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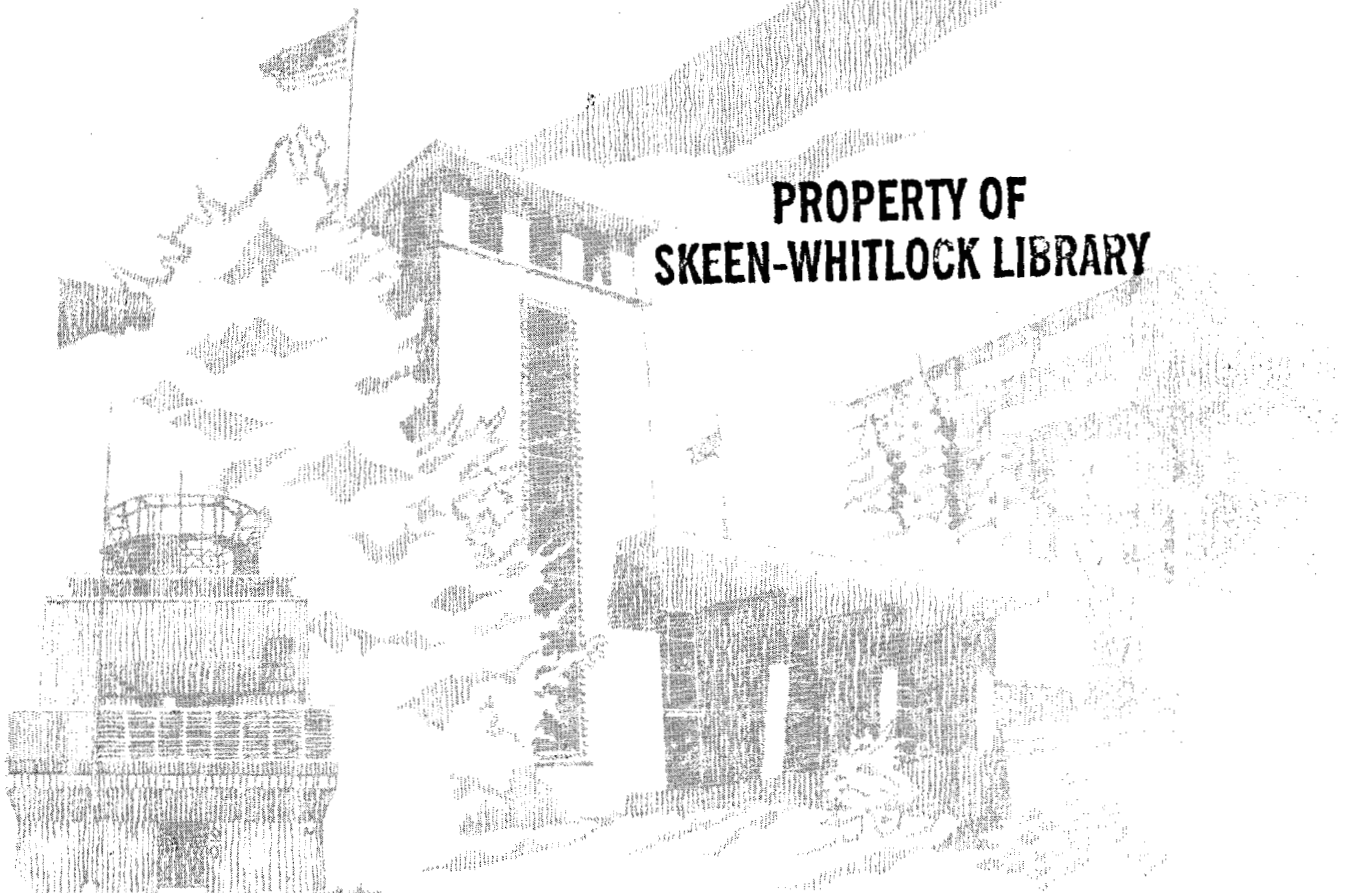
Dissolution of Evaporites in and Around the Delaware Basin, Southeastern New Mexico and West Texas

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Prepared by
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Albuquerque, New Mexico 87185 and Livermore, California 94550
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Abstract

Permian evaporites in the Ochoan Castile, Salado, and Rustler Formations in the Delaware Basin of southeast New Mexico and west Texas have been subjected to various degrees of dissolution (notably of halite and gypsum) through geologic time. Eastward tilting of the Delaware Basin has resulted in the exhumation and erosion of Ochoan rocks in the western part of the basin. Waters in the Capitan, Rustler, Castile, and Bell Canyon Formations have previously been proposed as agents or consequences of evaporite dissolution according to four principal models: solution-and-fill, phreatic dissolution, brine density flow, and stratabound dissolution (along bedding planes). Several geomorphological features of positive and negative relief have previously been cited as indicators of evaporite dissolution. Brine density flow has been used to explain the selective dissolution of certain evaporite horizons during the late Cenozoic. A review of available geological data has revealed that

- Halite deposition was probably not so extensive as formerly believed
- Waters with potential to dissolve evaporites are in the Rustler and Capitan, but not in the Bell Canyon, Salado mine seeps, or the Castile brine reservoirs
- Brine density flow has not been active in removing most of the "missing" halite, nor are "point-source" dissolution features likely to have their roots at the Bell Canyon
- Major evaporite dissolution has not been confined to the late Cenozoic, but much of it took place during the Permian, Triassic, Jurassic, and Tertiary periods
- The Bell Canyon Formation has not been a sink for dissolution-derived brine

Abstract (cont)

Stratabound dissolution is an efficient process for the removal of evaporites, and is well exemplified in Nash Draw. This process entails downdip migration of meteoric water within beds of competent fractured rock, with upward and downward excursions of the water into adjacent halite-bearing beds. The chief weakness in the stratabound model for dissolution is the as-yet-unidentified sink for dissolution brine. If the stratabound model of dissolution is active in removal of lower Salado halite, the threat of dissolution to the WIPP in the next 250 000 yr is comparable to the threat to the same area posed by the growth of Nash Draw during the past 600 000 yr. The regional geological history showed the past threat to be negligible.

Preface

For the past 6 yr, the Waste Isolation Pilot Plant (WIPP) proposed for the vicinity of Los Medaños in southeastern New Mexico has spawned a great deal of controversy. Various groups and individuals have questioned not only the suitability of the Los Medaños locality in particular, but of bedded deposits of rock salt in general, for siting of any radioactive waste storage facility, regardless of the nature of the waste or the required time of isolation. Unfortunately, much of the argumentation (both scientific and uninformed) about dissolution is fraught with unwritten speculation. The intent of this work is not to address all of this unwritten material, but to consider only those arguments appearing as scientific documentation prepared by scientists and distributed through official public agencies and the professional literature.

For various reasons, several workers with experience in dissolution studies have not published all their arguments. This is understandable; working hypotheses may change rapidly with time. To keep pace with the state of the arguments, I have resorted to *some* personal communication. A field trip (at which the author was not present) was held June 16-18, 1980, to "further clarify the different views on the geological processes active at the [WIPP] site [and vicinity]." I am indebted to Lokesh Chaturvedi of the New Mexico Environmental Evaluation Group, who prepared an excellent summary of that trip, thoroughly documenting the relevant discussion and arguments at each field trip stop. Much of that material exists in no other form than his document, EEG-7.

This report was prepared in response to a request by the State of New Mexico for a "detailed review paper . . . specifically addressing Roger Anderson's hypothesis about extensive deep dissolution in the lower part of the Ochoan evaporite deposits in the Delaware Basin." The report is, however, intended as much more than that; it is primarily intended as a critical but constructive assembly of the arguments I consider relevant to evaporite dissolution in general. It also contains a significant amount of original work. To meet the criticism of previously proposed models, a model of dissolution is presented that is consistent with all available data.

Acknowledgments

First, I thank Leslie Hill, P. D. ("Pete") Seward, Robert ("Uncle Bob") Statler, and Wendell Weart, who provided so much encouragement so essential in the organization of this work. I have had many helpful discussions with George Barr, Lawrence Barrows, and Dennis Powers. I received moral support and valuable thought-contributions from George Bachman, L. M. ("Bud") Gard, Charles Jones, Jerry Mercer and Richard Snyder, all with the United States Geological Survey, during much of this work. Sanford ("Eric") Erickson and John Golden provided analytical support and some new data contained herein, while stable-isotope measurements are courtesy of James O'Neil of the USGS. David Borns, Terri Ortiz, and Sue-Ellen Shaffer participated in interpretation of geophysical logs. Karen Robinson and Sue Shaffer assisted in the preparation of some of the illustrations. I am indebted to Rosalyn Baca, Carmen DeSouza, and E. Roberta Voelker for their assistance in the final preparation of the manuscript.

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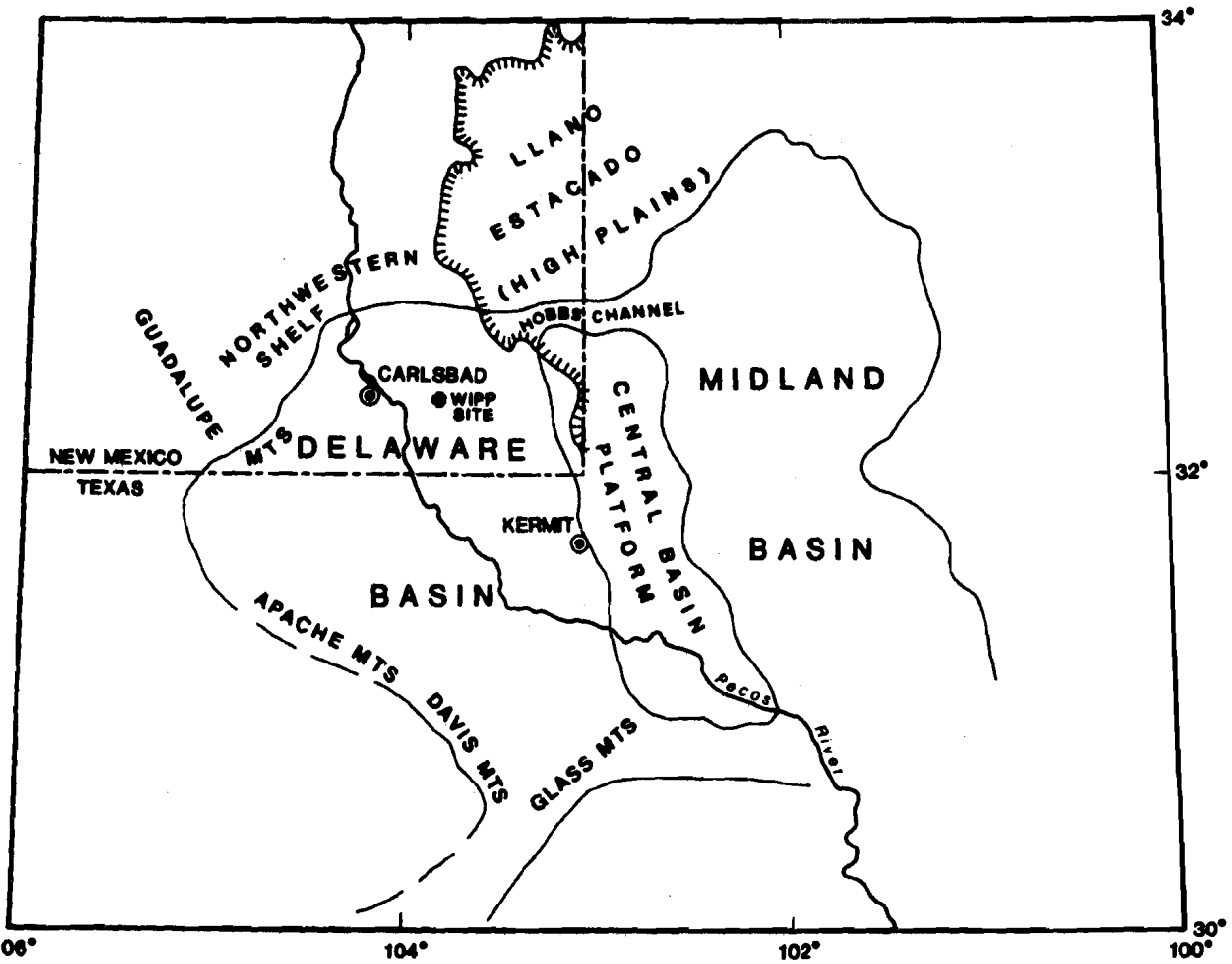


Figure I-1. Major Geographical and Tectonic Subdivisions in Southeastern New Mexico and West Texas (modified after McKee, Oriel, et al, 1967). The boundary of the Delaware Basin is here adopted as the *shelfward* limit of the Capitan limestone.

Chapter II. Models of the Dissolution Process

Previous studies of dissolution features and processes leading to them have given rise to several models of various geometries, some of which are inconsistent with one another. Associated with each model is the time span over which it was active.

Erosion by solution-and-fill, a model described by Lee (1925), has been proposed to explain shallow dissolution of gypsum terrain. Essential characteristics in solution-and-fill are incision of dendritic drainage patterns and development of underground drainage channels that collapse as dissolution removes material. The resultant irregular system of channels just below the surface is generally restricted to the soluble substrate; gradients increase downstream with a consequent headward-cutting. A chaotic jumble of less soluble rock is the end product. Bachman (1980) has proposed that erosion by solution-and-fill is active in certain surface exposures of gypsum terrain in southeastern New Mexico.

Phreatic dissolution requires proximity of soluble rock (limestone, anhydrite, gypsum, halite) to open conduits filled with freely moving fresh water (Bachman, 1980). The result is large open cavities greater than a few tens of feet below the surface. Open space may be maintained (such as in the Carlsbad Caverns), or collapse may result (such as Christiansen, 1971, proposed for the formation of the Crater Lake collapse structure resulting from salt dissolution in southeastern Saskatchewan).

Phreatic dissolution has led to spectacular development of caverns in the Guadalupe Mountains. Jagnow (1979) has suggested that caverns developed largely through the agency of sulfuric acid derived

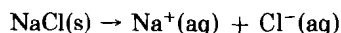
from the oxidation of iron sulfides in the lagoonal facies west of the Guadalupe Mountains. Jagnow proposed that the reaction of limestone with sulfuric acid left the massive and banded gypsum deposits observed in many of the caverns, and that the gypsum is not the result of extensive evaporite remobilization.

"Stratabound dissolution" refers here to the "blanket" or "regional" dissolution described by Bachman (1980). This mechanism involves lateral movement of water along a permeable bed, adjacent to which soluble rock dissolves. The process is said to be self-perpetuating, the dissolution of rock making space for the advance of the undersaturated solutions. Anderson (1981) has proposed that stratabound dissolution has removed 50% of the halite in the Delaware Basin and that this mechanism is still active, particularly in the lower part of the evaporite sequence.

Dissolution by brine density flow is a model proposed by Anderson and Kirkland (1980) to account for the domal features in the Delaware Basin. It is said to begin with a point-source reservoir of unsaturated solution underneath the evaporites. As the solution dissolves evaporites and becomes more dense, it sinks to the lower parts of the reservoir, allowing continuous contact of unsaturated solutions with the evaporites. Thus, linear vertical dissolution features are thought by Anderson (1981) to actively stope their way upwards through the evaporites throughout the Delaware Basin. Anderson (1981) has also invoked the brine density flow model as a triggering agent for stratabound dissolution in the lower part of the Delaware Basin evaporites.

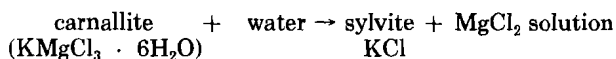
Chapter III. The Soluble Nature of Evaporite Minerals

Although evaporite mineral deposits, in general, are regarded as chemical sediments precipitated from aqueous solution, segregating aqueous ions into crystals is not simply a matter of removing water. During precipitation accompanying evaporation of a mother liquor, a previously deposited evaporite mineral phase may not remain in thermodynamic equilibrium with the fluid surrounding it, since the composition of the fluid may change during crystallization of other phases. Such a disequilibrium can give rise to peritectic relationships, in which the primary phase reacts with the solution to produce a new phase. In the reverse of precipitation (dissolution), a mineral phase is surrounded by a solution with which it is not in equilibrium. Many evaporite minerals (e.g., halite) will dissolve to give aqueous species whose chemical composition is identical to the chemical composition of the solid phase:



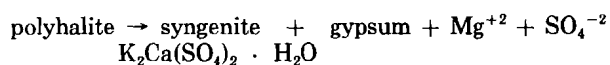
Such minerals are said to dissolve *congruently*.

Another class of minerals, upon contact with a fluid, will give rise to aqueous species in proportions different from those in the solid (incongruent solubility). A familiar example is the alteration of carnallite, illustrated as follows (Braitsch, 1971):



It is believed (Braitsch, 1971) that carnallite (and not sylvite) is a primary precipitate from evaporating seawater. The abundance of sylvite (and not carnallite) in the Delaware Basin potash ore zones is attributable either to (1) the secondary formation of sylvite as an alteration product of primary precipitates, or (2) the primary precipitation of sylvite from a mother liquor deficient in magnesium, which implies a solution other than seawater. Braitsch (1971) has suggested that examples of both mechanisms of sylvite formation occur in various salt deposits.

A more relevant example of an incongruently soluble evaporite mineral is polyhalite, $\text{Ca}_2\text{K}_2\text{Mg}(\text{SO}_4)_4 \cdot 2\text{H}_2\text{O}$. Perthuisot (1971) has suggested that the "rain weathering" of polyhalite takes place as follows:



The readily soluble syngenite dissolves incongruently, leaving gypsum and a solution enriched in magnesium, potassium, and sulfate.

The most important soluble evaporite minerals, because of their abundance in the Delaware Basin, are anhydrite, gypsum, halite, sylvite, polyhalite, and glauberite, $\text{Na}_2\text{Ca}(\text{SO}_4)_2$. Of these, the first four dissolve congruently. Polyhalite dissolves incongruently to give a residue of gypsum or anhydrite with a solution enriched in potassium, magnesium, and sulfate. Glauberite dissolves incongruently to give a residue of gypsum or anhydrite with a solution enriched in sodium and sulfate. These relationships will be important in a subsequent discussion.

The presence of certain species already in solution can profoundly influence the solubility of a mineral phase not in equilibrium with the solution. One example of this influence is a reduction in solubility of one phase by the addition of one of its constituents to the solution (the common-ion effect). In solutions of strong electrolytes, because the activity coefficients of the ions are hardly ever unity, linearity of the common ion relationship is not to be expected. For instance, equilibrium in the system $\text{NaCl-KCl-H}_2\text{O}$ (Figure III-1) shows that the presence of additional chloride from either halite or sylvite in solution will diminish the solubility of the other mineral, but a straight mixing line does not describe the system. Thus, a solution saturated only with halite, upon contact with solid sylvite, will "salt out," precipitating halite until the equilibrium mixture of Na^+ , K^+ , and Cl^- in solution is attained.

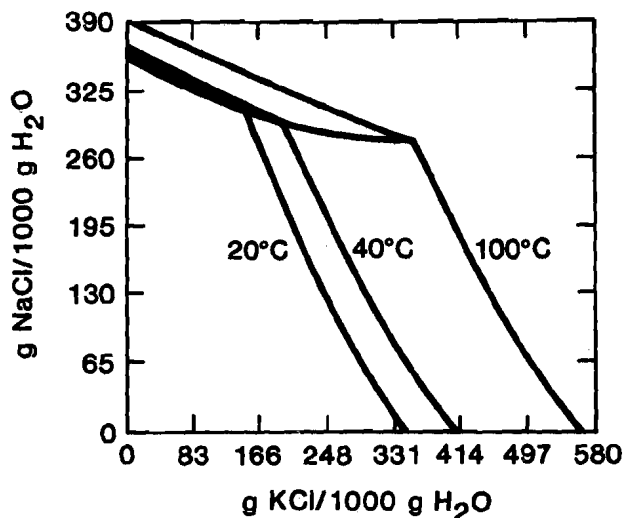


Figure III-1. Solubility Relations in the System NaCl-KCl-H₂O (after Braitsch, 1971). Note the locus of points of maximum solubilities of halite and sylvite as a function of temperature.

The solubility of gypsum or anhydrite is affected by the concentration of NaCl already in solution. Figure III-2 shows the relationship at 25°C for the solubility of calcium sulfates at various salinities.

Salinity (in terms of the thermodynamic activity of H₂O) also influences the hydration state of calcium sulfate (Figure III-3). An increase in hydrostatic pressure tends to expand the stability field of gypsum at the expense of anhydrite; an increase in lithostatic ("dry") pressure tends to expand the stability field of anhydrite at the expense of gypsum (MacDonald, 1953). The question remains as to whether anhydrite is first converted to gypsum before dissolving in a less saline water; this certainly appears to be the case at <800 ft in the Delaware Basin. Gypsum in the Delaware Basin evaporites is largely secondary after anhydrite (Murray, 1964) as a freshwater, near-surface alteration product. The anhydrite may indeed be a primary precipitate (Cody and Hull, 1980), or it may be an *early* diagenetic or even syngenetic-peritectic mineral in the bedded evaporites. In any event, gypsum rarely occurs at depth in an evaporite sequence that is considered "primary."

The dissolution behavior of the most abundant evaporite minerals halite, anhydrite (or gypsum), and polyhalite is important in subsequent discussions, where the dissolution behavior of individual minerals is seen to impose constraints on the composition of

fluids in contact with them. Note that none of the calcium-bearing evaporites can dissolve to yield a solution enriched in calcium with respect to sulfate. Further, a solution enriched in calcium cannot remain so after contact with soluble K- and Mg-sulfates or with sulfate-rich solution. The activities of calcium and sulfate coexisting in solution are inversely related by the solubility product of calcium sulfate. This argument is important in subsequent discussions.

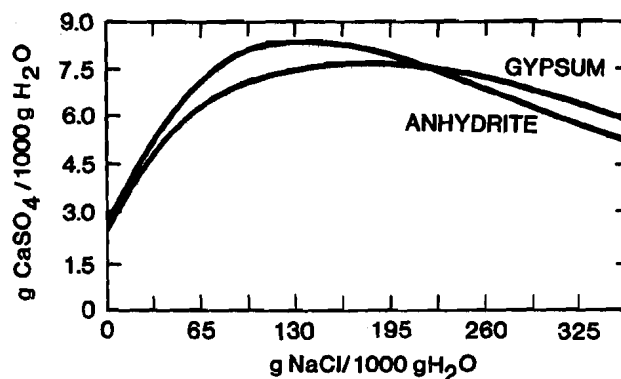


Figure III-2. Solubility of Calcium Sulfates at 25°C as a Function of NaCl Concentration Already in Solution (after Madgin and Swales, 1956).

