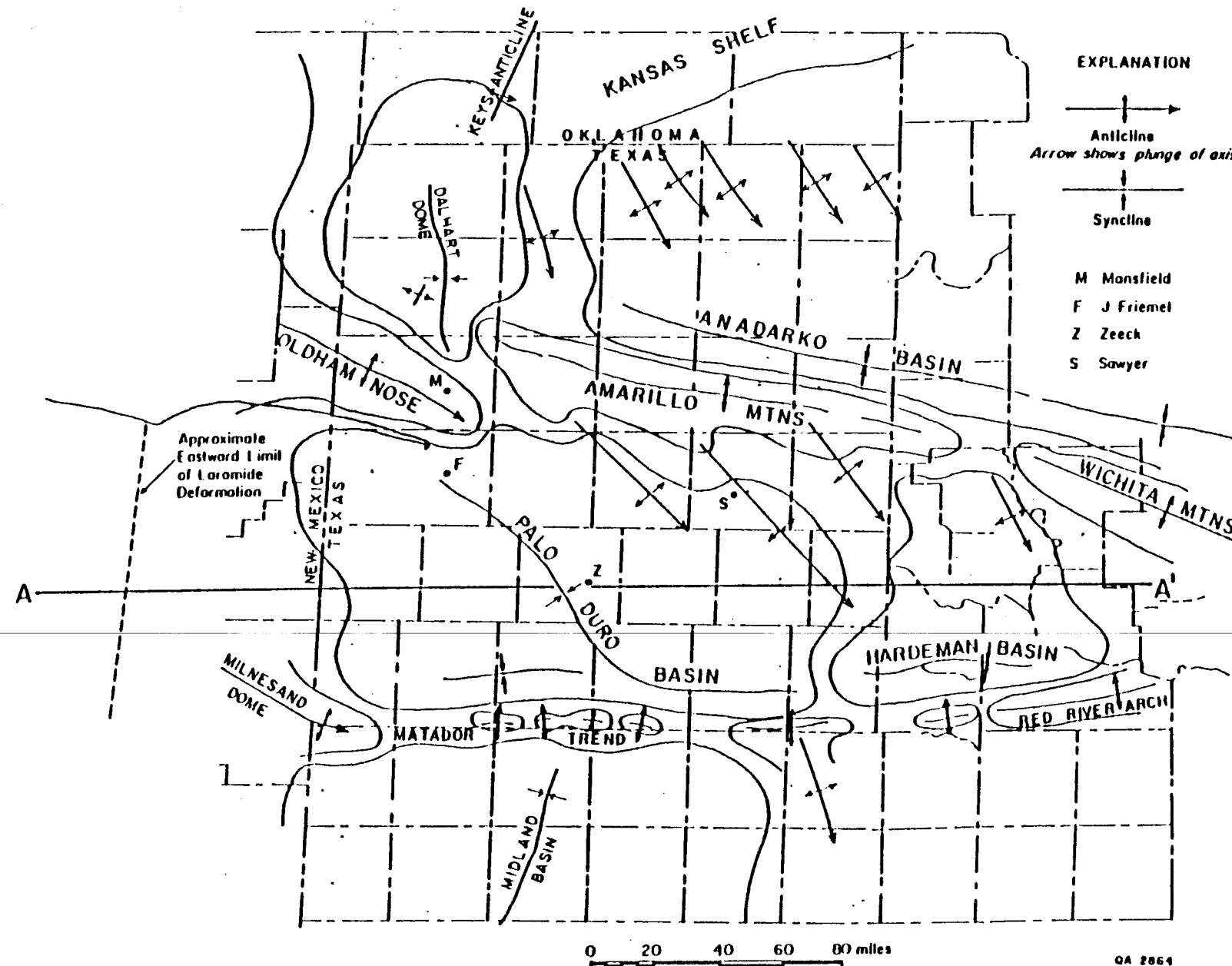


Figure 1. Structural setting of the Palo Duro Basin (modified from Nicholson, 1960).

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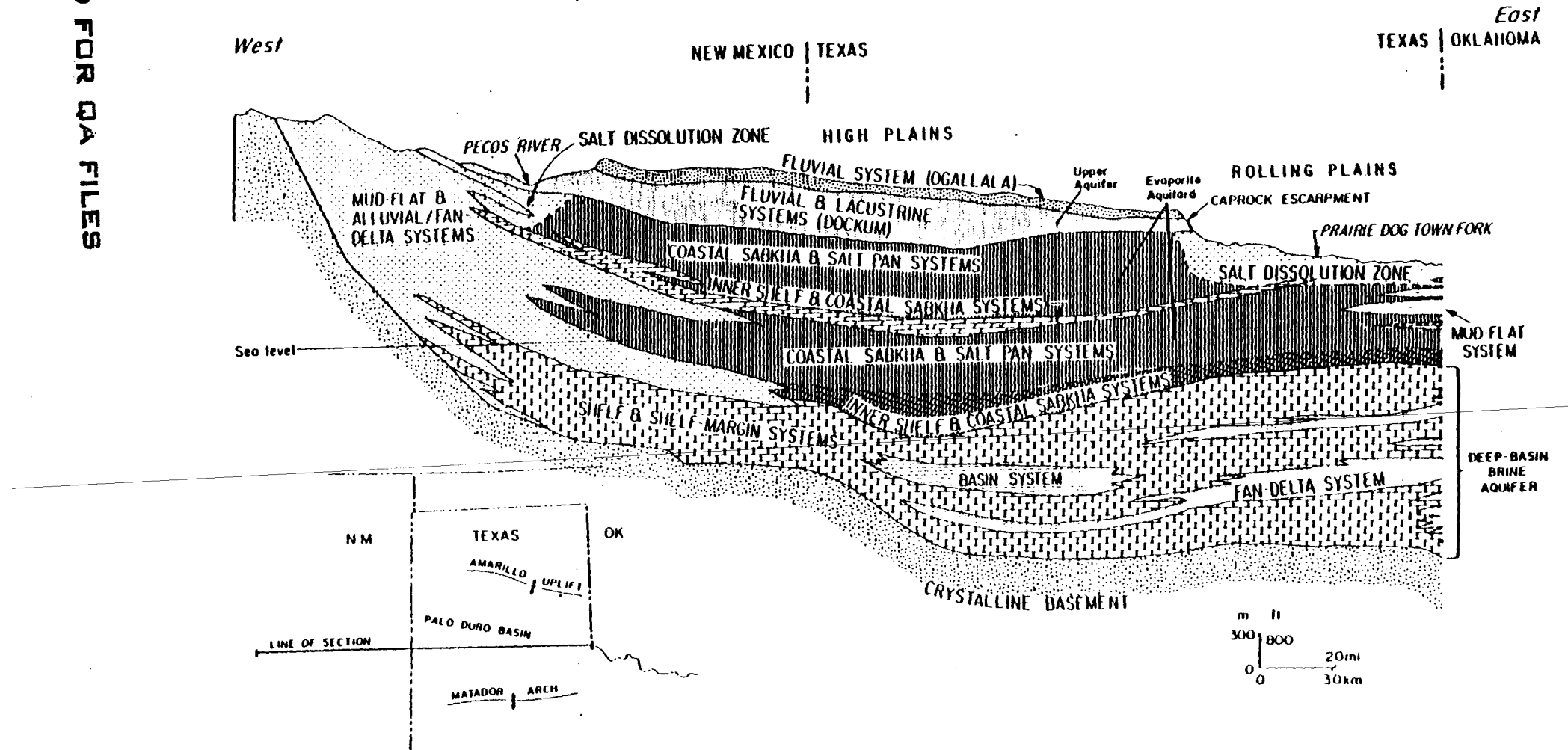


Figure 2. West-east cross section of Palo Duro Basin showing distribution of major geologic/hydro-logic facies (modified from Bassett and Bentley, 1983).