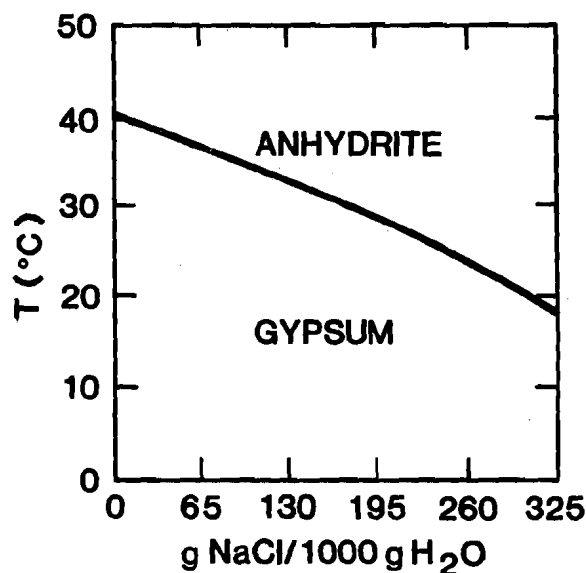


Figure III-3. Stability Relationships of Gypsum and Anhydrite as a Function of Temperature and Salinity (at 1 atm, total pressure). (Calculation after MacDonald, 1953.)

Chapter IV. Stratigraphy of Evaporites and Related Rocks

Introduction

Powers et al (1978) have thoroughly reviewed the stratigraphic relationships in the Delaware Basin. Therefore, only characteristics of the various stratigraphic units relevant to evaporite dissolution are discussed here, with emphasis on pertinence of their characteristics to later discussion.

The rock-stratigraphic units selected for treatment here include formations of the upper part of the

Permian Guadalupian series, the entire Permian Ochoan series, remnants of Triassic and Cretaceous rocks, and Cenozoic deposits related to geomorphic processes. A diagrammatic stratigraphic section near the Basin margin is given in Figure IV-1. A correlation diagram showing the time-relationships of deposition is given in Figure IV-2.

SURFICIAL DEPOSITS (LOCALLY TRIASSIC AND CRETACEOUS ROCKS, OGALLALA ON THE HIGH PLAINS; REGIONALLY GATUÑA, MESCALERO CALICHE AND AEOLIAN SAND)

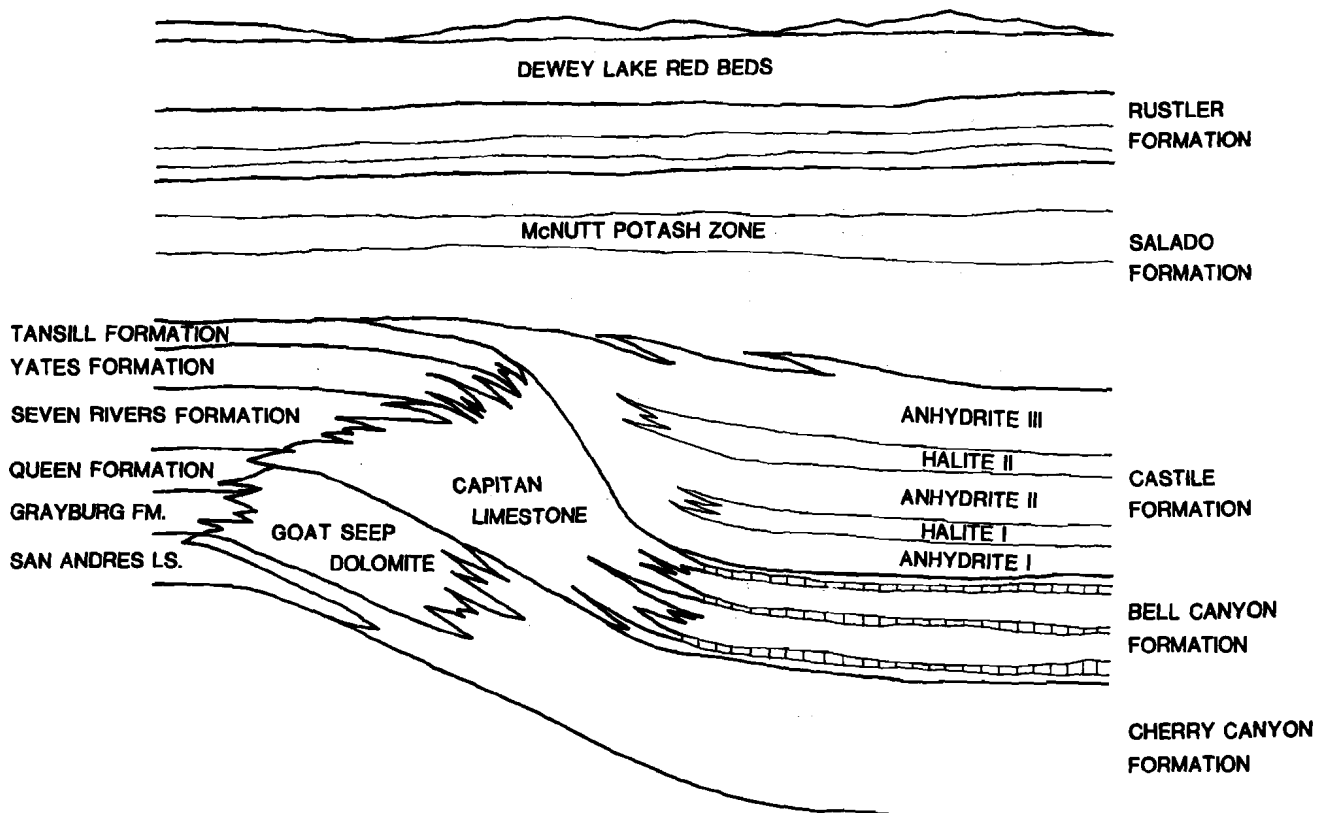


Figure IV-1. Diagrammatic Cross Section Near the Delaware Basin Margin. Shows stratigraphic relationships among some Permian and younger rocks (compiled from several sources)

		SHELF AREA (BACKREEF)	MARGINAL AREA (REEF)	DELAWARE BASIN	
PERMIAN	OCHOAN SERIES	DEWEY LAKE RED BEDS	DEWEY LAKE RED BEDS	DEWEY LAKE RED BEDS	
		RUSTLER FORMATION	RUSTLER FORMATION	RUSTLER FORMATION	
		SALADO FORMATION	SALADO FORMATION	SALADO FORMATION	
		NO CASTILE EQUIVALENT	NO CASTILE EQUIVALENT	CASTILE FORMATION	
	GUADALUPIAN SERIES	ARTESIA GROUP	TANSILL FORMATION	CAPITAN LIMESTONE	BELL CANYON FORMATION
			YATES FORMATION		
			SEVEN RIVERS FORMATION		
			QUEEN FORMATION		
			GRAYBURG FORMATION	GOAT SEEP DOLOMITE	CHERRY CANYON FORMATION
		SAN ANDRES LIMESTONE	SANDSTONE TONGUE BASINWARD EXTENT ?		

Figure IV-2. Correlation Diagram for Some Permian Rocks in the Vicinity of the Delaware Basin (after Hayes, 1964)

Upper Part of the Guadalupian Series

Bell Canyon Formation

The Bell Canyon Formation is the uppermost unit of the Delaware Mountain Group. It is mostly fine-grained sandstone consisting of 0.1- to 0.4-mm-dia grains of quartz, microcline, and plagioclase, with accessory biotite, chlorite, and other heavy minerals (Powers et al, 1978). It is a marine sandstone between 670 and 1040 ft thick (King, 1948), cemented with calcite.

Within the Bell Canyon are prominent limestone beds 10 to 40 ft thick, which have been traced shelfward into the Capitan limestone. The entire Bell Canyon sandstone grades shelfward into the Capitan limestone (qv). The facies change by thickening of limestone beds and pinchout of the intervening sandstone (Hayes, 1964), similar to the relationship between the Goat Seep dolomite and the Cherry Canyon Formation, which underlie the Capitan and Bell Canyon, respectively. Three of the limestone beds are closely spaced in the lower fourth of the Bell Canyon

(Hegler, Pinery, and Rader members). Several hundred feet higher, near the top of the Bell Canyon, is the Lamar member. The McCombs limestone member is about halfway between the Rader and the Lamar.

The top of the Bell Canyon Formation is the base of the Delaware Basin evaporite sequence. Distribution of the Bell Canyon is bounded on all edges of the Basin by the Capitan. Consequently, the Bell Canyon makes no contact with the Artesia group (qv) of the shelf facies, even though the Lamar has been shown as time-equivalent to the middle and lower Tansill Formation (Tyrell et al, 1978).

The Bell Canyon contains permeable sandstone strata (Hiss, 1975). Field and core testing has shown that these superposed saturated zones are ~15 ft thick, and are hydrologically isolated from one another by sandstone of low vertical permeability. The two saturated zones nearest the base of the evaporites are ~500 ft below the Lamar (Mercer and Orr, 1979). These beds represent the closest stratabound source water for dissolution of the overlying evaporites from below. The available quantities and solute content of that water are discussed later.

Capitan Limestone

The unit generally accepted as marking the margin of the Permian tectonic disturbance that created the Delaware Basin is the Capitan limestone, constructed by reef-building organisms whose habitat virtually encircled the Basin. The Capitan consists of local, informal massive and breccia members grading laterally and vertically into each other. The breccia member grades basinward into the Bell Canyon Formation (qv); the massive member grades shelfward into the Tansill, Yates, and Seven Rivers Formations of the Artesia Group (qv). The vaguely defined stratigraphic contact between the two members rises stratigraphically toward the basin (Hayes, 1964). The Capitan contact relationships with adjacent units are illustrated diagrammatically in Figure IV-1. Cavernous porosity is locally well developed in the massive member of the Capitan (Jagnow, 1979), although the contact between massive and breccia members appears to have controlled the development of some caverns in the Guadalupe Mountains.

Because of its cavernous porosity, the Capitan limestone is the most permeable carrier of water of all the rock units associated with the evaporites (Hiss, 1975). The fact that it is abutted laterally by the Castile Formation (qv) and is overlain by the Salado Formation (qv) warrants consideration of its water as a source of fluid that could dissolve the adjacent evaporites.

Artesia Group

Of the five units belonging to the Artesia Group (Grayburg, Queen, Seven Rivers, Yates, and Tansill Formations), the last three are deemed relevant to this discussion because they all have a transitional contact with the Capitan, and because the Ochoan evaporites overlie them.

The Seven Rivers is coeval with the lower Capitan (Lang, 1937). The bedded (dolomitic) carbonate facies of the Seven Rivers undergoes a basinward transition into the Capitan and a shelfward transition into evaporite (gypsum or anhydrite) facies (Sarg, 1981). The carbonate facies is ~5 to 7 mi wide. The Seven Rivers is 460 to 600 ft thick (Hayes, 1964).

The Yates overlies the Seven Rivers with a sharp but conformable contact and is one-third to two-thirds siltstone. The rest is dolomite similar to that of the Seven Rivers. The Yates also grades basinward into the Capitan and shelfward into evaporites. It is 260 to 330 ft thick (Hayes, 1964).

The Tansill Formation overlies the Yates and Yates-coeval rocks in the Capitan massive member. It grades basinward into the uppermost Capitan, and is composed mostly of dolomite near the basin margin, similar to the dolomite in the Yates and Seven Rivers. It also grades shelfward into evaporites, as do the two underlying bedded formations. The Tansill is 90 to 200 ft thick in the subsurface near the northern edge of the Basin (Hayes, 1964).

From transitional contacts between bedded dolomites and massive limestone near the basin margin, the upper three units of the Artesia Group grade basinward into (successively) pisolitic, oolitic, evaporitic, and (finally) clastic facies. After Hiss (1975), the carbonate facies is called the Carlsbad, the evaporitic is called the Chalk Bluff, and the clastic is called the Bernal. In succeeding discussions, the Carlsbad facies of the Artesia Group has the greatest importance because of (1) its lithologic similarity to the Capitan, (2) contact relationships with the Capitan, (3) presumable similarity in water-carrying capacity to the Capitan, and (4) the proximity of the Ochoan evaporites, some of which overlie the Tansill.

Ochoan Series

Castile Formation

There is a transitional albeit abrupt contact between the Bell Canyon Formation and the overlying Castile Formation (Cys, 1978), the oldest of the Permian Ochoan evaporite units. The Castile Formation is bounded laterally by the Capitan, and Figure IV-1 suggests that this boundary may represent sedimentary on-lap, though this remains to be demonstrated. There is little disagreement over the basal and lateral boundaries of the Castile Formation (Adams, 1944; Snider, 1966; Anderson et al, 1972). The nature of the contact between the Castile Formation and the overlying Salado Formation in the Delaware Basin proper (within the confines of the Capitan "reef") is problematical, as pointed out by Anderson (1978).

The Castile Formation is composed of an alternating sequence of anhydrite and halite, laminated to various degrees; some of the anhydrite is locally blocky, nodular, or brecciated. The laminations, well-developed in anhydrite, are defined by thin layers of carbonate, clastics, and organic matter.

In complete sections, the Castile Formation was divided into seven members, all varying somewhat in

thickness throughout the basin. Table IV-1 contains two recently used systems of subdivision of the Castile Formation.

The Castile Formation originally included the overlying Salado Formation (Kroenlein, 1939), but was formally divided by Adams (1944), who alluded to an angular unconformity separating the two in the basin. The rocks Adams called the Castile Formation occur only in the Delaware Basin (Figure IV-2); and the basal contact of the Salado with the underlying Capitan, Tansill, or Yates outside the basin was at the time simple to define: it was the base of the Fletcher anhydrite. Adams (1944) did not, however, accept the time-equivalents of the Fletcher in the basin as the contact. Controversy continues over the nature of the contact; the Adams (1944) and Anderson (1981) schools of thought attach great significance to the unconformity, which Anderson believes represents some pre-Salado dissolution on the top of Castile

Halite III and Anhydrite IV, and some later intraformation dissolution at this same horizon. The thinking of Jones (1954) made the Castile-Salado contact into a transitional intertonguing relationship underneath the Cowden anhydrite ("marker bed 143" of modern usage is equivalent to Adams' "Cowden anhydrite"). The choice of Kroenlein (1939) and Snider (1966) for the contact would be the top of an unspecified anhydrite marker. Jones (1954) adopted Kroenlein's definition of the Fletcher anhydrite, but defined marker bed 143 (not 144) as the Cowden, ~50 ft above Kroenlein's Cowden. Adams (1944) called marker bed 144 the Fletcher, ~50 ft above the Fletcher of Kroenlein and Jones. Later, Jones (1981) applied the name Cowden to what he had formerly called the Fletcher. These various nomenclatures have been applied only to marker beds in the northeastern part of the Delaware Basin, Eddy, and Lea Counties, New Mexico. Table IV-2 is a comparison of the various nomenclatures.

Table IV-1. Lithologic Members of the Castile Formation

Nomenclature		Thickness Variation Throughout Delaware Basin (ft)	Lithology
Snider (1966)	Anderson et al (1972)		
Anhydrite I	Anhydrite I	170 – 300 (SW) (NE)	Anhydrite, divided into laminations by carbonates and organic matter.
Halite I	Halite I	20 – 350 (breccia on west) (NE)	Anhydrite-halite laminated couplets.
Anhydrite II	Anhydrite II	75 – 125 (W) (NE)	Calcite-anhydrite laminated couplets.
Halite II	Halite II	20 – 250 (breccia on west) (NE)	Anhydrite-halite laminated couplets; contains 5 discrete beds, 2 to 5 ft thick, of calcite-laminated anhydrite.
Anhydrite III	Anhydrite III	250(?) – 350 (W) (E)	Calcite-laminated anhydrite, locally indistinguishable from overlying member where basal halite in that member is missing.
Anhydrite IV	Halite III	0 – 300 (anh W) (E)	Halite-anhydrite couplets, with 5 discrete calcite-laminated anhydrite beds 10 to 60 ft thick.
Anhydrite V	Anhydrite IV	0 – 300	Calcite-laminated anhydrite, locally contains halite beds or stratigraphically equivalent breccias, restricted to eastern part of basin. Elsewhere, possibly merged with Anhydrite III where Halite III is missing.

Table IV-2. Comparison of Marker Bed Nomenclature Near the Castile/Salado Contact (Adapted from Snider, 1966)

Present Usage Marker Bed No.	Lithology	Terminology as Used by			
		Kroenlein (1939)	Adams (1944)	Jones (1954)	Jones (1981)
141	Polyhalite				
142	Anhydrite				
143	Anhydrite		Cowden	Cowden	
144	Anhydrite	Cowden	Fletcher		
unnumbered "magnesian" bed	Anhydrite	Fletcher		Fletcher	Cowden

The different stratigraphic terminologies led to confusion in interpreting how much evaporite section was removed by dissolution, particularly in the upper Castile and lower Salado. The only consensus is that the Castile contains mostly anhydrite, and the Salado, halite. A further difficulty arises in the use of marker beds to delineate dissolved zones; the relevant marker beds are not all persistently identifiable throughout the Delaware Basin. They change thickness, pinch out, merge, and split (Jones et al, 1960).

There is no doubt as to the importance of the Castile Formation; it is the oldest (and deepest, where buried) of the Ochoan evaporite units. Thus, it is the evaporite unit nearest to the water in the Bell Canyon Formation, which has been proposed as a source for water to dissolve the evaporites. There is also no doubt as to the importance of the question of how much original evaporite was present and how much has dissolved; the choice of original boundaries of various halite units profoundly affects such quantitative estimates. In succeeding discussions reference will be made to the anhydrite members of the Castile Formation (Table IV-1), as they occur in the relevant portions of the basin and the numbered marker beds of the Salado Formation without regard to the precise stratigraphic position of the Castile/Salado contact. The Cowden anhydrite is here taken as the thick anhydrite bed below marker bed 144 (Table IV-2).

Salado Formation

Before there was a basis for differentiating the Salado Formation as a separate rock-stratigraphic unit, it was called the "upper Castile" or "Main Salt" (Kroenlein, 1939). This was because of abundance of halite, relative to anhydrite, in the subsurface. Near

the reef crest and a short distance backreef (shelfward), the Tansill/Salado division is clearly distinguishable by the interposition of the (Fletcher) anhydrite resting on Tansill dolomite. In the evaporitic facies of the Tansill (qv) there is apparently an uninterrupted sequence of evaporites from Tansill to Fletcher, and the contact is less obvious (Jones, 1954). If the Fletcher anhydrite is adopted not as the base of the Salado, but as a thin shelfward extension of the Castile Formation, the uppermost Castile, not the Salado, is the first unit to entirely overlie the Guadalupian series.

Potassic and magnesian mineralization occurs throughout much of the Salado, most commonly as the mineral polyhalite in the anhydrite marker beds and as nodular and disseminated deposits in halite beds. The occurrence of polyhalite does not follow stratigraphic boundaries, but in the basin the lower limit is generally above marker bed 143, and its upper limit extends into the overlying Rustler Formation (qv). In the backreef, polyhalite occurs in the Tansill evaporitic facies (Jones, 1954).

The zone of soluble potassium mineralization is contained as a subset of the polyhalite zone and is called the McNutt potash zone. The potash deposits are within mixed halite-clastic beds containing accessory clay-rich layers and seams or nodules of polyhalite. The other characteristic minerals are one or more of the following: marker beds of anhydrite, polyhalite, kieserite, or glauberite; sylvite, KCl; carnallite, $\text{KMgCl}_3 \cdot 6\text{H}_2\text{O}$; langbeinite, $\text{K}_2\text{Mg}_2(\text{SO}_4)_3$; kainite, $\text{KMgClSO}_4 \cdot 11/4\text{H}_2\text{O}$; leonite, $\text{K}_2\text{Mg}(\text{SO}_4)_2 \cdot 4\text{H}_2\text{O}$. Of these, sylvite and langbeinite are incorporated into products requiring potassium, such as commercial fertilizer.

Aside from the 45 numbered siliceous or sulfatic stratigraphic marker beds, most of the Salado consists of halite, which is commonly stained shades of orange, pink, or brown because of minor amounts of sulfates, silicates, and oxides. The marker beds are traceable in the subsurface over horizontal distances of several kilometers to tens of kilometers and are cited by number in succeeding discussions. They are sequentially numbered from the top down, from 100 to 144. Not every marker bed is clearly recognizable in every borehole.

Rustler Formation

The Rustler Formation is composed of ~40% anhydrite, 30% halite, 20% siltstone or sandstone, and 10% anhydritic dolomite. The various lithologies are organized into five recognizable members, following the nomenclature of Vine (1963). In ascending order, they are (1) an unnamed red-brown siltstone interbedded with anhydrite, 120 ft thick; (2) thin-bedded vuggy gypsiferous or anhydritic dolomite or dolomitic gypsum/anhydrite, 25 to 30 ft thick, named the Culebra member; (3) 190 ft of anhydrite, interbedded with halite and traces of polyhalite, the Tamarisk member; (4) 20 to 30 ft of gypsiferous or anhydritic dolomite or dolomitic gypsum/anhydrite, composed of thin undulatory laminae containing silty matter, the Magenta member; and (5) 80 ft of anhydrite, siltstone, and halite, the Forty-niner member.

Anhydrite in the Rustler Formation is locally altered to gypsum where (1) water moving through the Culebra and Magenta members is fresh enough to facilitate the alteration of adjacent beds, (2) the anhydritic members of the Rustler Formation are exposed at the surface, (3) gypsum karst is developed in the shallow subsurface. In places where gypsification has occurred, halite has certainly been removed, but removal of halite does not necessarily result in gypsification of adjacent anhydrite.

The Forty-niner member is locally deeply weathered to a red-brown siltstone mottled with green-gray reduction spots. Where gypsum is absent, it is indistinguishable from the overlying Dewey Lake Red Beds (Bachman, 1980).

The Rustler Formation, like the Castile Formation, has widespread surface exposure, resulting in the

development of gypsum karst. It differs from the other Ochoan rocks in its ability to carry significant amounts of water in the two dolomitic beds. These waters are isolated from one another to various degrees, and in many occurrences of the Rustler are totally confined within their respective host beds. Development of certain geomorphic and structural features in the Rustler has been used to describe certain of the dissolution mechanisms previously proposed.