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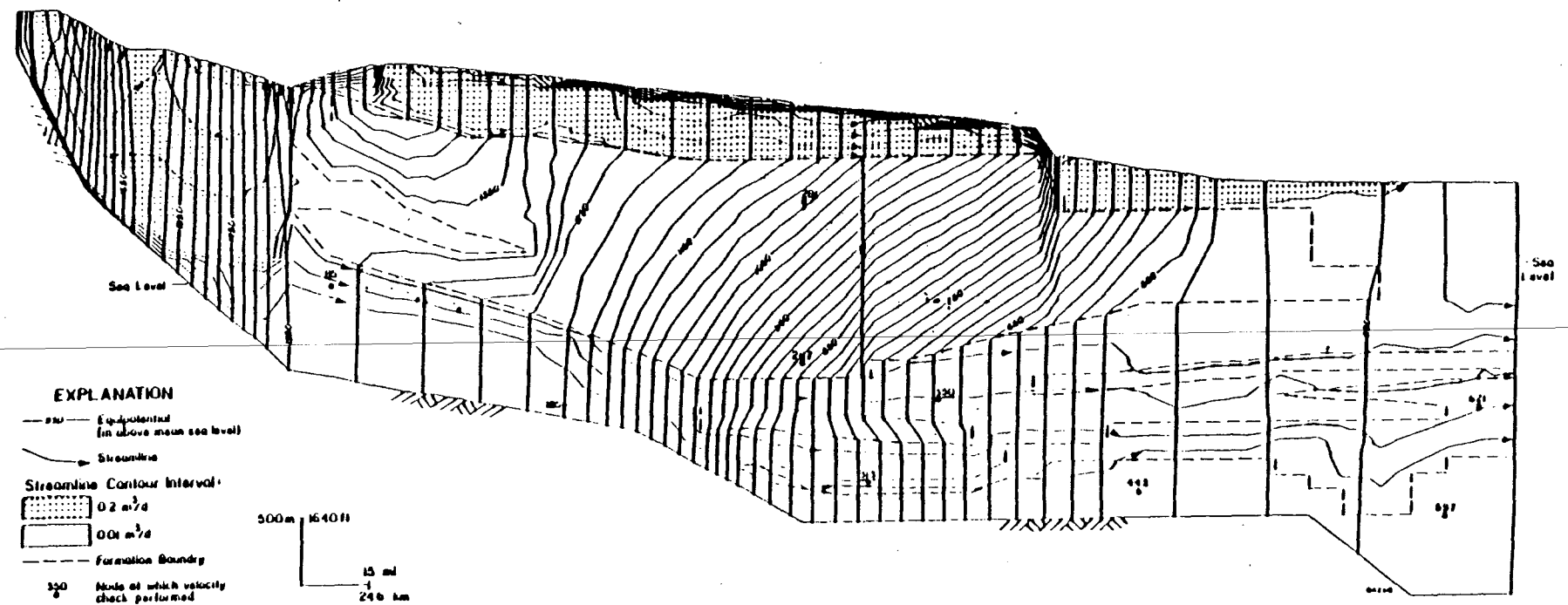


Figure 3. Modeled ground-water flow through west-east vertical section of Palo Duro Basin, showing equipotential contours and flow lines (Senger and Fogg, 1984). Note flow line through Evaporite aquitard. Two contour intervals for flow tubes had to be used to depict both shallow and deep flow systems.

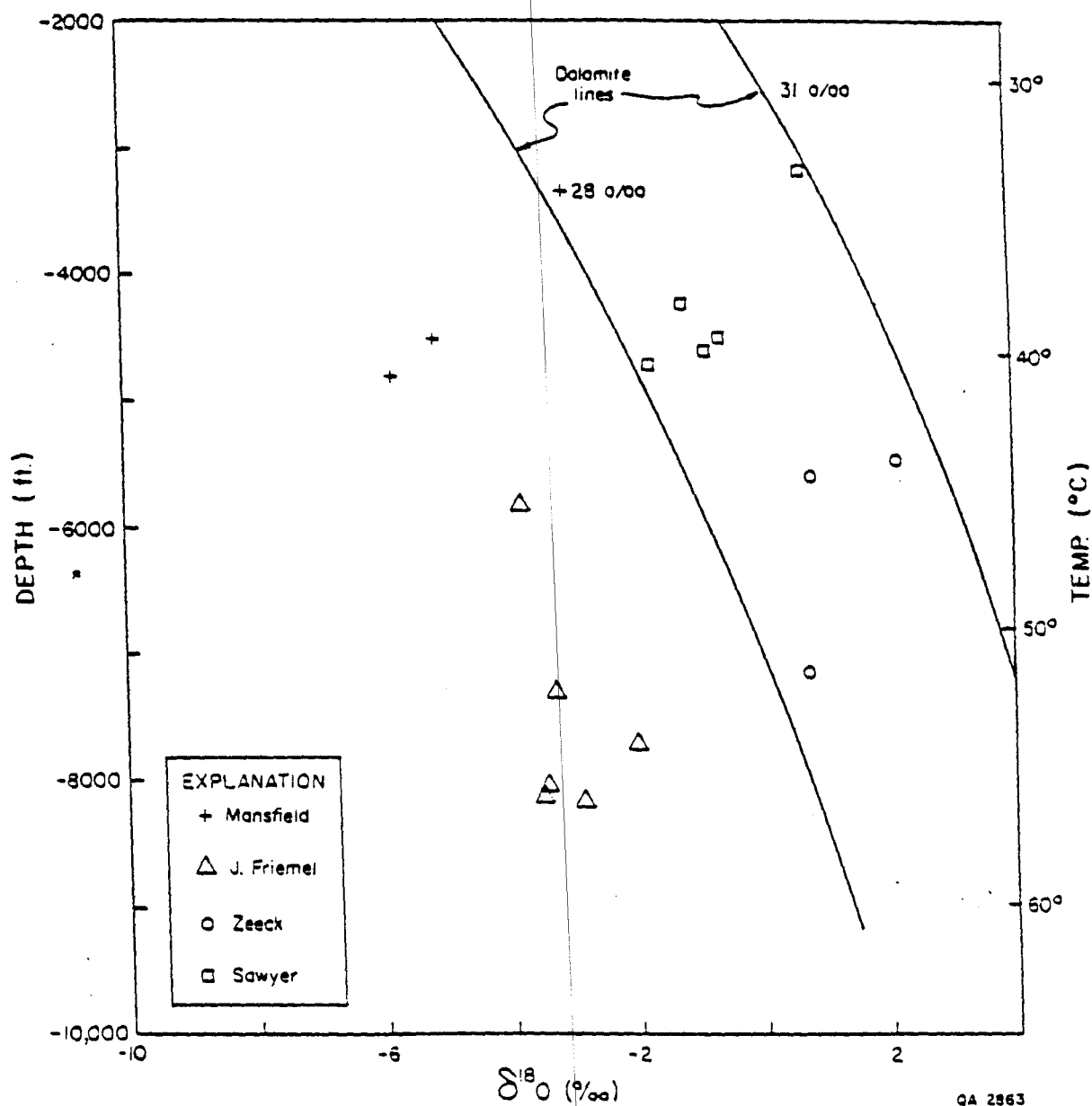