Dewey Lake Red Beds

The last of the Ochoan rocks has virtually no affinity with evaporites. The Dewey Lake Red Beds are fine-grained, red-orange, micaceous quartz sandstone, commonly veined with epigenetic selenite. They are conformable on the Rustler Formation and are truncated by an angular unconformity signifying the end of the Permian section in southeastern New Mexico.

Triassic Rocks

Erosional remnants of the Triassic Dockum Group are exposed in the northern part of the Delaware Basin (Bachman, 1980). These poorly sorted deposits of conglomeratic sandstone and (locally) shale overlie beveled Permian evaporites within the western edge of the Delaware Basin. This relationship shows that the evaporites were exposed at the surface and therefore were subject to erosion and dissolution as early as Triassic time.

Cretaceous Rocks

Early Cretaceous fossiliferous marine rocks (Lang, 1947) are preserved as debris in collapse features in the Pecos River drainage system. At two localities, this debris rests on gypsum of the Castile Formation. At a third, Cretaceous rocks are mixed with fragments of Triassic conglomerate and Culebra member of the Rustler Formation (Bachman, 1980). The interpretation of these geological relationships is not without ambiguity.

Cenozoic Rocks

Ogallala Formation

Well-sorted, windblown sand surmounted by a well-developed pedogenic caliche caprock characterizes the Miocene and Pliocene Ogallala Formation in southeastern New Mexico. It was deposited on an irregular erosional surface, and its upper surface forms the featureless High Plains caprock, or Llano Estacado (Figure I-1), sloping southeastward 2.1 m/km. This is interpreted as original dip (Bachman, 1976). This unit occurs near The Divide and east of San Simon Swale, and structure contours on its surface are taken to represent presence or absence of post-Pliocene deformation (e.g., subsurface dissolution and collapse) in those areas where it is preserved.

Gatuña Formation

Deposits of various kinds that accumulated in erosional depressions in the Pecos Valley in post-High

Plains (Ogallala) time are called the Gatuña Formation. These deposits consist of lenticular gravels reworked from Triassic and Ogallala rocks, pale red-brown sand and sandy clay, and light-yellow, well-sorted sand. In the upper part of the Gatuña Formation a volcanic ash bed occurs, identified as Pearlette type "O," derived from the Yellowstone region 600 000 yr ago (Bachman, 1980).

Flat-lying occurrences of the Gatuña Formation in depressions are believed attributable to dissolution and collapse, making it an important unit for placing limits on age of collapse.

Mescalero Caliche

The formation of pedogenic caliche requires a finite length of time on a stable land surface (Bachman, 1980). A caliche-covered surface, such as the Mescalero, can also be used to establish limits of time of formation of certain surficial features. The Mescalero caliche, occurring on a broad surface between the Pecos River and the High Plains, is 410 000 to 510 000 yr old, according to the uranium series disequilibrium method (Szabo et al, 1980).

Chapter V. Water Available to Dissolve Evaporites

Introduction

As with the previous discussion on stratigraphy, a complete review of hydrostratigraphic relationships and flow parameters is not presented here in favor of features relevant to the evaporites and adjacent rocks. In particular, limitations on quantities and sources of water available to dissolve the evaporites are described, especially in the light of previous estimates that are shown to be partly in error.

For this report, the relevant Guadalupian section is divided into the Yates-Tansill, Bell Canyon, and Capitan hydrostratigraphic units. The Yates and Tansill Formations are considered together because of lithologic similarities (silty bedded dolomite), and because they underlie the Ochoan evaporites in the backreef (or "shelf") area and contain the regionally persistent occurrence of water nearest the Salado. Locally the Yates may be in contact with the Salado where the Tansill is thinned. The other shelf units are considered only as they relate to the Capitan. The Capitan warrants separate status as a hydrostratigraphic unit because of its regional importance as an aquifer, and because of its proximity to evaporites of the Castile and, locally, the lower Salado. The Bell Canyon is the only basinal Guadalupian unit of importance here. It immediately underlies the Castile evaporites and is the only basinal Guadalupian unit in contact with the Ochoan section. There may or may not be a hydrologic connection between the Bell Canyon and Capitan Formations (Hiss, 1975), but contrary to the findings of Hiss, it will be argued that the entire Delaware Mountain Group is probably not a single, vertically interconnected hydrostratigraphic unit. Thus the amount of water contained in the Bell Canyon Formation (and therefore available to dissolve evaporites) is limited and is significantly smaller than that postulated by Anderson (1978).

The only Ochoan unit considered here to be a major hydrostratigraphic unit is the Rustler Formation. The Dewey Lake Red Beds contain a "moist" zone (Mercer and Orr, 1979), but this water is not laterally persistent. Similarly, Triassic and younger

rocks are not treated as a regionally pervasive source of water in the Delaware Basin, because they do not occur throughout the basin.

Bell Canyon Formation

The Bell Canyon Formation is a unit that may locally supply various amounts of hydrocarbons (Powers et al, 1978), but not domestic or industrial water because of the limited quantity and poor quality of the water. Consequently, nearly all the information available for interpreting Bell Canyon hydrologic conditions was originally obtained by hydrocarbon production companies (Hiss, 1975). Near the proposed WIPP site, information was obtained from three holes by the United States Geological Survey (USGS) in cooperation with Sandia National Laboratories (SNL) on behalf of the US Department of Energy (DOE).

The 1° eastward regional dip (Snider, 1966) of the Delaware Basin has resulted in erosional exposure of the Bell Canyon Formation and its intertonguing relationship with the Capitan limestone in the western part of the basin. Here "large springs near the base of the reef escarpment probably are supplied by groundwater moving through the upper beds of the Bell Canyon Formation" (Hendrickson and Jones, 1952). In boreholes the uppermost limestone member (Lamar) has yielded brine and, locally, petroleum (Mercer and Orr, 1979). During the renewed drilling of AEC 8 into the Bell Canyon Formation, geophysical logging revealed the vertical distribution of porosity in the sandstone units. Mercer and Orr (1979) did not comment on the Lamar member and the Ramsey sand (a locally important hydrocarbon "pay" zone) because of their low porosity. Two other sandstone beds (4832.5 to 4848.5, and 4809.5 to 4815.5 ft below Kelly bushing in AEC 8, known as the lower sand and upper sand, respectively) were the only potential water-yielding units encountered in the upper 700 ft of the Bell Canyon Formation. This 700-ft interval accounts

for two-thirds of the entire formation. Shortly after dual completion of the hole in these two zones, static levels of water derived from the lower and upper sands were 615 and 560 ft, respectively, below land surface (Mercer and Orr, 1979). This conspicuous difference in levels of water of similar density attests to the stratabound, vertically isolated nature of the water in the Bell Canyon Formation. In addition, the deposition-controlled porosity containing natural gas in isolated lens-shaped sandstone reservoirs also indicates only small degrees of vertical and horizontal connected porosity in the Bell Canyon Formation. Thus, in the upper 700 ft of the Bell Canyon Formation, the total saturated thickness is <30 ft.

Hiss (1975) compiled several laboratory determinations of permeability and porosity made by oil companies on selected cored intervals, mostly from the lower Bell Canyon and upper Cherry Canyon Formations. About 4900 ft of core was measured (4500 samples), mostly in the horizons of the most promising geologic section for hydrocarbon production. In addition, about half the data points are from wells in Ward County, Texas, with about 2000 from Eddy and Lea Counties in New Mexico. No aquifer performance tests were reported for any of the Delaware Mountain Group rocks. Hiss reports an "average" permeability for the "Delaware Mountain Group" samples in the four-county area (Eddy, Lea, Winkler, Ward) of 6.70 mD (~0.016 ft/d or 0.005 m/d expressed as hydraulic conductivity). The "average" porosity was 15.65%. These values must be considered local maxima, and not representative of the entire Guadalupian sequence or the Bell Canyon Formation, because the core samples were initially selected for their promisingly high permeability with respect to production of hydrocarbon.

The extremely limited quantities of water produced from but few thin horizons of the Bell Canyon Formation (Mercer and Orr, 1979) hardly qualify this unit as a "major aquifer" as proposed by Anderson (1978, p 31). In addition, the following facts all call into question the importance of the Bell Canyon waters as a dissolution agent:

- Much of the Bell Canyon water is highly saline, but not completely saturated with sodium chloride under the evaporites (Hiss, 1975).
- The salinity does not abruptly rise from west to east as evaporites appear in the overlying section (Hiss, 1975) as would be expected at a dissolution "front."

- The water contains solutes in combinations not found in the evaporites (Lambert, 1978).
- There is little evidence for appreciable movement of water in the Bell Canyon (Mercer and Orr, 1979).

These lines of evidence are examined in more detail in a subsequent chapter.

Capitan Limestone

Although the Capitan limestone occurs only at the margin of the Delaware Basin in a band 10 to 14 mi wide and 1600 ft thick (Hiss, 1975), its water-carrying capacity has implicated it as a major agent in evaporite dissolution. Due to the eastward regional dip, the Capitan east of the Pecos River is restricted to the subsurface (Figure V-1). It is exposed prominently in the Guadalupe Mountains to the west. The Capitan limestone is renowned for its locally well-developed cavernous porosity, particularly in its massive member. Many caverns have been formed by phreatic solution of calcium carbonate along joints in the massive Capitan, perhaps by the agency of sulfuric acid derived from the oxidation of pyrite in the nearby shelf facies. Such a reaction sequence would account for significantly large occurrences of gypsum in many of the Guadalupe Mountains caves (Jagnow, 1979). Cavern development is not thought to be active at the present time; most of the Guadalupe caves are considered "dry," with the local water table well below the explored cavernous porosity. Water in the Guadalupe caves is restricted to drip-catchment pools; no flowing streams or lakes can be shown to be fed by springs.

In the northern and eastern margin areas of the Delaware Basin, high-porosity zones have been discovered by interpretation of "breaks" in drilling through the Capitan. Five such occurrences were reported (Hiss, 1975), varying in thickness between 12 and 60 ft. None were as large as the subsurface relief exposed in Carlsbad Caverns, 100 to 300 ft.

Quantitative hydrologic measurements on the Capitan are few. Most of the data, derived from drill-stem tests, are only estimations. Eight well performance tests were reported by Hiss (1975). Figure V-1 shows the geographic distribution of hydraulic conductivity values derived from well tests in the Capitan. The values vary from 1.4 to 5.2 ft/d, with one locally high value of 25 ft/d on the Central Basin Platform margin, just 1/2 mi from a point with a value

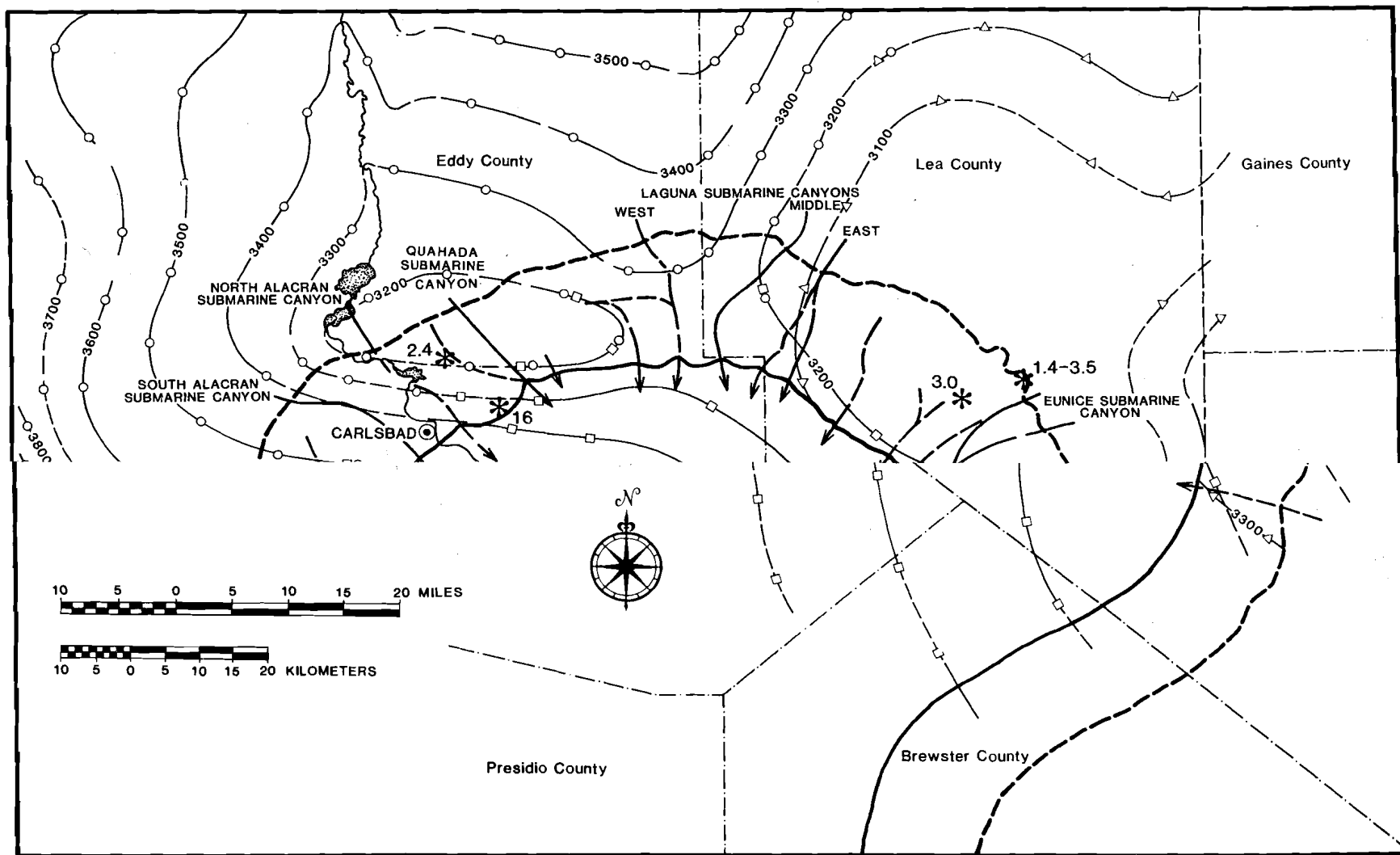


Figure V-1. Distribution of the Capitan Limestone and Potentiometric Water Levels in Guadalupian Rocks of the Delaware Basin (compiled from Hiss, 1975). Heavy lines are outcrop or subcrop limits projected to surface, approximately indicating boundaries of the Capitan limestone. Contours are fresh-water corrected, estimated to be in effect *before* major extraction of hydrocarbons by secondary stimulation, extraction of water for which resulted in significant later drawdown in Capitan contours. "Basin aquifer" contours are composites for the Bell Canyon, Cherry Canyon, and Brushy Canyon Formations. "Capitan aquifer" contours are composites for the Capitan limestone and Goat Seep dolomite. "Shelf aquifer" contours are composites for the Tansill, Yates, Seven Rivers, Queen, Grayburg, San Andres, and "Glorieta" Formations. Locations of Capitan well tests are shown, and trends of "submarine canyons" (inferred from local thinning in the Capitan) are depicted.

of 4.4 ft/d. Similarly, near the city of Carlsbad, points with values of 2.4 and 16 ft are <1 mi apart. Since there are no saturated zones to test in the western (Guadalupe Mountains) portion of the Capitan, no quantitative hydrologic data are available there.

Yates-Tansill Hydrostratigraphic Unit

Regional hydrologic data for the Yates, Tansill, and underlying units have been derived from the same sources as for the Bell Canyon; i.e., from drill-stem tests made by hydrocarbon production companies, and from laboratory core tests. Because of its higher potential for hydrocarbon production, more samples of the Yates Formation have been analyzed. In fact, for most of the four-county area reported by Hiss (1975) the number of Yates core analyses (11 000) accounts for ~32% of the total number for the entire shelf aquifer system. About 400 (1%) of the analyses are for the Tansill. The "average" permeabilities of the Yates and Tansill in the four-county area are 10.79 and 2.51 mD, respectively (0.026 and 0.006 ft/d or 0.008 and 0.002 m/d expressed as hydraulic conductivity). The respective porosities are 9.74% and 4.23%.

The average porosities of the remaining shelf units (Seven Rivers, Queen, Grayburg, San Andres, and "Glorieta") lie between those values. Their permeabilities are from 11 to 14 mD (0.025 to 0.035 ft/d or 0.008 to 0.012 m/d), except for the Seven Rivers, with 56 mD (0.14 ft/d or 0.043 m/d). Thus, the hydraulic conductivities of the shelf units are all one-tenth to one-hundredth as large as that of the Capitan. Hydrologic communication across the shelf/reef facies margin is restricted by the high contrast in permeability, and the lowest (a factor of 10) contrast is between the Seven Rivers and Capitan Formations.

Regional Hydrology of Guadalupean Rocks

Hiss' work of 1975 emphasized the change in groundwater flow patterns in historic time. The flow patterns were based on potentiometric surfaces characteristic of (a) the early 1920s and (b) the period 1960 to 1970. The main reason for the differences was the

voluminous recent withdrawal of Capitan water near Kermit, Texas, for use in stimulating secondary production of hydrocarbons from old wells by injection of water. The two sets of potentiometric contours (called "predevelopment" and "postdevelopment" for the early and later time periods, respectively) reveal the relative responses of the Capitan and adjacent groundwater reservoirs to pumping-induced stresses. Patterns of groundwater movement are based on trends in potentiometric contours. Similarly, the measured hydraulic conductivities are based largely on amounts of water supplied to wells in single-well tests. Aside from circumstantial responses of the Capitan to pumping around Kermit, there is no direct indication (such as multiwell tracer test results) of any movement of water in the Guadalupian section.

Several subsurface features have been described by Hiss as possible impediments to the movement of groundwater in the Capitan and adjacent aquifers: (1) igneous dikes, cutting the Capitan, that were dismissed as inconsequential, (2) the Laguna Submarine Canyon System through the Capitan (Figure V-1), which locally decreases the thickness and therefore the transmissivity through the Capitan near the Eddy-Lea county line, (3) the 100-fold difference in hydraulic conductivities that limits water exchange between Capitan and adjacent rocks.

In any consideration of the history of the dissolution of Delaware Basin evaporites through geologic time, the predevelopment potentiometric surfaces are the more relevant. Most of the potentiometric surfaces are results of individual water levels in wells open to the formation of interest. Hiss (1975) observed that obtaining conditions for the predevelopment period was difficult, and in many cases they were estimated from the regional history of fluid production, or were extrapolated back in time on the basis of observed rates of changes in potentiometric gradient. In a description of potentiometric conditions of the Capitan, Hiss chose to subdivide the Capitan as (see Figure V-1): (1) Guadalupe Mountains northeastward to Carlsbad on the Pecos River, (2) Carlsbad eastward to the Eddy-Lea county line, (3) the county line eastward and southward to Kermit, Texas, (4) Kermit southward to the Glass Mountains.

Water was said to enter the Guadalupe Mountains, move northeastward under water table conditions, and discharge near Carlsbad as springs on the Pecos River. The role of the Guadalupe Mountains area as the principal recharge area for this part of the Capitan was questioned by the stable-isotope work of

Lambert (1978). Nevertheless, the potentiometric gradient in the area was 1 to 2 ft/mi toward Carlsbad. The head distribution in the Capitan west of the county line appears to be controlled by good hydraulic communication with the Pecos River. The Capitan potentiometric data (Hiss, 1975) indicate that gradients in the Capitan were very small, except near the Glass Mountains, which may constitute a recharge area in the southeast part of the Delaware Basin. Another notable exception may be the county line area, where the locally diminished transmissivity may give rise to a locally higher gradient. Predevelopment potentiometric surfaces average 3140 to 3150 ft above sea level (corrected to fresh water) in the 20-mi stretch from the Guadalupes to Carlsbad, 3200 to 3300 ft at the groundwater divide near the county line, 3000 to 3100 ft between Hobbs and Jal, 3100 ft near Kermit, and 3300 ft in the Glass Mountains. Because the upper boundary of the Capitan is below its potentiometric level east of the Pecos River, it is under confined conditions in that area, whereas water-table conditions prevail west of the river.

The Capitan can be either a source or sink for water in adjacent shelf and basin units. At all locations along the Basin margin, the Bell Canyon has a higher head than does the juxtaposed Capitan, even after corrections are made for salinities; between the basin units and the Capitan, Hiss shows a potential difference of 100 ft at Carlsbad (and along the Central Basin Platform) to 800 ft at White's City. Thus there is no tendency for even fresher Capitan water to flow into the Bell Canyon Formation.

Osmotic pressure differential provides additional information about the separateness of the waters in the Capitan and Bell Canyon. If we assume a typical reef-margin Capitan value of 3000 mg/L total dissolved solids, and an NaCl-saturated value of 300 000 mg/L for the Bell Canyon, and that the Capitan/Bell Canyon contact acts as a semipermeable membrane, the calculated osmotic pressure differential (Lewis and Randall, 1961) at 30°C is ~1900 psi, or a fresh-water equivalent head of ~4400 ft driving from the Capitan into the Bell Canyon. Thus, through geologic time there has been little tendency, through solution mixing, to diminish this gradient and approach osmotic equilibrium. Further, it is not apparent that the fluid production from the Capitan in the last 30 yr has affected the head distribution in the Bell Canyon, according to the data of Hiss (1975). Thus, lateral flow

from the Bell Canyon into the Capitan, even where the differential head is highest near the Central Basin Platform, cannot be demonstrated. In view of (1) the high permeability contrast between Capitan and Bell Canyon, (2) the preservation of a higher Bell Canyon potentiometric head, and (3) the preservation of a higher Capitan osmotic potential, significant amounts of water flow in *either* direction is unlikely.

Hiss (1975) inferred a lateral connection between the Capitan and the Yates-Tansill hydrostratigraphic unit. The strongest evidence for this connection is the head differential pattern in the eastern and northeastern reef margin (Figure V-1), where the heads of both Capitan and undivided shelf units are all ~3100 ft. Most of the "shelf" data points are taken from units older than the Seven Rivers (Yeso, Glorieta, Queen, Grayburg, San Andres, Bone Spring/Victorio Peak), which have no apparent connection with the Capitan. On the northwestern shelf between White's City and Artesia, however, the shelf/Capitan head differential is between 50 and 800 ft, increasing southwestward. Because most of the data points in this region come from the San Andres, little can be said about the degree of connection between the Yates, the Tansill, and the Capitan in this region. Nevertheless, great confidence in direct connection between "basin" and "shelf," bypassing the Capitan as inferred by Hiss, is here deemed unwarranted.