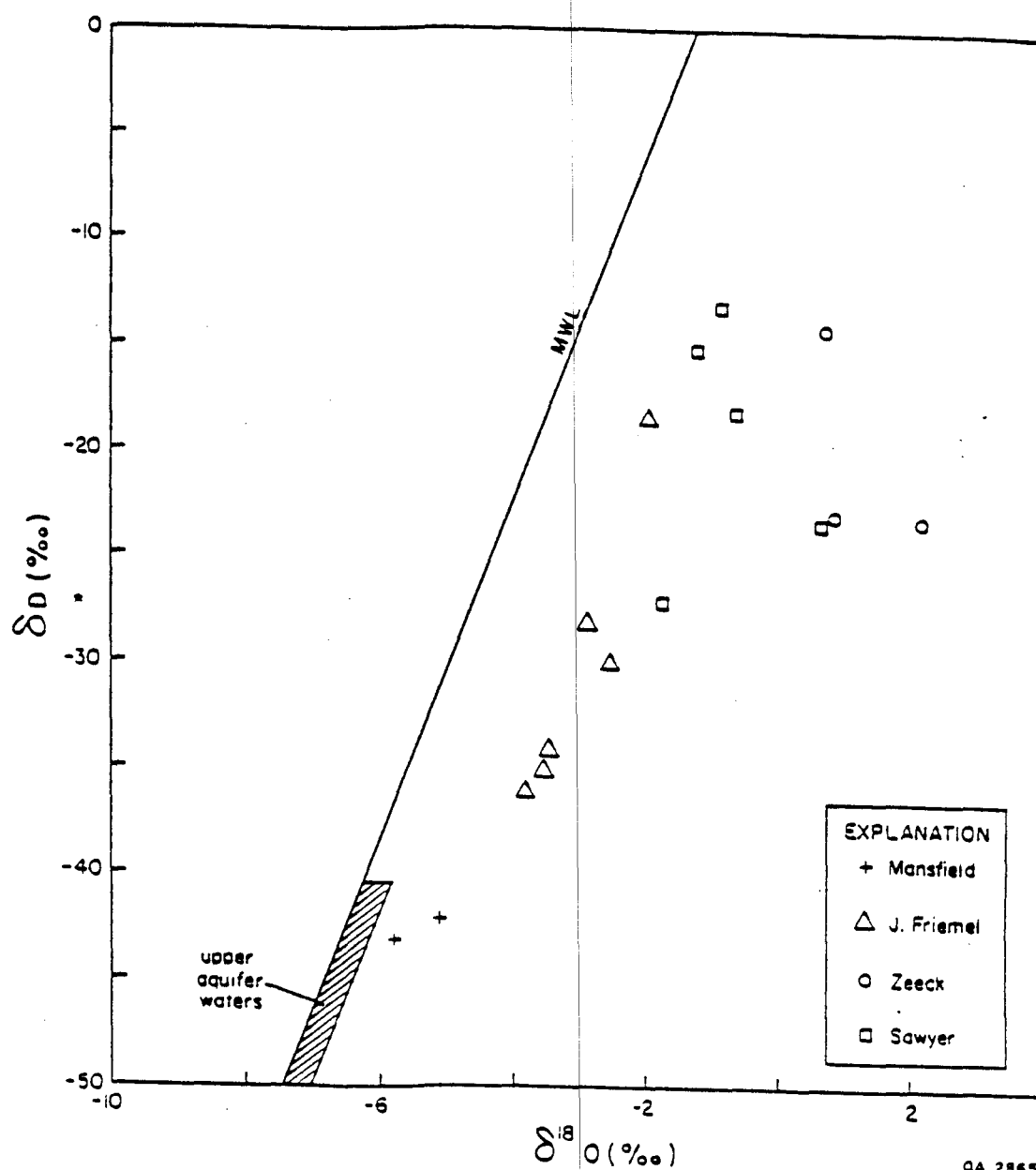