The "groundwater divide" at the northern apex of the reef appears to be manifest in the shelf system as well as in the Capitan. West of this divide, heads were not perceptibly influenced by recent hydrocarbon-field development. Near the Central Basin Platform the development was reflected in significant draw-down in both the Capitan and shelf units, to similar degrees, illustrating the good connection in this area.

Predevelopment potentiometric patterns suggest that the northeastern shelf area provided an outlet for water in the Capitan through the so-called "Hobbs Channel" (Figure I-1). Given this relationship on the east and the high permeability contrast on the west, it is difficult to imagine a significant amount of recharge for the Capitan from the shelf. The source of Capitan recharge is problematical.

Difficulties in the Hiss model of Capitan recharge in (at least) the Guadalupe Mountains were discussed by Lambert (1978). The relatively homogeneous groundwater reservoir in the Capitan is isotopically distinct from the seasonally integrated sampling of

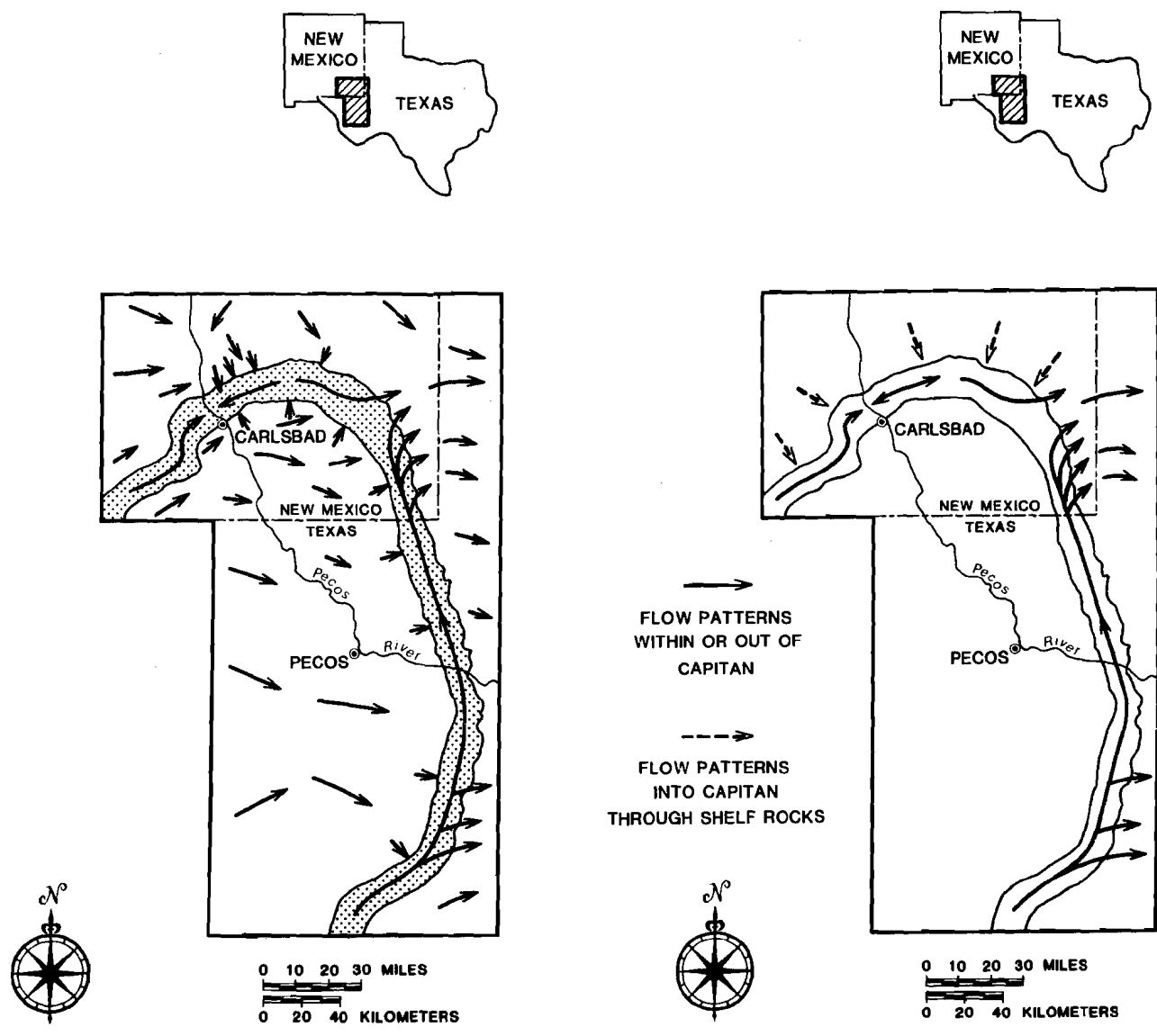
groundwaters in the vadose zone in the Guadalupe Mountains caverns. This indicates that such recharge cannot be a major contribution to the Capitan. Isotopic relationships are treated in detail in the section on interaction between rock and water.

The nondifferentiation of hydrologic descriptions of various individual stratigraphic units has made difficult the assembly of a self-consistent model for the movement of water in the Delaware Basin. Little significance can be inferred from treating the shelf units as one "aquifer," the basin units as another, and then postulating a connection between the two that bypasses the Capitan. One such difficulty is illustrated by the San Andres Limestone (Figure IV-1). The San Andres Limestone has been variously treated by Hiss (1975) as (1) an individual unit in which there were two single-well pump tests near the "Hobbs Channel," (2) part of an undivided Grayburg-San Andres unit, (3) part of the undivided shelf sedimentary rocks consisting of Tansill, Yates, Seven Rivers, Queen, Grayburg, and San Andres. The porosity and permeability data have been so grouped. The San Andres, however, has several unique features that justify its individual consideration as an aquifer.

First, it is the only shelf unit coming in direct contact with a basin unit, specifically the shelfward sandstone tongue of the Cherry Canyon Formation (Figure V-1). Second, it has a close affinity with the Goat Seep Reef partially underlying the Capitan (though hydrologic communication between the two reefs has not been documented). Third, porosity and permeability for the San Andres cores are typically higher (10% and 0.2 ft/d, respectively) than for most other shelf units. If there is to be an isotopically homogeneous body of water in the Capitan, the probable recharge sources are (1) the Pecos River where it cuts the Capitan near Carlsbad, (2) some *small*, undetermined recharge through outcrops in the Guadalupe

Mountains, (3) some undetermined amount of recharge through outcrops in the Glass Mountains, and (4) recharge from shelf units, particularly the San Andres Limestone by means of the Goat Seep Reef. Water from the Pecos River would mainly supply the portion of the Capitan between the river and the groundwater divide. Any recharge in the Guadalupe Mountains area would discharge at the River. Water from the Glass Mountains would have exited the Capitan through the Hobbs Channel. Thus, the major source of recharge in the northern portion of the reef must (by default) be the shelf rocks, specifically the San Andres. This hypothesis has yet to be tested.

Figure V-2A is a summary of the Hiss model of water flow in the Delaware Basin. Figure V-2B is a modified proposal, based on consistency with current data. The differences between the revised model and that of Hiss are (1) virtually no involvement of the Bell Canyon Formation in either recharge to or discharge from the Capitan, (2) a subdivision of the "shelf" system so that various individual shelf units can act as either recharge to or discharge from the Capitan, (3) essentially no connected flow within the Bell Canyon Formation. Because the Bell Canyon Formation has been invoked as a major source of water for dissolution of Ochoan evaporites by Anderson (1978), this last point is especially significant to this discussion. It was shown that the supply of Bell Canyon water is limited in that (1) sustained flow is limited by the low permeability, (2) reservoir size is limited by the irregular, local, unconnected nature of fluid reservoirs in the Bell Canyon (water, oil, natural gas), and (3) the Bell Canyon Formation is not actively and continuously recharged by any known meteoric or groundwater source in the Delaware Basin (Lambert, 1978). These points are discussed in detail in Chapters VIII and X with the evaluation of various dissolution models.



A. Hiss (1975)

B. This work, incorporating recent geochemical data

Figure V-2. Comparison of Models Proposed for Groundwater Movement in Guadalupian Rocks of the Delaware Basin. Arrows indicate relative magnitudes and directions of groundwater flow. Note no involvement of the "basin aquifer" and a greater involvement of the "shelf aquifer" in B.

Rustler Formation

As indicated previously, the Rustler Formation contains two members of dolomitic anhydrite (locally altered to gypsum) that carry varying amounts of water. Quantitative hydrologic data for the Rustler Formation were developed slowly over the past 30 yr. Because the dolomitic beds locally yield relatively small amounts of water of marginal quality, the Rustler Formation was not historically considered a significant aquifer. Its chief use has been to supply local stock tanks where yields of at least 15 gal/min with total dissolved solids <10 000 mg/L can be produced from a well by a wind-driven pump. Typically, the quantity and quality of Rustler waters are much lower. In Nash Draw, an area of active shallow-seated evaporite dissolution, gypsum of the Tamarisk and Forty-niner members has developed a cavernous porosity. Increased permeability afforded by this porosity and fracturing of the lower, more intrinsically permeable (Culebra) dolomitic member delivers 300 to 700 gal/min to some wells in the western part of Nash Draw (Hendrickson and Jones, 1952).

Interest in the Rustler Formation was renewed with the advent of Project Gnome, a nonweapons ("Plowshare")-related nuclear detonation in the Salado Formation in December 1960. Again it was observed that yields of water varied "considerably from place to place," yield depended on degree of solution alteration and fracturing of the Culebra dolomite, and solution alteration and fracture density decreased with increasing overburden (Cooper and Glanzman, 1971). Quantitative measurements of hydraulic parameters of the Culebra near the Gnome site were until recently taken to represent conditions east of the Pecos River: effective porosity 10%, transmissivity 500 ft²/d, hydraulic conductivity 16 ft/d, storage coefficient 2×10^{-5} , velocity ~ 0.5 ft/d (Cooper and Glanzman, 1971; Mercer and Orr, 1977). Water was found to be under confined ("artesian") conditions, with a static level of ~ 75 ft above the top of the Culebra near the Gnome shaft.

Because the Rustler Formation has such a low intrinsic permeability, quantitative hydrologic testing by conventional methods is difficult. The low permeability of the Rustler has required the modification of usual procedures. Details are specified by Mercer and Orr (1979) and Mercer and Gonzalez (1981). Initially 24-h drill-stem tests were performed in open holes.

Low rates of fluid pressure and level buildup, and unstable hole conditions made such tests of limited value. More easily interpretable results were obtained from long-term observation of wells completed in one producing zone. The zone of interest is open to the hole or is accessed by perforation through casing. If multiple completion is desired, casing affords a more reliable seat for a packer or bridge plug. Thus, wells can be pumped, bailed, or swabbed for extended periods of time, appropriate for low rates of fluid production. Such specially completed wells are available for long-term monitoring of fluid level changes, such as slow recovery after a drawdown. Several such wells have been developed in the Los Medaños and Nash Draw area.

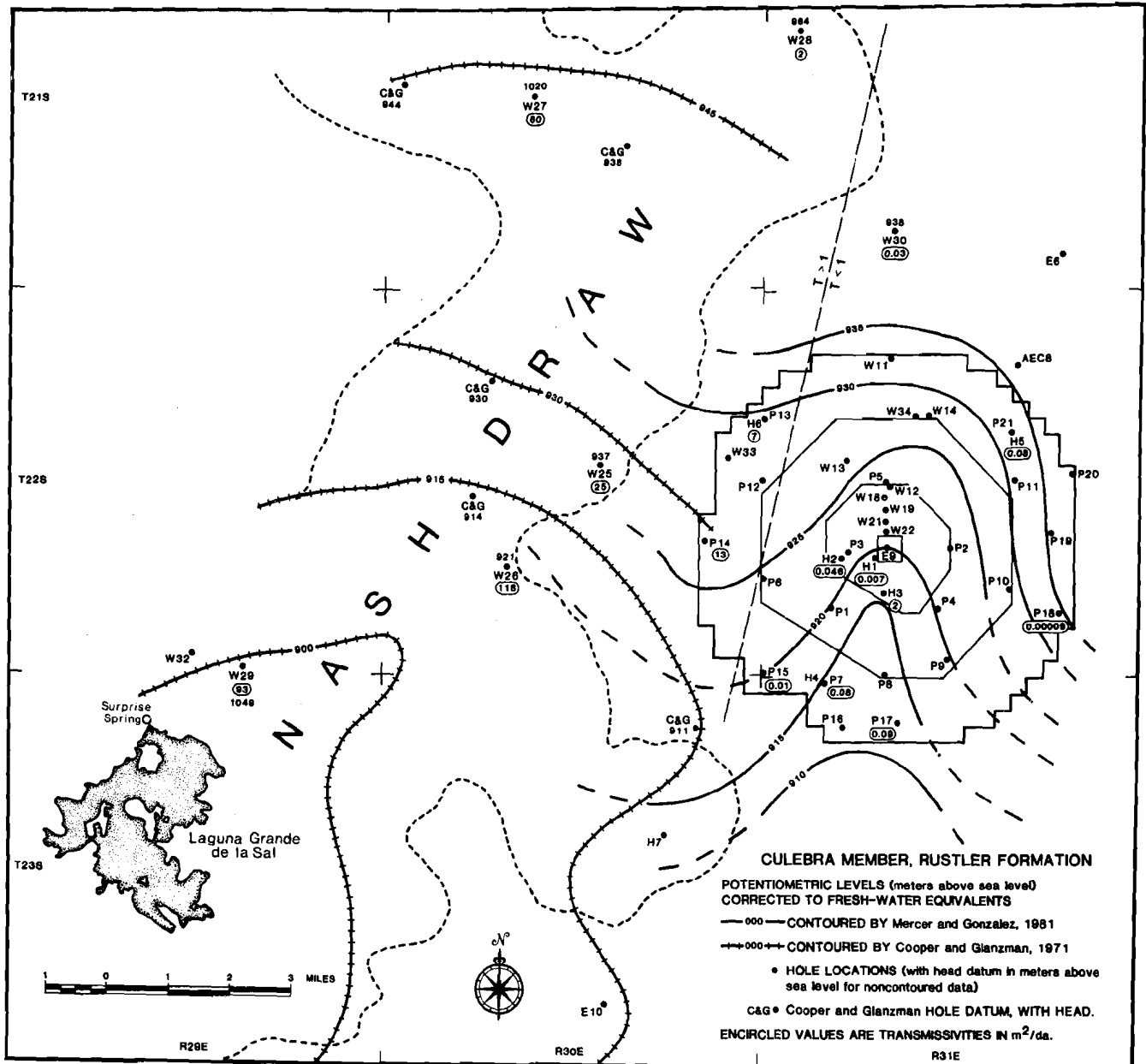
Figure V-3A shows the potentiometric contours for the Culebra member of the Rustler Formation near Los Medaños and Nash Draw (Mercer and Gonzalez, 1981). Contours are corrected to unit fluid density. In addition, contours are shown as drawn by Cooper and Glanzman (1971). The two sets of contours cannot be reconciled at this time, since it is not known if the 1971 interpretation (specifically for Nash Draw) contains a density correction. In addition, point-values of transmissivity are shown on the map (Mercer and Gonzalez, 1981). Note especially the highly variable values throughout the area. Note also the suggestion of a dividing line at $T=1$ m²/d, possibly reflecting the more highly fractured (by collapse?), more permeable nature of the Culebra to the west toward Nash Draw. Where the permeability is indicated to be lower, the potentiometric contours are more closely spaced (to the east). Conversely, to the west (toward Nash Draw), permeabilities are higher and contours are widely spaced. The relationships between the hydrology of Nash Draw and its origin as a dissolution feature are discussed elsewhere.

Measurements of water levels in privately owned stock wells may be less reliable in contributing to regional potentiometric maps of the Rustler waters. Cooper and Glanzman (1971) pointed out that many of the stock wells completed in the Rustler may tap the Culebra, Magenta, Tamarisk, Forty-niner, lower member, or any combination of members. Owing to locally high permeabilities that make a certain point in the Rustler productive and useful, that point may be tapped and the well opened to whatever zones seem appropriate to the owner. The most regionally productive Rustler member is the Culebra, and the Culebra is

most likely to be the major contributor to a well. Little reliance can be placed in a single-well measurement unless the producing zone can be positively identified.

There is virtually no natural vertical hydraulic connection between the Magenta and Culebra members, with the possible exceptions of locations at WIPP 25 and 27 (Gonzalez, in preparation). Else-

where, each unit supports a water column in a well differing in height from the other by several meters. Mercer and Orr (1977, 1979) pointed out that potentiometric levels of respective Rustler "aquifers" decrease in elevation with increasing depth of the water-bearing layers, at any given point (e.g., a single borehole).

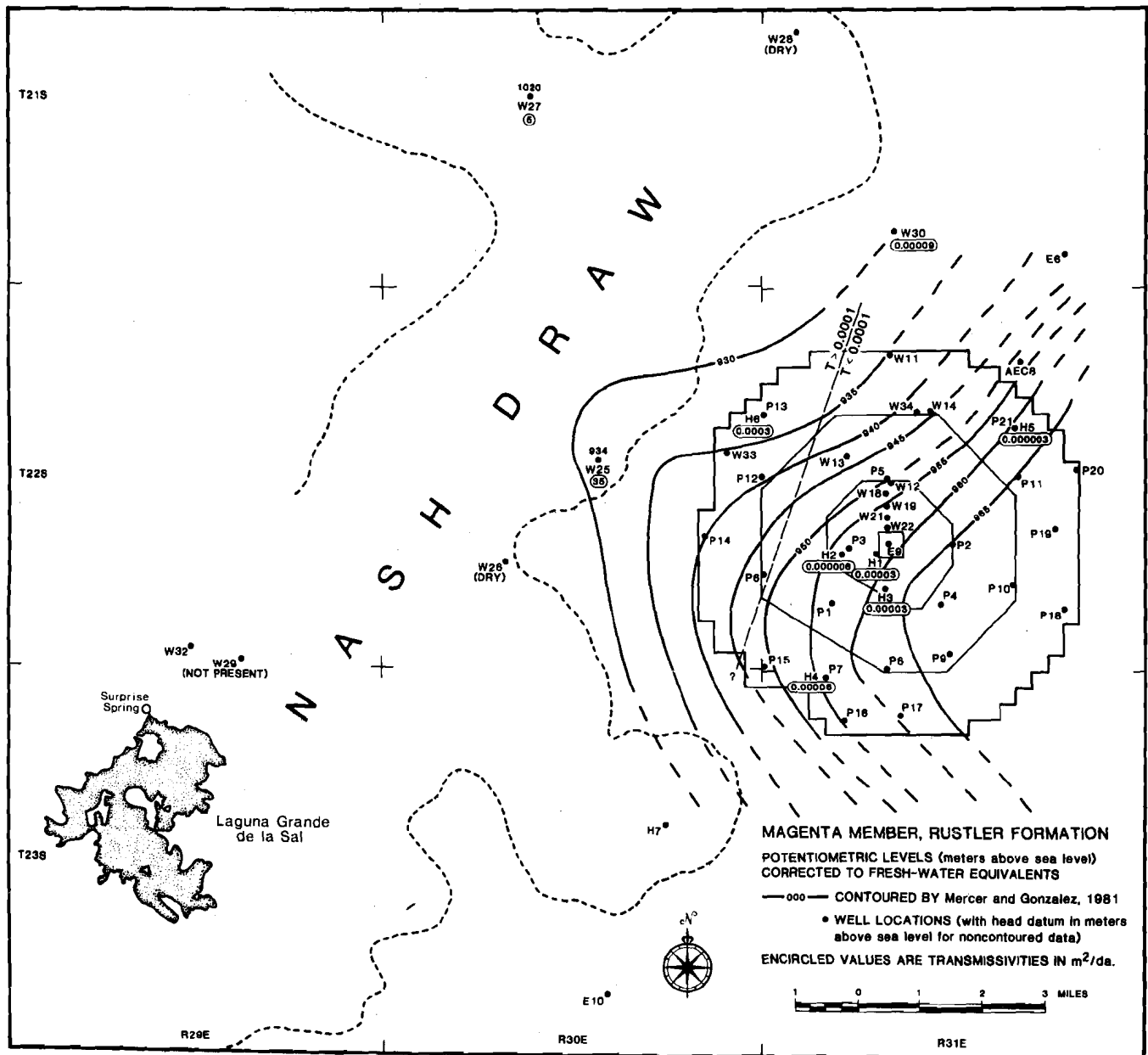


A. Culebra member

Figure V-3. Hydrologic Features of the Rustler Formation in the Vicinity of Los Medaños and Nash Draw. Solid contours are from Mercer and Gonzalez (1981). Hachured contours are from Cooper and Glanzman (1971). Noncontoured data points are treated by Mercer (in preparation). Irregular and octagonal boundaries are proposed WIPP land-use control zones (Powers et al, 1978).

Figure V-3B shows the potentiometric contours for the Magenta member of the Rustler Formation near Los Medaños and Nash Draw (Mercer and Gonzalez, 1981). Again, contours have been corrected to unit fluid density. Production of water from the Magenta is less regionally persistent, as indicated by the presence of dry holes. Note the suggestion of a dividing line based on a transmissivity value of $0.0001 \text{ m}^2/\text{d}$. East of the line the overall transmissivity values are a few orders of magnitude less than for the Culebra at the same points. The contour patterns are almost at

right angles to the Culebra patterns, and also tend to become more widely spaced toward Nash Draw, reflecting an increase in permeability. This increase is consistent with the suggestion of Cooper and Glanzman (1971) that yield is related to size and density of fractures and thinness of overburden. In Nash Draw, surface exposures of Rustler are common, and brittle Rustler rocks are extensively fractured owing to the action of erosion, dissolution of halite and gypsum, and collapse of intervening material.



B. Magenta member

Figure V-3. (cont)

Robinson and Lang (1938) described an occurrence of brine "at the base of the Rustler Formation," particularly near Nash Draw. It is shown that this "brine aquifer," whose discharge is taken to be the seeps and springs on the Pecos River near Malaga Bend, is not so much associated with the Rustler Formation as with the upper boundary of halite in the Ochoan section (Chapter VIII). In this work, the brine occurrence is treated as a consequence, rather than an agent, of dissolution.

Surficial Features

The Pecos River and Laguna Grande de la Sal are perennial bodies of surface water said to have relevance to Delaware Basin groundwater hydrology (Figure VI-1). The possible relationship of the River with the Capitan has already been discussed. The "basal brine aquifer" underlying parts of Nash Draw (Hale et al, 1954) is said to discharge to the River near Malaga Bend.

Drainage in the northeastern Delaware Basin is poor. Just east of the Pecos River flood plain and bordering terraces are several large depressions of low topographic relief. Laguna Grande de la Sal, ~5 mi north of Malaga Bend, is one of those large depressions. Water in Laguna Grande de la Sal is locally derived from a lenticular body of fill that surrounds the lake and playa (Robinson and Lang, 1938). Three sources of water had at that time been identified: surface drainage during rainfall, spring discharge, and potash refinery effluent. The salinity of the lake varies with rainfall. The largest spring (Surprise Spring) at the north of the lake uniformly discharges 115 to 125 gal/min. Surprise Spring contains 57 000 mg/L total dissolved solids (30 000 of which is chloride). In dry periods the lake precipitates a salt crust, indicating saturation with sodium chloride at ~330 000 mg/L. The water of Surprise Spring is not derived from either the Culebra dolomite or the "basal brine aquifer," both of which in that vicinity have chloride

contents at least 60 000 mg/L chloride (Lambert and Robinson, unpublished data). There is apparently no discharge from Laguna Grande southward through the alluvium. Hale et al (1954) made calculations to show that very little of the water from Laguna Grande made its way through the alluvium, diluted by infiltration of surface irrigation water to 10 000 to 20 000 mg/L chloride, to discharge at Malaga Bend. Laguna Grande de la Sal is thus related to surficial processes only, is fed by shallow groundwater discharged through runoff or alluvium, is now receiving waters characteristic of potash refinery effluent (Robinson and Lang, 1938) having percolated through gypsic rocks (Table V-1), and is depleted through evaporation. Thus, Laguna Grande and the question of its relevance to subsurface dissolution processes in the Delaware Basin are not treated further here.

Other lakes in Nash Draw have appeared intermittently in closed depressions. Water levels of these lakes fluctuate in concert with rainfall and recharge from local surface runoff (Bachman, 1974).

Table V-1. Solute in Surprise Spring North of Laguna Grande de la Sal

Element	Concentration (mg/L)
HCO ₃ ⁻	170
SO ₄ ⁻⁻	3470
Cl ⁻	30000
Br ⁻	16
Na ⁺	14400
K ⁺	580
Ca ⁺⁺	2420
Mg ⁺⁺	2000
Sr ⁺⁺	60
Li ⁺	4

Chapter VI. Geomorphological Aspects of Dissolution

Introduction

Karst topography typically develops in regions underlain by soluble rock (limestone, dolomite, gypsum, anhydrite, halite, etc). As the rocks dissolve during exposure to circulating solutions in which their components are not saturated, permeability locally increases and overburden collapses. Thus karst topography is characterized by depressions, internal drainage, valleys without streams, sinkholes, and caves. Karst is most spectacularly developed in carbonate terrain, where the relatively high mechanical strength can maintain underground open space. Evaporite karst is more likely to be obscured by rapid collapse following dissolution, subsurface creep of rock salt, or infilling by aeolian detritus. It was perhaps the resulting pattern of closed-contour topographic depressions that first attracted attention to subsurface dissolution of evaporites in the Delaware Basin. There are positive-relief features, variously described as hills, mounds, castles, domes, "breccia pipes," etc, as well as the sinks that were taken as indicators of collapse. This chapter describes and reviews discussions of origin of various of these features.

Features of Positive Relief

Breccia Chimneys

In 1960, J. D. Vine described a series of "anomalous structural and geomorphic features" circular in map view with various degrees of exposure of internal structure. In particular, four of these just north of Nash Draw (Figure VI-1) involved doming of the Mescalero caliche and brecciation of a central core of Triassic rocks. These four he designated Domes A, B, C, and D; the northern pair (A and B) near Highway 62-180, and the southern pair (C and D) ~2 mi southwest near the Maroon Cliffs (Figure VI-1). Each dome is 1000 to 1500 ft across.

Dome A, cut by a railroad spur, is dissected by erosion to form a shallow basin with 40 to 50 ft of relief. Detailed mapping by Bachman (1980) has disclosed the following: a rim of Mescalero caliche, which also covers the outer slopes, dipping 15° away from the center; Dewey Lake Red Beds and Triassic rocks inside the rim, dipping 15° (locally 20° to 22°) away from the center and in normal stratigraphic relationship with one another; a central brecciated core of previously lithified clay, coarse sandstone, and conglomerate derived from Triassic rocks and separated from the normal Permian and Triassic section by a ring fault; a channel-filling of Gatuña Formation gravel cutting across the ring fault and lying unconformably on the brecciated core; a structureless caliche, with abundant root casts, developed on the gravel. Bachman reconstructed the history of Dome A as follows: collapse dropped Triassic rocks against older rocks along a cylindrical-shaped fault; surrounding Triassic rocks were eroded and Gatuña gravel was deposited in a closed depression formed by the collapse; a shallow intermittent pond developed in the depression, as the Mescalero caliche formed over a wide geographic area; the surrounding area subsided, probably from regional dissolution in the Rustler and upper Salado Formations, leaving a more resistant core and resulting in tilting of beds up to 15° away from the center.

The history of Dome C was similarly described, but because the internal structures of Domes B and D have not been exposed, no statement can be made about them. Vine attributed the origin of the domes to any of several near-surface processes, but when the Mississippi Chemical Corporation (MCC) mine workings in potash Ore Zone 7 encountered well-cemented breccia of angular blocks of halite and anhydrite encased in clay-sized material at a depth of 1100 ft directly below Dome C, the deep-seated nature of Dome C became apparent.

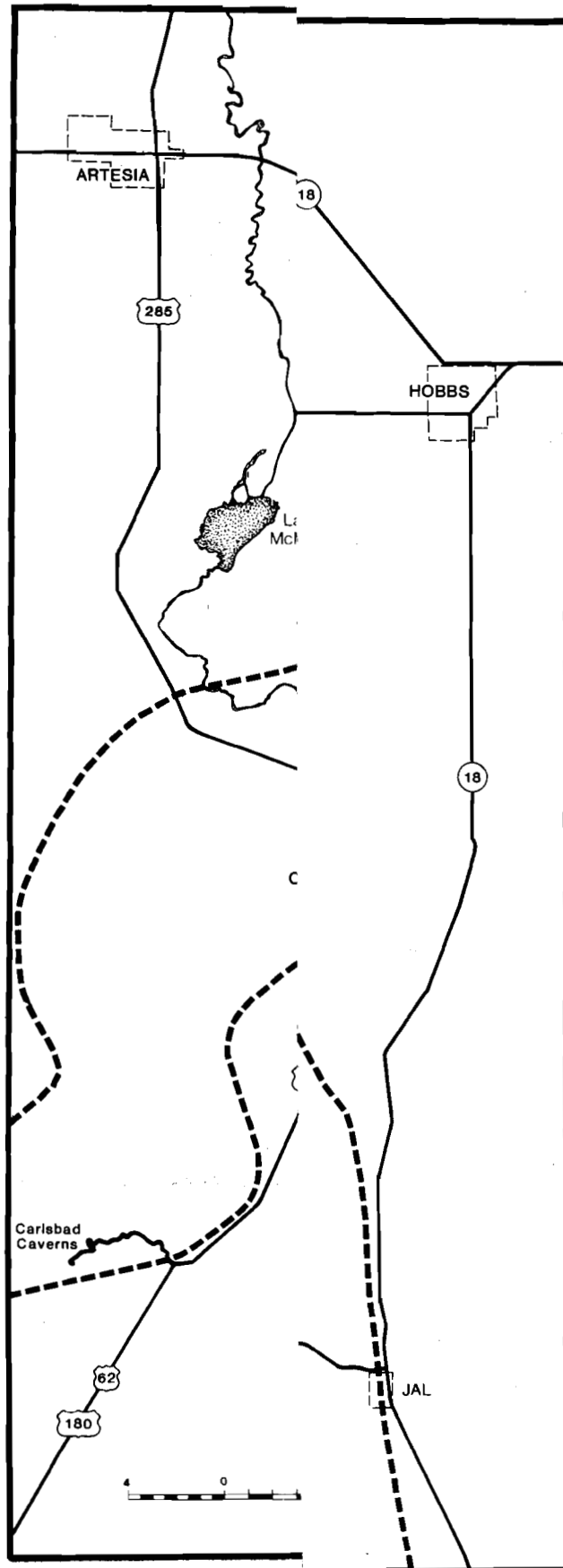


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The discovery of the subsurface manifestation of Dome C inspired several additional hypotheses not only for Vine's domes, but for many other surficial expressions of presumed dissolution at depth. Vine's domes were at first thought to extend to the base of the Castile Formation by Anderson (1978), who proposed that water from the underlying Bell Canyon Formation had stopped its way upward by the brine density flow model (Anderson and Kirkland, 1980) through the entire evaporite section to the surface. Later Anderson (1981) acknowledged that Vine's domes were situated atop the Capitan reef. Nevertheless, the brine density flow model was applied to account for the origin of various surficial domes, sinks, valleys, castiles, and subsurface removal of halite. Bachman (1980) suggested that brine density flow created Vine's Domes A and C at a time when potentiometric levels in the Capitan were higher than at present. Bachman said that incision of the Pecos River (post-Gatuña time) and its breaching of the Capitan reduced the potentiometric level in the Capitan so that it could no longer support the column of water necessary to create such features. Deep drilling of Domes A and C (Snyder and Gard, 1982), core studies, and mapping in the MCC suggested that neither Anderson nor Bachman is correct: that Domes A and C (and perhaps the Weaver Pipe, Figure VI-1) represent collapse into a cavity not in the Ochoan evaporites, but in the Guadalupian carbonates. Snyder and Gard (1982) have fully documented the development of their arguments. They agree with Bachman (1980) that this type of breccia chimney is unique to the reef area.

Karst Domes

The term "karst dome" was introduced by Bachman (1980). Several of Bachman's karst domes were cited earlier by Vine (1960) as possible examples of the same kind of feature as Domes A and C. A particularly high density of such features is found west of the Pecos River near Malaga Bend, in the Queen Lake vicinity (Figure VI-1). Anderson (1978) inferred a possible "deep dissolution" origin for the Queen Lake domes. The Queen Lake domes, described by Reddy (1961), differ from the breccia chimneys in that their cores contain rocks older, as well as younger, than their flanks. In map view they are elliptical to circular,

100 to 2100 ft in diameter, up to 35 ft high. Relationships between cores and flanks are of three types: domes with an arcuate or circular fault enclosing a brecciated core, domes with a normal stratigraphic sequence from center (leached Salado) to margin (Culebra), and domes with a circular fault enclosing a plug. Flanks of the Queen Lake domes consist of the Culebra dolomite member of the Rustler Formation, dipping 4° to 70° (average 35°) outward. Brecciated cores of faulted domes are composed of the Magenta dolomite member of the Rustler Formation, down-dropped into juxtaposition with the Culebra flank. The brecciated cores, 200 to 900 ft in diameter, also contain "blocks of white and greenish-gray gypsum, fine sand, silt, and a few blocks of Culebra dolomite in gypsum." The "plug" (nonbrecciated core) in other domes is a circular outcrop of gypsum with red shale. Reddy (1961) reported two such domes northwest of Queen Lake, with central plugs (bounded by circular or arcuate faults) 200 to 400 ft across, interpreted as the upper leached zone of the Salado Formation. Erosional depressions in some of the breached domes have accumulated Gatuña deposits and caliche. There is also doming of caliche over some of the structures.

Reddy (1961) ascribed a salt-flowage mechanism to the formation of the Queen Lake domes, citing differential unloading of overburden during the cutting of the Pecos Valley, enhanced by tectonic compression caused by basin tilting to the northeast. Such a near-surface process is similar to one proposed by Vine (1960) for the origin of Domes A, B, C, and D. The distribution of "domes," whether expressed as positive topographic relief, rings of Culebra outcrop, or simply as circular vegetation patterns viewed from the air, is from northern Nash Draw southward at least as far as the Texas line (Figure VI-1). Such a wide distribution led Vine (1960) and Anderson (1978) to explain all of them with the same origin. In particular, Anderson cited point-source dissolution at depth as the origin for Vine's domes (including the breccia chimneys), hills in Nash Draw, the Queen Lake domes, and castiles on the Gypsum Plain in Texas.

The unique relationship between Vine's domes (breccia pipes) and the Capitan reef makes unlikely a similar origin for all the topographically positive features ~1000 ft across (Bachman, 1980; Snyder and Gard, 1982). In particular, Bachman's description of

the Queen Lake domes is similar to that of Reddy, but Bachman compared them to "tower karst" of pure limestone in tropical regions. Bachman's description of the karst domes near Queen Lake emphasizes that "the presence of older rocks surrounded by younger strata can be observed only by tracing and mapping rock types on the ground, suggesting that little is to be gained from a reconnaissance of scattered dome-like features and assignment of similar origin to them all." Bachman's explanation of the origin of the Queen Lake karst domes is as follows:

"The karst domes are considered to have formed in the subsurface near the dissolution front of the Salado Formation. Near this front the upper surface of the Salado is highly irregular and dissected by dissolution channels. [cf Nash Draw, discussed below] Insoluble residues of the Salado Formation and overlying Rustler Formation subside into these channels to bury and protect the core of the karst dome as it develops . . . In the vicinity of Malaga Bend the distribution of karst domes appears to be random and they are presumed to have been initiated on an irregular (not controlled by systematic jointing or fracturing) dissolution surface."

In summary, Bachman's view is that the karst domes are erosional remnants of collapsed outliers of dissolution residue in the lower Rustler and upper Salado Formations. No relevant subsurface data exist on the deep-seated versus shallow nature of any of the features Bachman called karst domes. Because of insufficient subsurface control in the Queen Lake area, it is not known whether the domes are underlain by any of Anderson's "deep-seated sinks" in Castile Halite I (Chapter VII).

Karst Mounds

Bachman's (1980) karst mounds are "residual hills of dissolution breccia," circular in map view and a few hundred feet in diameter. Because of their appearance and surficial brecciation, they were once suspected to have originated at depths similar to Vine's Domes A and C. Several such features have been mapped by Bachman in Nash Draw and near Malaga Bend. It is conceivable that several of Reddy's domes near Queen

Lake are, according to Bachman's definition, karst mounds. As with the karst domes, their relief is ~10 to 30 ft. Whereas some karst mounds are capped with Mescalero caliche or Culebra dolomite, some are structureless heaps of insoluble residue from the Rustler Formation. The heaps are not in all cases circular in map view, but may be discontinuous elongated masses up to 1 km long.

One karst mound in particular that bore a surficial resemblance to a karst dome or one of Vine's domes was drilled to a depth of 390 ft (T22S, R29E, Sec 33; WIPP 32, Figure VI-1). Halite was present in the recovered Salado core, including the identifiable marker beds through number 119, indicating a normal stratigraphic sequence and evidence for a nondeep-seated origin for the feature (SNLA and USGS, 1980b). The feature in the center of Nash Draw (T22S, R30E, Sec 17) identified by Anderson (1978) as a "collapsed dome" was mapped by Bachman (1981) as a karst mound developed in the Forty-niner member of the Rustler Formation.

It is probable that Anderson (1978) considered other surficial karstic features in the Delaware Basin to be deep-seated dissolution features, including several of the domes near Malaga Bend and southward on both sides of the river.

Castiles

Castiles are included here for completeness because they are also positive topographic features and have been cited as deep-seated dissolution features. They are primarily restricted to the Gypsum Plain of west Texas.

"Castile" is the term applied to "irregular masses where the gypsum is completely replaced by limestone and banded calcite" (Adams, 1944). They are products of differential weathering standing up above the less resistant Gypsum Plain (Figure VII-1). In area, a "castile" or "limestone butte" covers a few tens of square feet to several acres and rises 10 to 100 ft above the plain. At least 73 have been reported by Kirkland and Evans (1976), who have presented a hypothesis for their origin as follows:

1. With the late Cenozoic uplift of the western Delaware Basin to an eastward dip of 100 ft/mi, the updip migration of oil and gas in the Bell Canyon Formation was stimulated.

2. The western part of the Castile Formation was exposed to erosion, resulting in the karst topography of the Gypsum Plain.
3. Through fracture systems developed in the anhydrite or gypsum (the state of exhumation and hence hydration of the castile is not obvious at this point) water and hydrocarbons were introduced at various points in the Castile Formation.
4. Sulfate-reducing bacteria flourished in the faults or fractures or brecciated zones, and used methane as a nutrient and sulfate as an energy source.
5. Metabolized (oxidized) carbon impregnated the calcium sulfate and converted it to calcite. Hydrogen sulfide was liberated as a by-product and was oxidized to form native sulfur deposits.

Whereas the logic of Kirkland and Evans (1976) is hardly unassailable, especially in view of the previously established limited mobility of hydrocarbons in the Bell Canyon Formation, this report does not contain a critical review of castiles. Rather, the relevance of castiles to other surficial features (domes and sinks) is evaluated.

Anderson's (1978) speculation on a generic relationship between castiles in the lower Castile Formation and breccia chimneys, domes, mounds, and sinks is tenuous at best. It would appear, for example, that if a breccia chimney developed in the lower Salado (say, Dome C), the presence of fractured anhydrite, hydrocarbons, and water (Snyder and Gard, 1982) would give rise to a replacement limestone mass at least at the level of the MCC. "It is unlikely," as Kirkland and Evans (1976) said, "that conditions existed to preclude growth of the microorganisms." This is especially true since traces of hydrogen sulfide were detected in the drilling of a breccia chimney (Snyder and Gard, 1982), yet neither of the cored breccia chimneys exhibits replacement of calcium sulfate by calcium carbonate.

There remains a question of when the castiles were formed. If they are a consequence of uplift of the Delaware Basin, erosion of the Ochoan section would first occur on the west, ultimately exposing the lower Castile Formation. Some of the castiles appear to be still actively generating hydrogen sulfide (Kirkland and Evans, 1976), suggesting recent formation of the castiles, probably soon after removal of the Salado

halite in the western part of the basin. Thus, there is evidence to indicate that castiles did not originate as extremely deep-seated features, and are probably not related to any deep-seated dissolution at a point source. A more likely source for the water to supply the formational process for castiles is in the nearby solution-subsidence troughs of the type described by Olive (1957), which are near-surface, and have recently contained water. Further, the thermodynamic incompatibility between freely circulating, meteoric water (postulated for the Bell Canyon Formation) and the highly reduced hydrocarbons in the Bell Canyon and the castiles seriously limits the amount of continuously recharged water necessary for a "deep dissolution" origin for the castiles.

Features of Negative Relief

Nash Draw

Nash Draw is a partially closed northeasterly trending depression ~16 mi long and 3 to 9 mi wide (Figure VI-1). The Nash Draw 15-min quadrangle was previously mapped by Vine (1963), and contains a nearly complete exposure of the Rustler Formation, Dewey Lake Red Beds, Gatuña Formation, and Mescalero caliche. This valley was recently mapped by Bachman (1981) at a more detailed scale.

Gypsiferous members of the Rustler Formation in Nash Draw have been dissolved to form "caves, sinks, and tunnels in a complex karst topography." Bachman observed that relationships between the Pleistocene Gatuña Formation and the karstic features indicate that many of the topographic features contain horizontally bedded Gatuña; others contain collapsed Gatuña. Thus, the subsidence of Nash Draw rocks spanned the time of Gatuña deposition, and may be active at present (Bachman, 1981).

Bachman ascribed the origin of Nash Draw to the process of "erosion by solution and fill" (first described by Lee, 1925), whereby fractures in soluble rocks are widened by dissolution by surface waters and develop into interconnected grikes, resulting in collapse sinks. The final erosional stage is manifest in a series of karst mounds.

An extensive drilling program in and around Nash Draw revealed that it formed by subsidence as a result

of dissolution of halite in the Rustler and upper Salado Formations. Complete Salado sections are found in Nash Draw below the Vaca Triste bed (near marker bed 116) on the west and below marker bed 103 on the east. Because the results of the Nash Draw drilling program have revealed much information about the dissolution process at depths <1000 ft, a more complete description (as well as possible implications for dissolution at depth) are presented in a subsequent section as a separate topic.

San Simon Swale and Sink

San Simon Sink and the surrounding Swale (Figure VI-1) have been the subjects of much speculation about deep-seated dissolution with little formal interpretation. One of the earliest descriptions was given by Nicholson and Clebsch (1961). San Simon Swale is a 100-sq-mi depression that, contrary to the name, contains no marshy areas. Most of it is covered by dune sand. Just west of its eastern escarpment is its lowest point, San Simon Sink, ~1/2 sq mi in area and ~100 ft deep, which includes a secondary collapse a few hundred feet across and 25 to 30 ft deep. The last reported episode of active subsidence in 1927 left large annular fissures ~5 ft deep at the edge of the sink. The sink contains fill of fine sand and calcareous silt. Both San Simon Sink and San Simon Swale, like the breccia pipes, are located over the Capitan (Figure VI-1).

San Simon Sink is a collapse feature. The sink contains a famous sugarberry tree whose trunk in 1976 was buried in fill up to the level of the lowest large limbs. By then the tree was dead, but it is said that at one time a horse and rider could pass beneath the lowest large limbs. Seismic crews reported drilling over 400 ft in the Sink without encountering Triassic red beds, while elsewhere in the surrounding Swale red beds are encountered at fairly shallow depth (Nicholson and Clebsch, 1961). These findings led Nicholson and Clebsch to suggest that the area of collapse is constrained to the sink area and is not pervasive over the entire Swale. Rather, collapse in the sink areas steepened the local drainage gradient, resulting in headward cutting and widening of the Swale.

The ultimate nature of San Simon Sink, however, remains unresolved. The sink was formerly covered by a lake ~35 sq km in area (Anderson, in Chaturvedi,

1980). Evidence for the lake includes beach and dune ridges and diatomite deposits. Continuous core in the center of the sink was taken to a depth of 810 ft from WIPP 15 (SNLA and UNM, 1981). As yet no refereed interpretation has been published and, consequently, the origin of San Simon Sink is speculative. The origins of Capitan-associated features are not intended for ultimate resolution by WIPP-related work, unless these features are shown by consequence analysis as worthy of consideration as direct threats to the integrity of the WIPP rocks.