Figure 4. $\delta^{18}O$ versus depth (temperature) for water samples from Deep-Basin Brine aquifer, Palo Duro Basin. Plot shows waters from No. 1-Mansfield and No. 1-J. Friemel wells not in isotopic equilibria with dolomite equilibria curves. Waters from No. 1-Zeeck and No. 1-Sawyer approach or are in isotopic equilibrium with dolomite. The +28 line and +31 line (relative to SMOW standard) represent a range of expected dolomite equilibria values (from Land, 1982). Temperature scale based on measured bottom-hole temperatures in DOE test wells, Data in Table 2. Well locations on Figure 1.

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Figure 5. $\delta^{18}O$ versus δD for waters from Deep-Basin Brine aquifer. Isotopic composition of waters from Upper aquifer from Senger and others (in press). MWL is Meteoric Water Line. Data in Table 2. Well locations on Figure 1.

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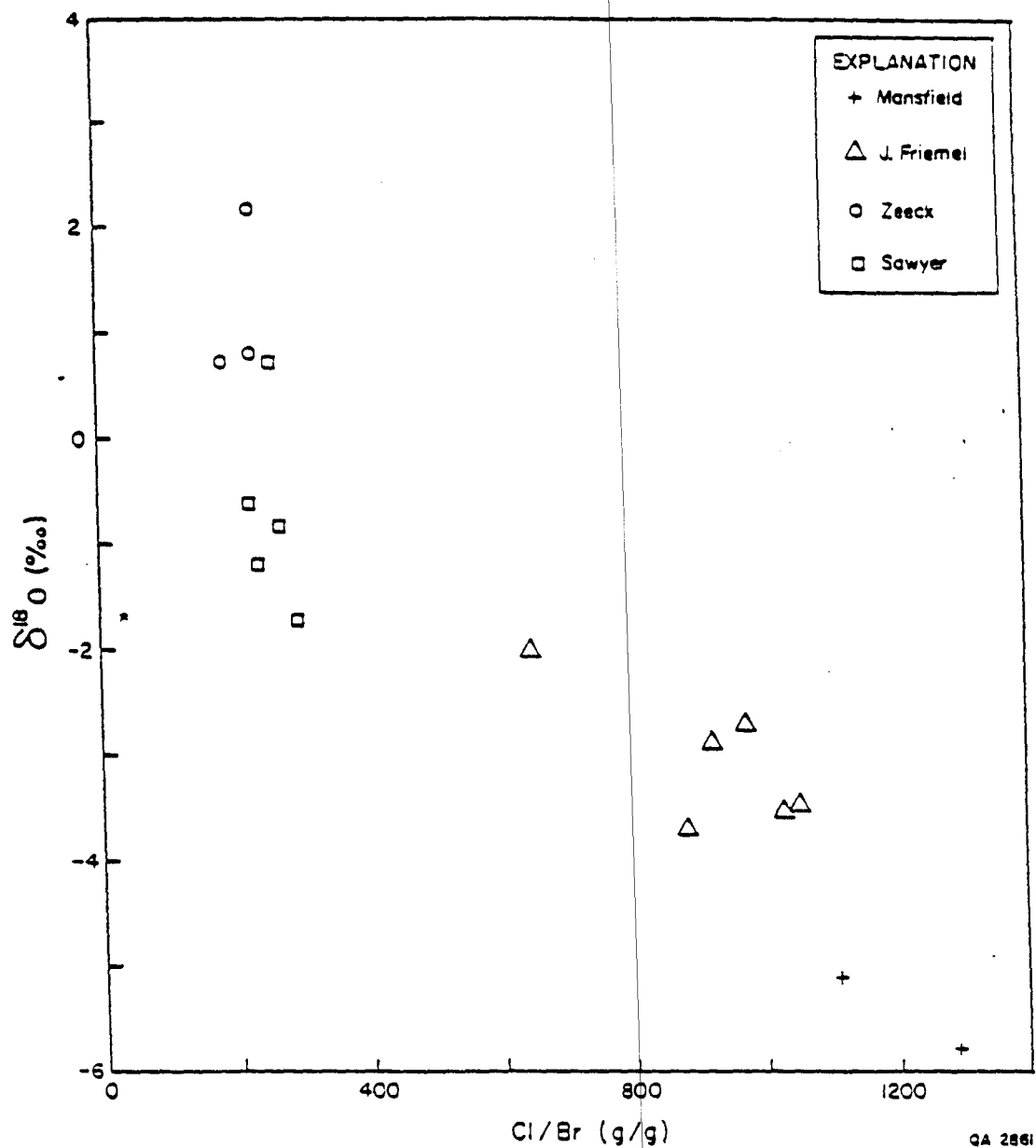


Figure 6. Cl/Br weight ratio versus $\delta^{18}\text{O}$ for water samples from Deep-Basin Brine aquifer. Note the inverse relationship between Cl/Br and $\delta^{18}\text{O}$ suggests a mixing between waters from different sources. Data in Table 2. Well locations on Figure 1.