The possible amount of stratigraphic offset associated with the origin of San Simon Sink is unclear. Anderson (in SNLA and UNM, 1981) tentatively identified a "light gray sandstone which may be part of the Santa Rosa Formation" in the bottom of WIPP 15 and Triassic Chinle Formation at 545 ft. Triassic palynomorphs are cited as the evidence for the assignment of Triassic age. Anderson (in Chaturvedi, 1980) equated the bottom of WIPP 15 with "petrologically similar" Triassic "sandstone and shale" outcrops on San Simon Ridge (1 mi northwest of the Sink), which Nicholson and Clebsch (1961) mapped as Chinle. The net displacement, if any, cannot be calculated as 900 ft, as Anderson proposes; Jones (in Chaturvedi, 1980) pointed out that Triassic red beds in WIPP 15 were not cored to their bottom, and there are no correlative stratigraphic datum points between WIPP 15 and San Simon Ridge outcrop to allow a calculation of displacement. Bachman and Jones (in Chaturvedi, 1980) both proposed erosional beveling of the Triassic rocks between hole and outcrop to explain the relationship.

Overlying the red beds in the WIPP 15 core, Anderson (in Chaturvedi, 1980, and in SNLA and UNM, 1981) reported 321 ft of uniform grain-sized sand, similar to the Old Mescalero Soil. About 150 ft and upward above the base of the sand section, pebbles of caliche and "Triassic" sandstone occur in the sand. The proportion of clay matrix material increases upward in the sand, culminating in 40 ft of predominantly clay, at 200-ft depth. The base of the clay is calcareous and contains "pluvial" *Artemesia* pollen. Sand overlies the clay, in turn overlain by a white marl at a depth of 35 to 95 ft that Anderson correlated with the surficial diatomite at the sink margin. The upper 40 ft of WIPP 15 core contains plant fragments and gypsiferous playa-like deposits. The basis for this correlation of diatomite and marl is not obvious. Radiocarbon dates on charophyte oögonia from the diat-

omite were $>32\,000$ yr (the maximum limit of radio-carbon dating); gastropod shells from the marl were 20 570 yr old. Anderson (in Chaturvedi, 1980) used the correlation of diatomite and marl and the 20 570-yr determination on the marl to calculate a "maximum estimate" for an average subsidence rate of 0.705 ft/yr.

Bachman (in Chaturvedi, 1980) proposed that the axis of San Simon Swale follows an old (surficial) drainage channel over the reef in which dissolution began; he cited gravels of reworked Triassic and Gatuña northeast of the sink. Bachman did not address the origin of the sink. Jones (in Chaturvedi, 1980) said that he believed San Simon Sink was formed by collapse caused by halite removal in the Rustler. Anderson (in Chaturvedi, 1980) has equated San Simon Swale with the several depressions containing Cenozoic fill, described by Maley and Huffington (1953) and discussed below. He has cited the reef margin occurrence of several such "string-of-beads" depressions as subsidence in response to localized removal of halite by waters moving upward from the Capitan through northwest-trending faults and/or fractures near the reef margin. Several faults of unspecified strike, dip, or location, but of the type postulated by Haigler (1962) or described by Smith (1978) have been invoked by Anderson (1978) to facilitate vertical displacement in San Simon Swale and water movement in many parts of the Basin. Conclusive evidence of faulting associated with San Simon Sink is lacking.

As discussed in Chapter X, the catastrophic 1927 collapse at San Simon Sink is strong evidence of collapse into a phreatic cavity in the brittle Capitan limestone. Natural cavities in halite with plan-view areas as large as San Simon Sink have not been found at the 2000-ft depth (the depth of the Capitan/Salado contact near San Simon Sink); such cavities are not preserved in subsurface rock salt so as to catastrophically collapse.

Bell and Willow Lakes, Nash Well, and Slick Sinks

Four topographic depressions in the Delaware Basin have attracted a certain amount of attention in that they contain Quaternary aeolian deposits that may provide evidence of the age and climate associated with their formation. Gypsum clay dunes are found at Bell and Willow Lakes and in the depression surrounding the Slick Windmills, called "Slick Sink" by

Widdicombe (1980). Gypsum sand dunes are found at "Nash Well Sink," also locally called "Laguna Quarto" (Figure VI-1). Laguna Quarto once contained a wind-mill-supplied stock tank (Nash Well), but artificial surface recharge from nearby potash operations intermittently fills the depression.

Conditions for formation of clay dunes include a nonarid climate with seasonally high temperatures and evaporation rates, seasonal flooding of a nearby lake, and moderate unidirectional winds during the dry season. In addition, there must be a nearby source of alkaline evaporites. Clay dunes at Bell Lake and Slick Sinks overlie (presumably Quaternary) hardpan near the sink edges. They are overlapped by Mescalero (quartz) sand dunes. At Willow Lake, clay dunes unconformably overlie the Gatuña Formation and have no soil cover. The only one of these depressions containing indication of age is Bell Lake Sink, "where (Mescalero) caliche has been faulted and displaced 3 m (10 feet) or more" (Widdicombe, 1980).

The total relief at Willow Lake and Nash Well Sinks is ~ 2 m (10 ft). At Bell Lake and Slick Sinks, the relief is ~ 9 m (30 ft). Widdicombe (1980) attributes the formation of the Willow Lake and Nash Well Sink depressions to "shallow dissolution in the Rustler Formation." Presumably, deep-seated dissolution of the type postulated by Anderson (1978) was not intended to account for the first two depressions because of the abundance of nearby shallow holes that showed normal subsurface stratigraphy. The same appears true for Laguna Grande de la Sal (Figure VI-1). Widdicombe's statement that "offset of more than 3 m (10 feet) at Bell Lake Sink and similar features at Slick Sink indicates that they also apparently originated from deep-seated collapse resulting from dissolution of the underlying evaporites" is not specific to any evaporite bed. The presence of alkaline evaporites in the playas has been used to infer a connection between Rustler waters and the playas, at a time when potentiometric heads were higher. That vertical distance could imply that Widdicombe's use of the expression "deep-seated" refers to depths of ~ 1000 ft. Anderson (1978) generalized the preliminary findings of Widdicombe to conclude that Bell Lake and Slick Sinks are evidence of dissolution of salt as deep as lower Salado. The absence of deep drill holes in these two sinks allows neither confirmation nor denial of any of Anderson's "deep-seated sinks" directly beneath the two surface sinks. More recent borehole

data will be evaluated, as discussed in a subsequent section.

Depressions in the Northern Delaware Basin

There are four named 1-km-dia playa-bearing depressions in the northernmost Delaware Basin whose topographic closures exceed 6 m (20 ft): Laguna Toston, Laguna Plata, Laguna Gatuña, and Laguna Tonto (Figure VI-1). The largest of these (Plata) contains evidence of a collapse origin (Nicholson and Clebsch, 1961). On the north slope of this playa, Dockum (Chinle?) shale crops out ~20 ft above the lake bed. Due south of the outcrop, equivalent red beds were encountered in two potash core holes at depths of 20 and 41 ft below the lake bed. Brokaw et al (1972) report no anomalous thinning of the Rustler beneath Laguna Plata, but nearby in the northern third of T20S, R32E they show a 100-ft depression in structure contours drawn on the top of the McNutt potash zone, and a local thinning of the Salado to 1100 ft just west of the center of T20S, R32E. In spite of these coincidences, the McNutt has not been significantly perturbed to disrupt the workings of the National Potash Company's Lea Mine, which extends into the very same region. Detailed subsurface data from coreholes in and near Laguna Plata are unavailable for examination at this time, and the origin of Laguna Plata is not precisely known. Examination of available stratigraphic data from any existing "Laguna" coreholes would have to be part of future work.

Williams Sink is a closed depression 1-1/2 x 4 mi with ~40 ft of relief in the east part of T20S, R31E and the west part of T20S, R32E. The underlying evaporite beds show normal thickness. It, too, has been undermined by the National Lea workings and is associated with no evidence of origin as a collapse structure.

Nicholson and Clebsch (1961) describe the numerous undrained depressions on the Llano Estacado. Previously termed "buffalo wallows," several were drilled by White et al (1946). Some depressions lacked caliche caprock, and their origin was attributed to leaching of the Tertiary Ogallala caprock or underlying calcareous cement in the Ogallala sandstone.

Pecos River Drainage

The last of the depression-like features to be discussed is also the largest; it extends almost the length of the Delaware Basin, near its center. It is not a new concept that the position of the Pecos River at a given period in geologic time may be either a cause or an effect of subsurface dissolution. "Collapse sinks are common along the Pecos River Valley southward from Roswell" (Bachman, 1974). Some of these sinks formed in historic time, such as the one near Lake Arthur in 1973. Bachman further proposed that the Pecos River follows a major belt of collapsed sinks from Carlsbad to the Texas State Line. The absence of a well-developed flood plain near broad meanders, the tendency of intermittent tributaries to follow semicircular collapse valleys, and the linear scarp along the east side of the river (presumed to have formed along a collapse structure) led Bachman to this conclusion.

The earliest preserved antecedent of the modern Pecos River drainage system is represented by gravel and stream deposits of the Gatuña Formation. Near Pierce Canyon, Bachman (in Chaturvedi, 1980) describes the modern relationship as follows:

"Gatuña drainage was dispersed in a broad system which flowed south to southeasterly. Today, Pierce Canyon flows westerly and cuts at right angles across channel and overbank deposits of the Gatuña Formation. These early stream deposits are as much as 100 feet above the Pecos River. Since Gatuña time, the Pecos River has entrenched itself a mile or more to the west of this ancient drainage."

If indeed these relationships indicate a certain amount of westward migration of the Pecos drainage in the last 600 000 yr, this sense of migration is contrary to a monotonically eastward progression of postulated evaporite dissolution, if the River is said to keep pace with the dissolution "front." Presumably, entrenchment of the modern Pecos River along dissolution-related channels in preference to erosional channels was not always the dominant factor controlling the River's course. Wisconsin-aged (10 000 yr before present) limestone gravels in collapsed sinks in the Pecos riverbed north of Carlsbad are also thought

to demonstrate the importance of dissolution features in controlling modern drainage (Bachman, 1980).

Another possible example of relationships between dissolution and erosion is Maley and Huffington's (1953) accumulations of "Cenozoic" fill. In the areas of fill exceeding a thickness of 1500 ft, there is a coincident thinning of "post-Rustler—pre-Tertiary" (Triassic?) strata (Figure X-1). Since there is no evidence of evaporites (to be removed by dissolution, causing subsidence) in surviving Mesozoic rocks, it can only be concluded that many areas now occupied by fill were once exposed to erosion, including possibly the eastern reef area of San Simon Swale.

The genesis of these erosional and dissolutional features remains unresolved, but geologic data in the cross section in Figure X-1 indicate that the observed depths of the depressions cannot be attributed entirely to dissolution. San Simon Swale may belong to this general category of depression. Thus, Bachman's explanation of San Simon Swale as a surficial drainage channel cannot be entirely dismissed in favor of a unique explanation (dissolution).

The fact that the Gatuña Formation has been mapped at the surface in the areas occupied by the fill led Anderson (1981) to consider the entire 1500 or more feet of fill "late Cenozoic" in age. He then assumed that the fill had accumulated in depressions formed by collapse in response to deep dissolution, and thus inferred that all dissolution in the Delaware Basin took place in the late Cenozoic, since no halite has been found in the depressions described by Maley and Huffington (1953). In actual fact, the "late Cenozoic" date for a maximum age of the fill is unjustified, as no faunal or radiometric data are known on which to base any maximum age. Gatuña-age deposits compose part of the fill, but demonstrate only a minimum age. The poor cuttings record from holes penetrating the fill did not even permit reliable lithologic comparisons with other known rock types (Maley and Huffington, 1953). Thus, if the mere presence of Gatuña in the surficial fill in halite-deficient depressions is alone used by Anderson (1981) to constrain all dissolution to the Pleistocene, this constraint is unsupported.

Numerous small (<1 km across) features in the Pecos River drainage are associated with a topographic depression or collapsed Gatuña Formation. Some of these, as indicated by Bachman (1974), are deflational (Williams Sink, Laguna Plata, and Laguna Gatuña) and are associated with leeward dunes of quartz sand. Some are etched into the regional carbonate cover (dolines). Bachman (1974) cites at the south end of

Nash Draw a sink that collapsed during Gatuña time but was stable during Mescalero time, the feature having since been exhumed. He also cites many locales in Clayton Basin (north of Nash Draw) where blocks of Cretaceous and Gatuña rocks have been preserved as sinkhole debris. Extensive mining activity of the Potash Corporation of America in Clayton Basin has failed to reveal a deep-seated origin for topographic depression contours in Clayton Basin, and such collapse may originate in the Rustler, much as it has in Nash Draw. There are several other examples of collapse of Gatuña alone or intermingled with other rock types, with or without a caliche cover, but these features are not individually discussed in this report.

Summary

As indicated at the beginning of this chapter, it was the surficial features such as depressions, hills, and disrupted drainage patterns that first attracted attention to the possibility of subsurface dissolution of evaporites in the Delaware Basin. These surficial features, together with exhumation of evaporites on the westward side of the Basin caused by eastward tectonic tilting, has led several investigators to identify areas of postulated current or past dissolution in the subsurface, but with little or no surface manifestation of the process.

The beginning of this work described several models of dissolution processes. The near-surface-related process (solution-and-fill) appears to be active in gypsiferous terrain, and it results in karst domes and karst mounds. Phreatic dissolution has certainly been active in limestone terrain in the Delaware Basin subsurface, and cavernous porosity in the Capitan limestone has developed from this process. Probably phreatic dissolution in the Capitan, and not in the evaporite section, has given rise to the breccia chimneys.

The geological evidence is meager at best for brine density flow resulting in point-source dissolution and certain surficial collapse features. The physics of the process (Wood et al, 1982) is favorable; the geological factors are not.

Regional stratabound dissolution is more difficult to identify on the basis of surficial features. How much soluble evaporite rock was originally present in the Delaware Basin versus how much survives is the subject of the next chapter.

Chapter VII. Original and Surviving Extent of Evaporites

The Controversy

Assumptions have been made that discrete evaporite beds were continuous throughout the Delaware Basin during Castile and later time (Anderson, 1981). This assumption was made largely on the basis of the ability to correlate calcite-anhydrite varve couplets in parts of the Castile Formation from subsurface cores in various parts of the Basin (Anderson et al, 1972). This correlation was extended to infer that individual breccia beds in core from UNM Phillips No. 1 (Figure VII-1) were correlative with halite beds in core from Union of California No. 4 University "37" (Anderson et al, 1978). From this it was concluded that halite was intraformationally removed from the Castile and lower Salado Formations, with dissolution proceeding along bedding planes from west to east. To this process Anderson (1981) has ascribed removal of 43% of Halite I, 49% of Halite II, 65% of Halite III (33% pre-Salado and 32% post-Salado), an undetermined percentage of halite associated with Anhydrite IV (undetermined probably because of the difficulty in determining the Castile/ Salado boundary), 73% of the lower Salado (top of Cowden [*sic*] to top of 136 marker bed), and 54% of the middle and upper Salado (assuming an original thickness of 1250 ft). Not reckoned here is the halite between the "Cowden" anhydrite and the base of Salado, because there is evidence that the so-called infra-Cowden salt was deposited on an erosional (dissolutional) angular unconformity that truncates Halite III to the north and restricts it to the region south of T23S. Conversely, the infra-Cowden salt is not well developed south of T24S (Figure VII-2).

The Anderson estimates for amounts and rates of halite removal relative to original volume are made on the basis of several implicit assumptions:

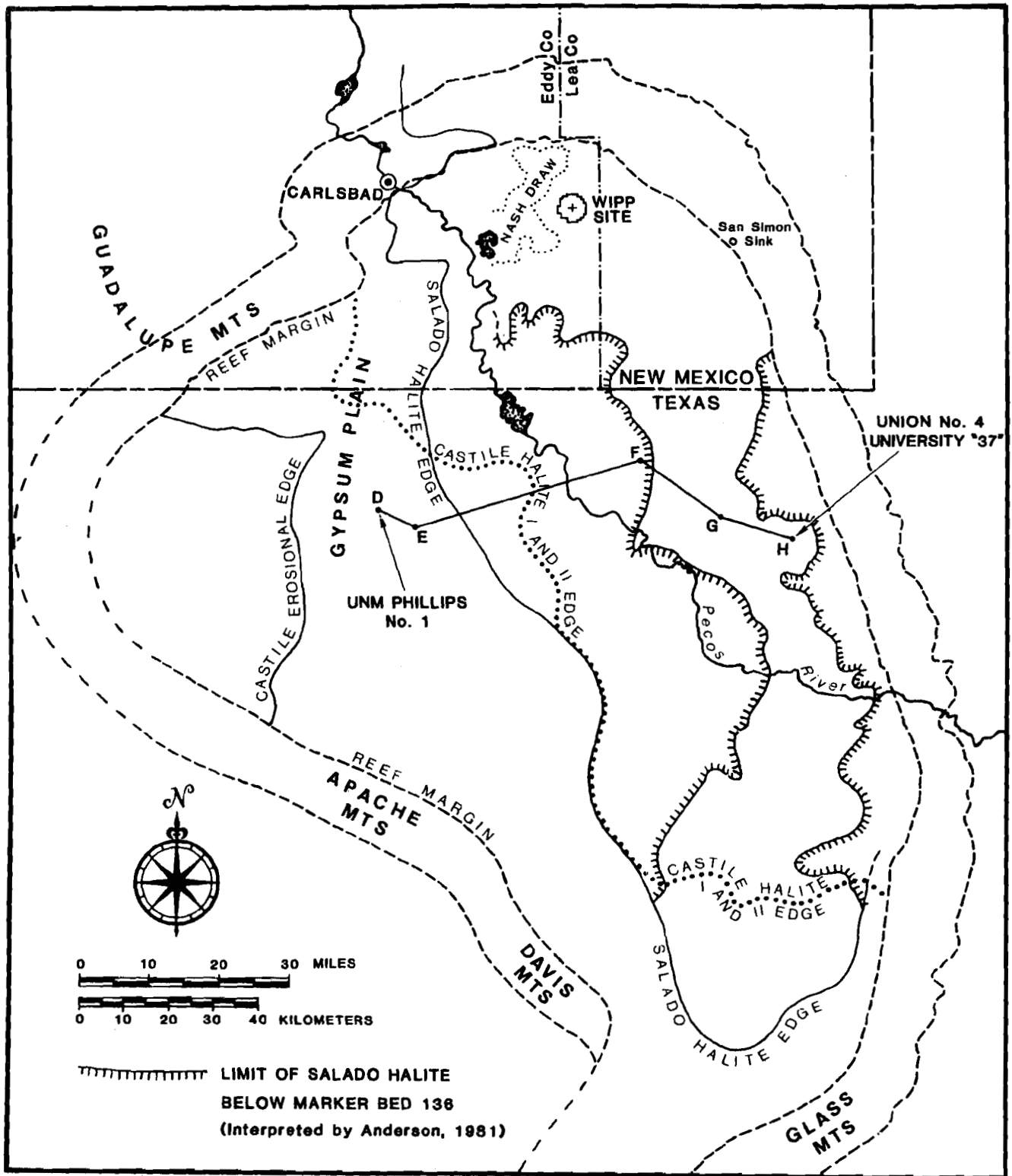
1. All halite units were once uniformly thick (within a few tens of feet) throughout the Delaware Basin at the time of their deposition on a uniformly flat surface.

2. All halite removal, except the development of the erosional unconformity on top of Halite III, is by the dissolution of discrete halite beds by some mechanism prevalent during the late Cenozoic (Pleistocene).
3. Acoustic borehole logs are correlateable to the core of UNM Phillips No. 1, and thicknesses are determined from correct log interpretations.
4. All Castile halite units originally abutted the Capitan.
5. All local variations (reductions) in thickness from a preselected value for a halite bed are due to halite removal by dissolution.

The consideration of validity of the above premises, and the conclusions drawn therefrom, can be condensed to three fundamental questions:

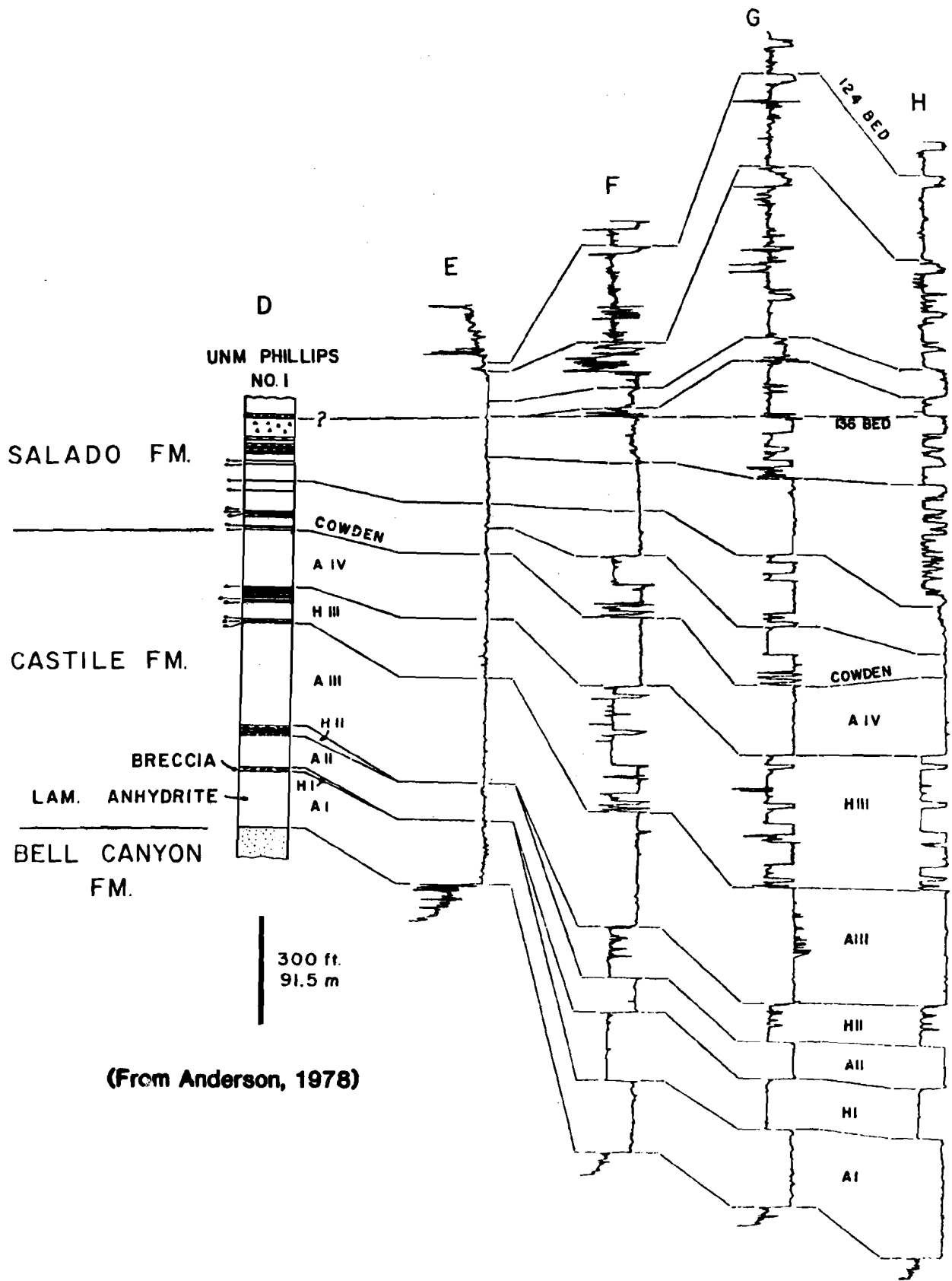
- What was the original volume of halite in the Castile and Salado Formations, particularly at the close of the Ochoan Epoch of the Permian Period?
- At localities of demonstrable removal of halite by dissolution, what was the dominant dissolution mechanism (including state of exhumation, prevalent hydrologic conditions, climate, etc) at the time of halite removal (and/or gypsification of anhydrite)?
- When in geologic time did removal of halite occur?

The last question gains overriding importance, regardless of amount or mechanism of dissolution, in any consideration of future threats to a radioactive waste disposal facility in the bedded evaporites of the Delaware Basin. Rate of advance of such a threat to rock integrity, if known, can allow bounding calculations of duration of rock integrity.

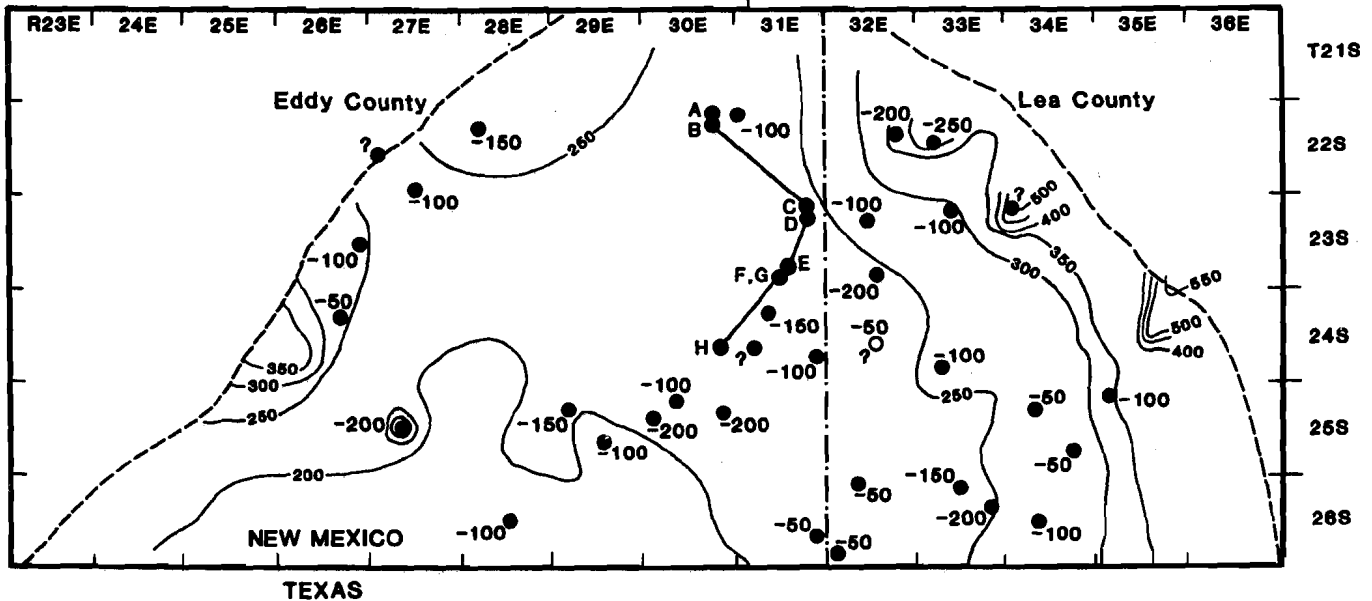


A: Map showing surface projections or outcrop of limits of Castile Formation, Halite I and II, Salado halite, and lower Salado halite (between marker bed 136 and Cowden anhydrite).

Figure VII-1. Subsurface Limits of Ochoan Evaporite Beds in the Delaware Basin (after Anderson, 1978, 1981).



B: Section from core and geophysical logging, from which Anderson inferred a "dissolution wedge" based on local thinning of the halite component in the lower Salado.



ISOPACH MAP OF CASTILE ANHYDRITE I

CONTOUR INTERVAL 50 FEET

INTERPRETED BY SNIDER (1966)
FROM DATA FROM APPROXIMATELY
175 BOREHOLES.



- 150 ● "SINK" IN HALITE I,
FROM ANDERSON'S (1978) ISOPACH
MAP, WITH APPROXIMATE DEVIATION
FROM REGIONAL THICKNESS
- ? ● "SINK" IN HALITE I,
NOT EXPRESSED AS DEPRESSION
CONTOURS, AS INTERPRETED BY
ANDERSON (1978)
- 50
? ○ AREA OF DEPRESSED HALITE I CONTOURS,
NOT INTERPRETED AS A "SINK" BY ANDERSON (1978)

Figure VII-2. Thickness Distributions of Anhydrite I in the Lower Castile Formation, Southeastern New Mexico (after Snider, 1966). Superimposed are locations of "sinks" in Halite I according to Anderson (1978). The location of the section line through holes A through H of Figure VII-3 is also shown.

Variations in the Castile Formation

Halite I and II of the Castile Formation do not abut the reef margin (Figure IV-1). This is not evidence that the Castile halite has been subject to dissolution because of proximity to reef waters. There is evidence, however, that anhydrite was deposited in place of halite at the reef margin. This relationship was not illustrated by Anderson et al (1972), but is well illustrated by earlier work by Snider (1966). Figure VII-2 shows a predominant belt within 1 to 10 mi of the reef margin that contains greatly thickened sequences of nonhalite strata (mainly calcareous anhydrite, massive anhydrite, some limestone and shale). This marginal thickening of anhydrite is not principally due to recrystallization or tectonic movement because in this zone (a) anhydrite laminae are preserved, and (b) thickening of anhydrite occurs around all margins of the basin, with no preferred association with any regional orientation, structural trend, or known stress pattern. Further, large accumulations of anhydrite are known not to flow as readily as halite. A similar depositional absence of halite near the reef margin has been documented by Jones (1972) and Bachman (1980). This depositional relationship indicates that (a) reef margin absence of halite cannot a priori be ascribed to dissolution, and (b) there are depositional variations in thickness in the Castile (and probably Salado) Formation that are unrelated to dissolution. The presence or absence of equivalent dissolution breccias (Anderson, 1981) notwithstanding, removal of some halite (if originally deposited) near the basin margin must have occurred during Castile time (not Pleistocene). The absence of halite was compensated for by a depositional thickening of anhydrite. That the development of the Castile dissolution breccias (Anderson et al, 1978) may be Permian cannot easily be dismissed. That anhydrite should be preserved in favor of halite at the basin margin is reasonable, if less concentrated solutions were pouring into the basin over the reef during Castile time.

Figure 16 of Anderson (1978) is a "summary of deep-seated dissolution features in the northern Delaware Basin." In it, Anderson had located several "deep-seated sinks" in Halite I. Aside from "domes," with or without collapse (later identified by Bachman, 1980, as karst domes or karst mounds), these "sinks"

were the most numerous features. Identification of almost all these "sinks" in Halite I of the Castile Formation are based on interpretation of geophysical (acoustic) logs of individual boreholes or a small number of closely spaced boreholes. The "sinks" were designated as dots in the centroids of closed depression contours in Anderson's (1978) isopach map for Halite I. Relief on these depressions varies between 50 ft (the contour interval) and 250 ft, with the most common values 100 to 200 ft. Locations of these "sinks" in Halite I are reproduced in Figure VII-2, which gives a value for relief of each sink. Apparently, one "sink" identified on the isopach map for Halite I (Anderson's Figure 4) failed to be transferred to the summary map; it is in T24S, R32E and has 50 ft of relief. Conversely, three Halite I sinks identified on the summary map fail to have a corresponding thinning of Halite I: T22S, R27E; T24S, R31E; and T23S, R34E. The last is actually situated on a thickened section of Anhydrite I; the other two correspond to no conspicuous anomaly in the Halite I isopach. The two "sinks" in T25S, R34E actually are the locations of two boreholes whose data were used to describe a single depression contour covering most of that township in the Halite I isopach.

Included in Figure VII-2 is a set of isopach contours for Anhydrite I as given by Snider (1966), which has more detail than those of Anderson et al (1972). We see that several of the Halite I "sinks" occur near the Basin margin, and correspond to thickened sections of the underlying Anhydrite I. One such association is in T25S, R27E, where a 200-ft "sink" in Halite I sits atop a 150-ft "hill" in Anhydrite I. As suggested elsewhere in this chapter, several of these "thinnings" actually represent localized, lesser depositional thicknesses of halite because of a local elevation of substrate above base level.

An example of regional variations in thicknesses of individual Ochoan beds is shown in the cross section, Figure VII-3. The line of this bent section is shown in Figure VII-2, and skirts but does not cross some of the so-called "deep sinks" in the Castile Halite I. Holes A and F, notably, show "thinnings" in Halite I relative to the regional thickness in surrounding holes. Reasons for these two thinnings are different in each hole. The "thinning" in hole A is immediately overlain by a "thickening" in the superjacent Anhydrite II, so that the composite thickness of Anhydrite I, Halite I, and Anhydrite II is identical in holes A and B. In hole

F there is a local rise in the elevation of the tops of the Bell Canyon Formation and Anhydrite I relative to holes E and G on either side. Tops of Halite I, Anhydrite II, and Halite II show no such local elevation. Rather than preferential dissolution of Halite I in hole F to account for the resultant local thinning in hole F, a geologically more reasonable interpretation is the development of a structural uplift of the basin floor near hole F following deposition of Anhydrite I. This local "mound" was later buried by Halite I deposition, which developed a flat-lying upper surface for the later deposition of Anhydrite II.

If preferential dissolution in the Cenozoic had been responsible for the local "thinning" of Halite I in hole A, all higher geological contacts drawn between holes A and B would have been parallel to the contact between Halite I and Anhydrite II. The superjacent local thickening of Anhydrite II due to flowage, as in salt tectonics, is exceedingly unlikely, given the brittle, rather than plastic, behavior of anhydrite. Thus, we can only conclude that thinning of Halite I in hole A, by whatever means, took place in the Permian, before or perhaps during the deposition of Anhydrite II.

To evaluate the magnitude and distribution of thickness variations of the Castile Formation, 348 borehole logs from the Northern Delaware Basin were examined as available from the petroleum industry. Most of the geophysical logs used to infer stratigraphic information were the combination of borehole-compensated acoustic (sonic) and natural gamma-ray logs. Although the acoustic-gamma pair proved most useful for identifying stratigraphic markers, a few neutron-density logs and electrical properties logs were used also, if those logs were all that was available in certain locations, and if the stratigraphic markers could be unambiguously identified on the logs. An isopach map was prepared, showing variations in thickness of the interval between the top of the Bell Canyon Formation and the top of Halite II, a unit of the Castile Formation easily distinguishable on the basis of the five discrete anhydrite layers it contains. The selected interval thus includes Anhydrite I, Halite I, Anhydrite

II, and Halite II, but not Anhydrite III, which locally can be a composite of two or more merged anhydrite layers in the upper Castile/lower Salado transition zone, a difficulty discussed elsewhere.

The resulting isopach map, covering a 5-by-6 township area (Figure VII-4) thus describes the entire lower and middle Castile Formation, not simply variations in individual halite beds. The map shows a profound thickening of the contoured Castile interval along the northeastern reef margin, not a thinning as would be expected if dissolution were active there. Note, however, that thicker accumulation of anhydrite, at the expense of halite, occurs near the reef margin. This feature confirms the findings of Snider (1966) and is illustrated diagrammatically in Figure IV-1. The map also shows an extremely uniform thickness of the lower and middle Castile interval, between 700 and 800 ft thick (except thicker at the basin margin). Note several closed depressions based on "one-hole anomalies," and also several mounds. The composite section isopach eliminates the "sinks" in Halite I (Figure VII-2) of relief >100 ft. The 300-ft "sink" in T25S, R30E lies adjacent to a 300-ft "mound," and these paired features more likely represent structural deformation involving salt flowage, not dissolution, in the Poker Lake Area (Anderson and Powers, 1978).

From the remaining isopach variations (<100 ft in 1 mi), it is concluded that there are no abrupt overall thinnings in the lower and middle Castile attributable to postdepositional removal of halite by dissolution. The total thickness of Castile anhydrite/halite paired units tends to remain the same throughout much of the Delaware Basin (except in areas of erosional exposure) regardless of variations in thicknesses of individual halite or anhydrite beds. The observed variations in the individual beds are thus syndepositional or deformational, not dissolutional. Further, direct evidence (Chapter VIII) shows that water from the Bell Canyon Formation has not participated in dissolution of halite in the overlying Castile Formation.

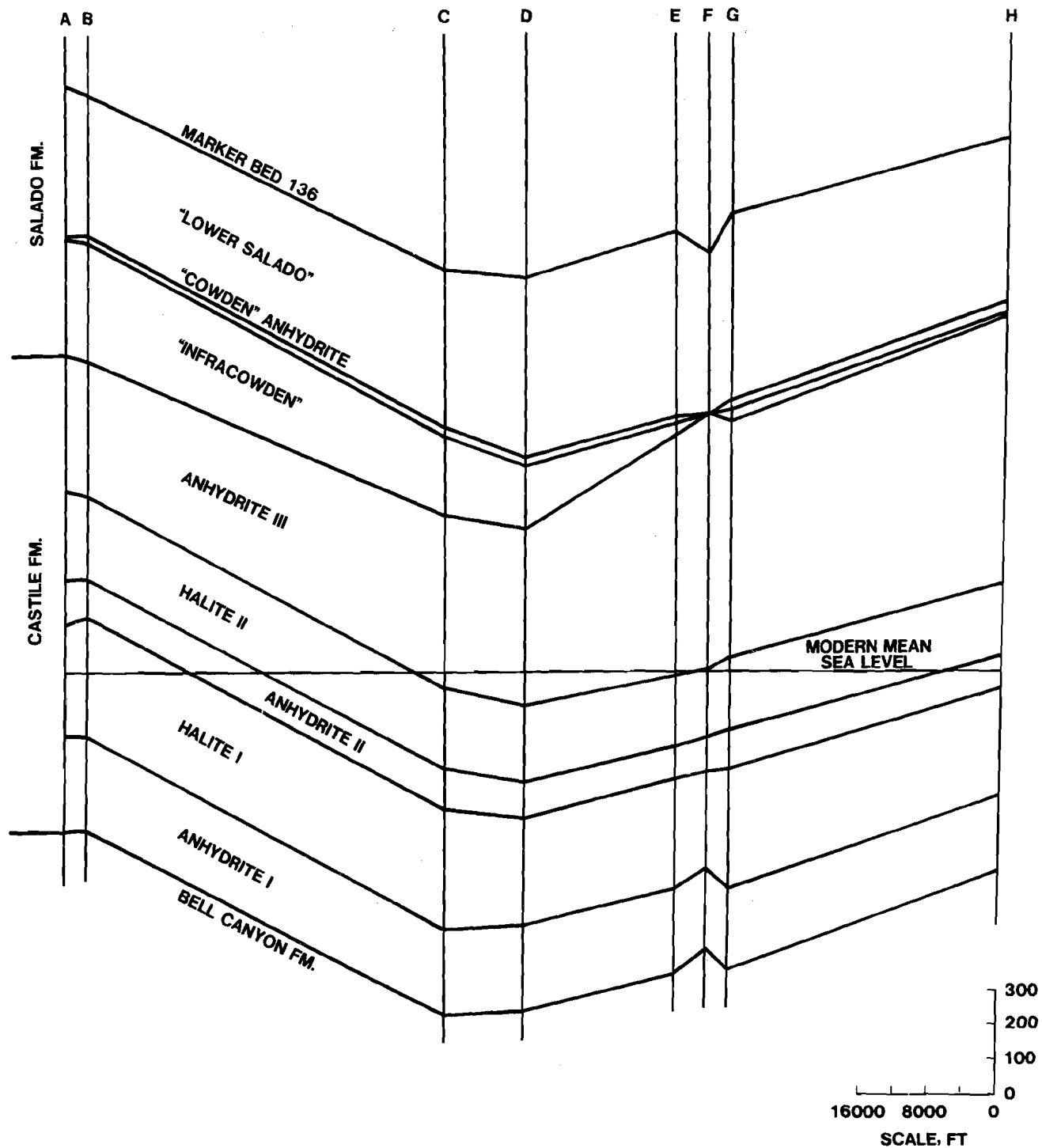


Figure VII-3. Cross Section of Beds in the Castile and Lower Salado Formations. The line of the section is shown in Figure VII-2. This cross section was compiled by S.-E. Shaffer from acoustic logs of the following boreholes.

- A: Phillips Petroleum Co. James "A" #1, T22S, R30E, S2;
- B: Phillips Petroleum Co. James "E" #1, T22S, R30E, S11;
- C: Continental Oil State AA2 1, T23S, R31E, S2;
- D: Max M. Wilson Bauerdorf Federal #1, T23S, R31E, S11;
- E: Patoil Corp. Wright Federal #1, T23S, R31E, S27;
- F: Patoil Corp. Wright Federal #2, T23S, R31E, S33;
- H: Hill & Meeker Shugart Federal 23 #1, T24S, R30E, S23.

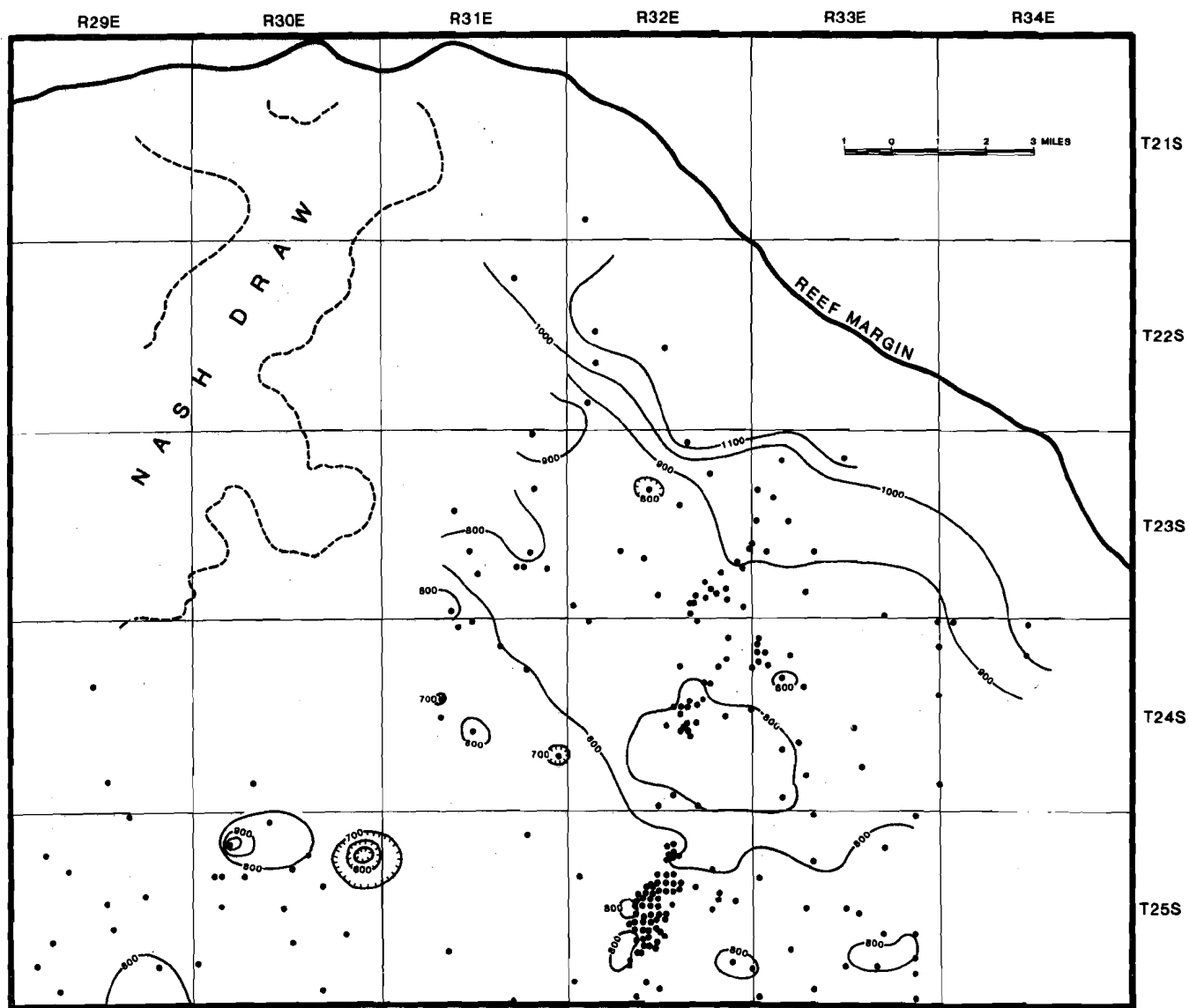


Figure VII-4. Isopach Map of Anhydrite I, Halite I, Anhydrite II, and Halite II in the Castile Formation in the Northern Delaware Basin. This map is based on interpretation of 348 borehole logs in a 5-by-6 township area. Contour interval is 100 ft. "Reef margin" is the approximate basinward extent of the Capitan limestone and the approximate limit of Castile deposition.