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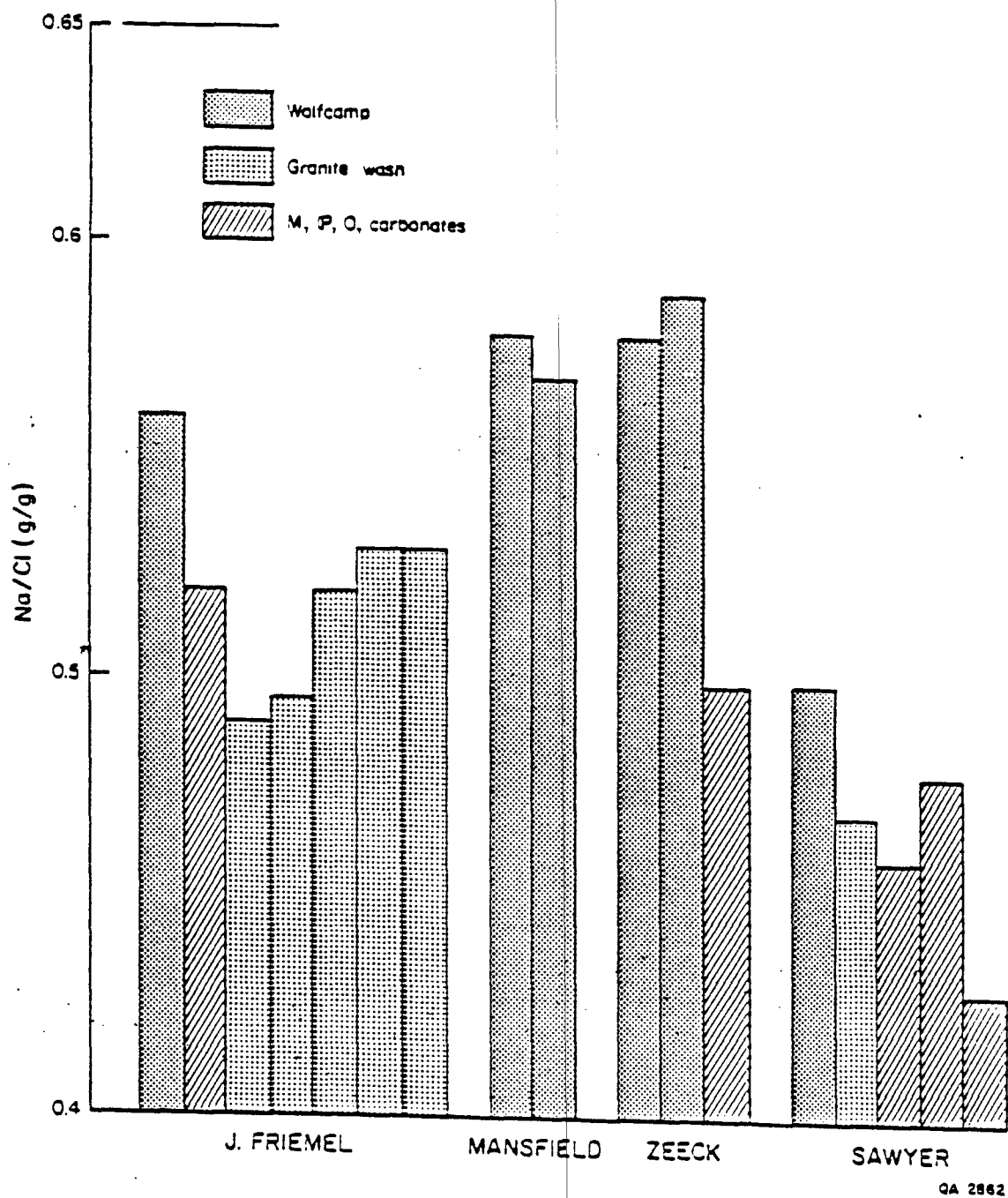


Figure 7. Na/Cl weight ratios for water samples from Deep-Basin Brine aquifer. Waters from Wolfcamp carbonates consistently have higher Na/Cl ratios. Data in Table 2. Well locations on Figure 1.

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FIGURES

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