Unconformity on the Top of Halite III

Adams (1944) described the boundary between the Castile and Salado Formations along the north and east margins of the Delaware Basin as an angular unconformity. Anderson (1978) examined the unconformity in detail. Discussions of the major conclusions of that examination follow:

1. Halite III (Snider's Anhydrite IV, Table IV-1) occurs as an elongate lens, now confined to the southeastern part of the Basin, with suspected northern outliers and slivers in basin margin structures. Where Halite III occurs, it splits Anhydrite III into a superjacent Anhydrite IV and a subjacent Anhydrite III. Where such a split does not occur, Anhydrite IV is not considered a separate unit, but is merged with Anhydrite III. This relationship gives rise to some basin-wide nonuniformity of nomenclature as described in Chapter IV. Such restriction of Halite III, as well as the overlying Anhydrite IV (Snider's Anhydrite V), led Snider (1966) to formulate the concept of the "Ochoan trough," the eastern portion of the Delaware Basin that by late Castile time was the last remaining depression in the Basin that could receive evaporites. This restriction was interpreted by Anderson (1981) as an Ochoan dissolution remnant, although no direct evidence of Halite III erosion was presented in terms of dissolution breccias in this horizon.
2. The locus of thickest preserved halite section in Halite III is ~30 mi to the south of that in Halites I and II and the infra-Cowden (Salado) halite. This was taken as evidence of dissolution, since it was believed unreasonable for the deepest point in the Basin to have undergone such shifting during the Permian. The analogy between Halite III and the infra-Cowden may be invalid, since significantly different basin geometry, by definition, governs the deposition of the Salado with respect to the Castile; the latter was confined by the Capitan reef (except for the Fletcher anhydrite); the former observed no such restriction.

Anderson (1978) did not consider the possibility that infra-Cowden thinning over the reef was caused by reef-related dissolution from

below rather than by reef-controlled nondeposition.

3. In the northern part of the Basin, the infra-Cowden (Salado) halite occurs to the exclusion of the underlying Halite III. In other words, Halite III thins and infra-Cowden halite thickens northward. The northward merger of Anhydrites III and IV as intervening Halite III thins corresponds to a southward merger of the Cowden into the underlying Anhydrite + IV as the intervening infra-Cowden thins. The merger of the Cowden into the uppermost thick anhydrite (defined as the top of the Castile Formation) is illustrated in hole F, Figure VII-3.
4. The lateral (southward) equivalent of infra-Cowden halite is 1.3 ft of dolomitic mudstone underlain by nodular anhydrite with a dolomitic mudstone matrix, and overlain by the Cowden. Thus the infra-Cowden edge of halite is depositional, not dissolutional.
5. The unconformity was identified in AEC 8 core from the extreme northern end of the Delaware Basin. The suspected Halite III equivalent was 100 ft below the infra-Cowden halite and was described as "about 40 ft (12 m) of disrupted, vuggy, recrystallized, and reorganized laminae," obscuring original brecciated structure (if any). The unconformity is said not to extend into the west central part of the Basin. The permeable zone "associated with clastics or salt residues around the margin of the basin" is said to be "widespread," acting as an active horizon of Cenozoic-age dissolution. Regionally pervasive examples of a dissolution breccia at this horizon are not known. Anderson (1978) also refers to a "thickened section of Anhydrite III, where Halite III is suspected of having been dissolved" traceable southward along the eastern basin margin to the New Mexico-Texas State Line. That thinned Halite III overlies thickened Anhydrite III is more suggestive of nondeposition on a local high, rather than removal by dissolution, provided the effects of thickening by gypsification can be discounted. Discussions of such compensating variations in adjacent beds of halite and anhydrite, so as to maintain a relatively uniform thickness throughout much of the basin, appear elsewhere in this chapter.

Thinned Halite Beds in the Salado Formation

Previous Work

It follows logically that, with the uplift of the Guadalupe Mountains and resultant tilting of the Delaware Basin as a unit to the east-northeast with a 1° regional dip, the evaporite beds to the south and west should be exhumed and exposed to erosion. Halite is not encountered at depths <300 ft and more typically <600 to 800 ft below ground surface (more or less independent of stratigraphy) owing to the action of near-surface fresh waters. In the south and west parts of the Basin, where the oldest Ochoan rocks have been exposed for the longest time, such exhumation would naturally give rise to the dissolution breccias after halite (Anderson et al, 1972), the type section of which is ~311 m of Castile core from UNM Phillips No. 1 (Figure VII-1). From the two widely separated core localities, UNM Phillips No. 1 near the western margin of surviving Castile, and Union No. 4 University "37" near the extreme eastern basin margin, Anderson et al (1972), Anderson et al (1978), Anderson (1978), and Anderson (1981) interpolated with geophysical logs the thicknesses of various halite units across the basin. Location of the section, containing the postulated eastern and western dissolution "wedges" (undercutting of the Salado below marker bed 136 by selective dissolution of the lower Salado) is shown in Figure VII-1A. Figure VII-1B shows the actual section, from which it was deduced that the missing halite between the Cowden and marker bed 136 was removed by dissolution.

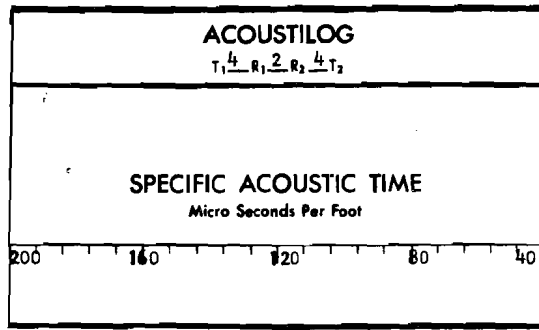
This rationale has been applied to well logs throughout the basin to formulate the concept of the "dissolution wedge." Note in Figure VII-1B Anderson plainly identified the Cowden across the entire section, even though he states in the text (1978, p 22) that the Cowden "cannot be recognized as a separate stratigraphic unit" in UNM Phillips No. 1. In holes E and F in Figure VII-1B the marker bed 136 cannot be clearly recognized, just as the Cowden cannot be recognized at the indicated positions in holes E and H.

The section in Figure VII-1B shows that the existing lithology of the Salado between Cowden and

marker bed 136 consists of interlayered halite and anhydrite. Where the anhydrite beds merged to give an acoustic log signature of anhydrite (individual beds of which *cannot* be clearly recognized), this is interpreted by Anderson (1978) as removal of the halite interlayers by dissolution. Several features of the Cowden-136 section do not support this interpretation:

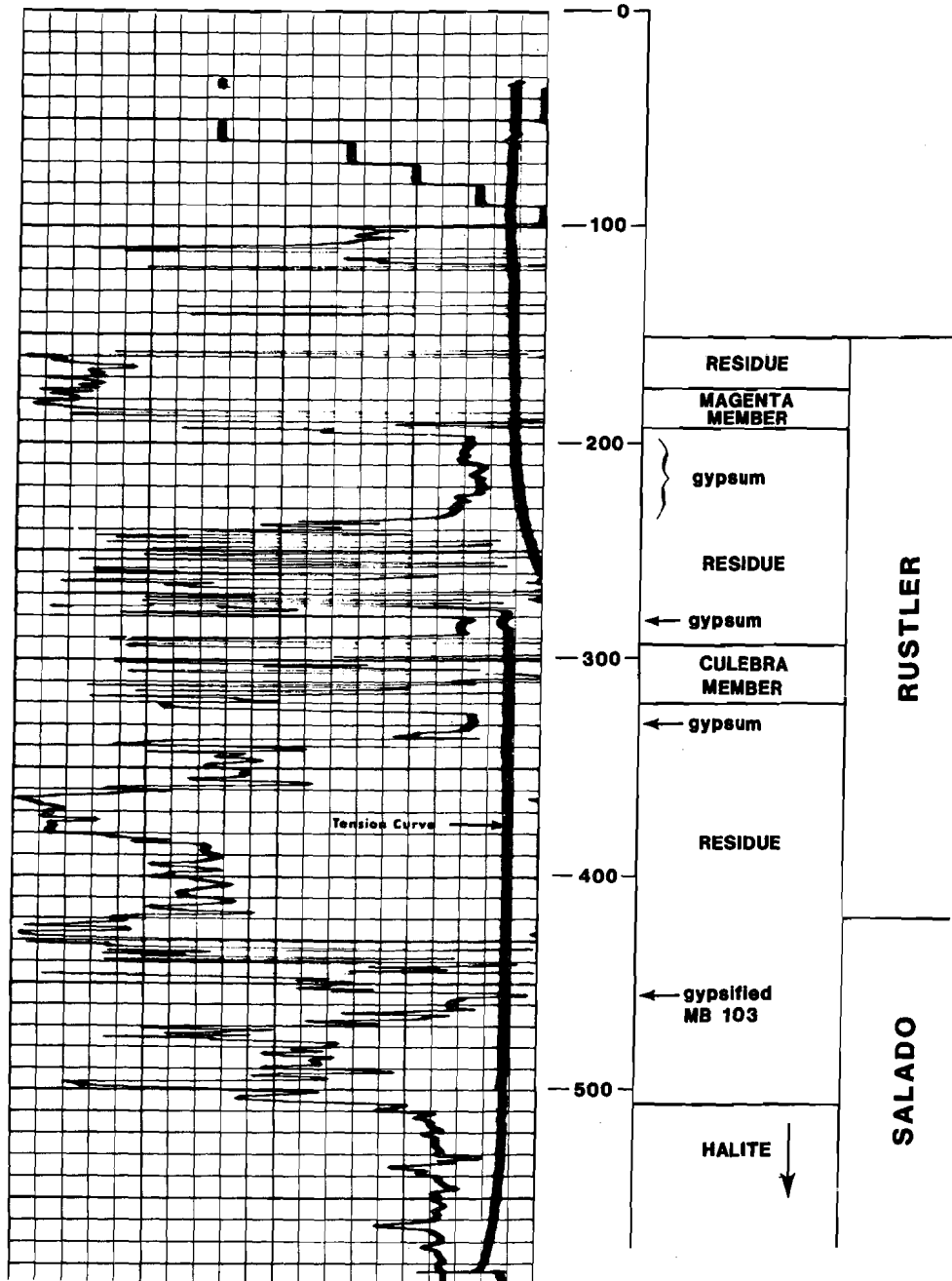
1. Although the ratio of halite to anhydrite varies within the beds between the correlation lines in holes G and H, the total thickness of the Cowden-136 section is the same in holes G and H.
2. Sections of anhydrite between Cowden and 136, which are postulated to be dissolution residue because they contain no interlayered halite, show no log signature of chaotic dissolution residue. The acoustic (?) logs used by Anderson (1978) to make the correlations, on which the "dissolution wedge" is based, bear the signatures of very pure anhydrite, not mixtures of residue and gypsum as would be present in abundance if large quantities of fresh water were circulating in subsurface open space. A discussion on the nature of gypsification is taken up in Chapter VIII.
3. There is a variation in thickness of ~100 ft in an anhydrite bed in holes F and G, which contains no halite.
4. Core was available only from the halite-free UNM Phillips No. 1 in the western (exposed) part of the Ochoan (west of the westernmost occurrence of Ochoan halite). There is no direct evidence (e.g., cores containing dissolution breccia) from any of the other holes to demonstrate that halite (if originally present) has been removed by dissolution.

Coring in Nash Draw has revealed the nature of residue remaining in the wake of active dissolution, showing that vuggy porosity, recrystallization, and open-space filling are indeed likely subsurface developments. An active dissolution zone (as indicated by core examination) would yield an acoustic log signature distinctly discernible from that of merged anhydrite beds. Examples of acoustic logs containing signatures of dissolution residue known from core are given for WIPP 27 in Figure VII-5A and for WIPP 29 in Figure VII-5B.



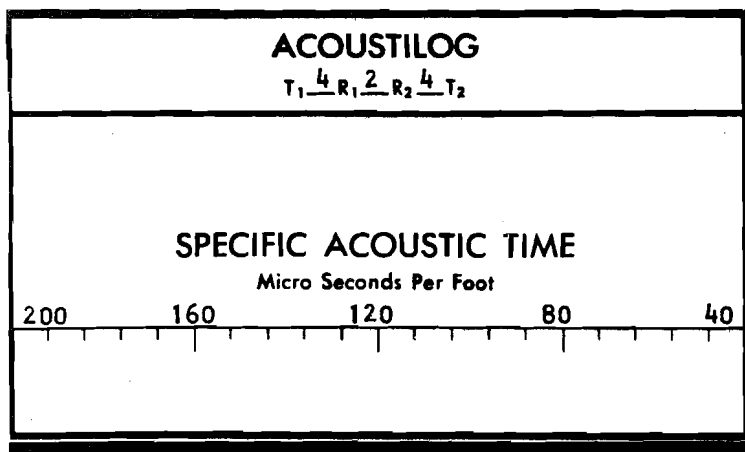
WIPP 27

DEPTH **LITHOLOGY**
 (ft) (FROM CORE DESCRIPTIONS)

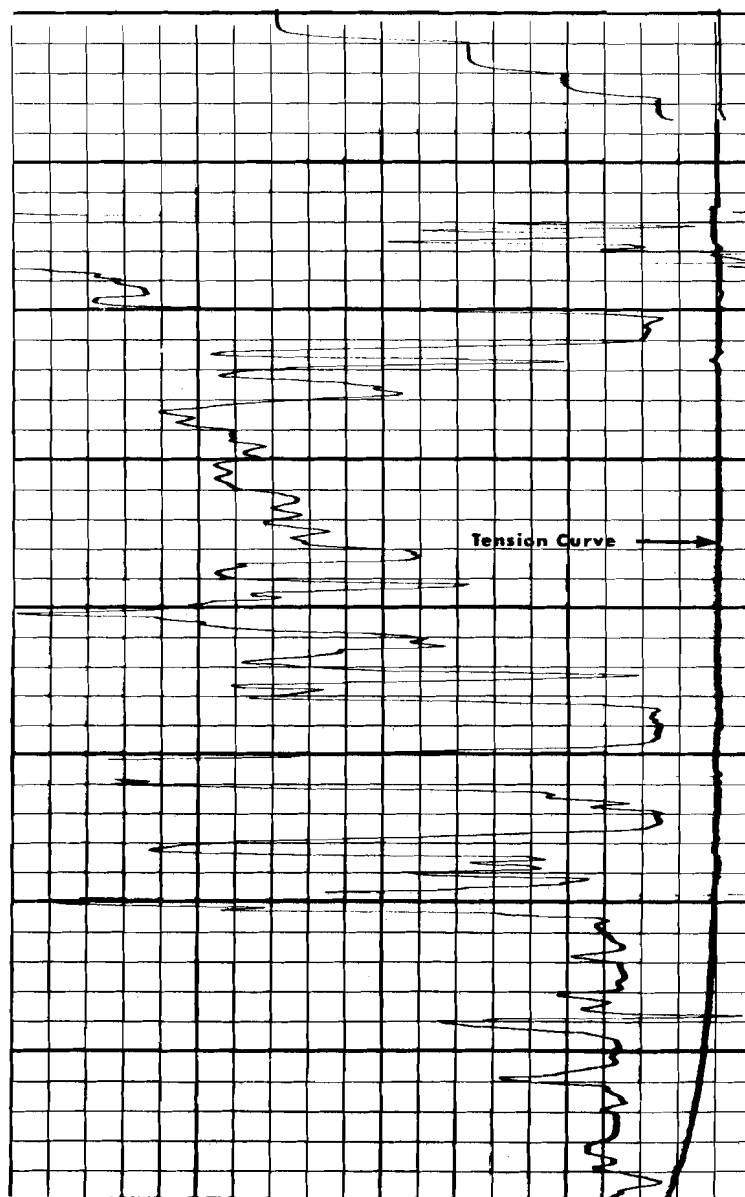


A. WIPP 27

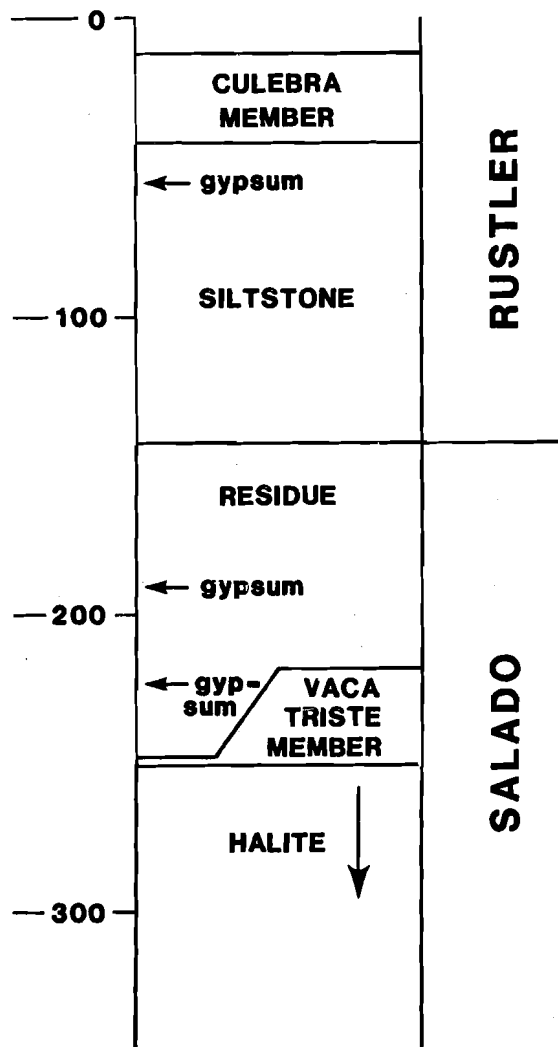
Figure VII-5. Geophysical and Lithologic Logs in the Rustler and Upper Salado Formations in Nash Draw. Shows wildly chaotic acoustic signatures of dissolution residue consisting of mudstone, gypsum, and connected porosity (lost circulation zones).



WIPP 29



DEPTH **LITHOLOGY**
 (ft) (FROM CORE DESCRIPTIONS)



B. WIPP 29

Limitations of Log Interpretation

Throughout most of the Delaware Basin, borehole geophysical logs are the only available data from which to derive the subsurface stratigraphy. Every piece of information derived from logs depends upon interpretations that may be neither unambiguous nor inescapable. For example, in the absence of core there would be difficulty in identifying specific numbered marker beds of the Salado Formation, given only the geophysical log depicted for WIPP 29 in Figure VII-5B. Such ambiguities give rise to errors in identifying stratigraphic marker beds, which in turn can indicate false thinnings or thickenings in various horizons between certain marker beds.

Possible errors in stratigraphic interpretation of geophysical logs become especially significant in the case of an isolated borehole that may provide the only data for miles around. Thus, an apparent local "thinning" in such a hole could be interpreted as a sink with several hundred feet of isopach relief. The resulting "one-hole anomaly" should be treated with a certain amount of suspicion. Most of the "deep-seated sinks" in Figure VII-2 are one-hole anomalies.

Among the Salado stratigraphic horizons most difficult to recognize uniquely and uniformly throughout the Delaware Basin are the Cowden anhydrite and marker bed 136, the two units used by Anderson (1978) to delineate the "dissolution wedge" in the lower Salado. It has been shown that the stratigraphic position of the Cowden varies between the Salado and the Castile throughout the basin (Figure VII-3, also discussed in Chapter IV), and that in the southern part of the basin the Cowden merges into the uppermost thick anhydrite of the Castile Formation as part of a depositional, not dissolutional, relationship. Further, the Cowden cannot be uniquely recognized with the same high degree of confidence throughout the basin, as it occurs with other anhydrite beds of similar thickness (Table IV-2) and its gamma signature does not always have the characteristic spike at its base. Thus, the Cowden is an unfortunate choice for a marker bed on which to base isopach diagrams because of its inherently complex stratigraphic associations.

Similarly, marker bed 136, which varies in thickness and degree of polyhalitization even throughout the Carlsbad potash district (Jones et al, 1960) is an unfortunate choice. Figure VII-1B clearly shows the

immense difficulty in uniquely identifying this bed, especially in holes D, E, and F, and probably in holes G and H as well.

For these reasons, the present study has ignored the Cowden anhydrite, and has selected as a lower stratigraphic marker either the top of Halite II (the base of the uppermost thick anhydrite of the Castile Formation) or the top of the Bell Canyon Formation. The upper boundary of this study is still the marker bed 136. Logs in which this horizon was not clearly recognizable were not used. Such choices will (1) take into account the mutually compensating thicknesses of adjacent halite and anhydrite beds (see previous discussion), and (2) avoid drawing a lower boundary in the stratigraphically nonuniform region of the Castile-Salado transition. A disadvantage to such a choice is the fewer number of boreholes providing log data for lower horizons, as most of the WIPP-specific boreholes of T22S, R31E do not penetrate to sufficient depth.

Because of the ambiguities in identification of the Cowden and 136 beds in Figure VII-1B and Figure 15 of Anderson (1978), and the absence of the characteristic acoustic log signature of dissolution residue, there is no clear evidence of either an eastern or western "dissolution wedge" preferentially developed in the lower Salado.

To evaluate the magnitude and distribution of thickness variations of the Ochoan evaporite section, an isopach map was prepared for the interval between the top of the Bell Canyon and the top of marker bed 136. The same boreholes were used as in the Castile study discussed before. Selection of this interval has several advantages. The interval spans the transition zone between the Castile and Salado Formations; for many years there has been disagreement over where to draw a contact between the two. The interval spans the Cowden anhydrite; the difficulties of identifying this marker bed were discussed above. The interval also spans the interval containing the mutually exclusive relationship between Halite III in the Castile and the "infra-Cowden" in the Salado (see "Unconformity in Halite III," above). Finally, and most importantly, the interval incorporates several beds of halite and anhydrite; this avoids the problem of thinned individual beds of halite that, it has been shown, were improperly interpreted as strong evidence of preferential dissolution. The disadvantage of the equivocal identification of marker bed 136 in some logs caused some

logs to be eliminated from this study. If dissolution has indeed been a major process in the area of study, the composite Ochoan section will be abruptly and conspicuously affected, since there will have been no opportunity for evaporite dissolution to be compensated by later sedimentary infilling in Ochoan time.

The resulting isopach map is Figure VII-6. Throughout the study area the thickness of the Castile and lower Salado Formations consistently remains >2000 ft. Only to the southwest, toward the erosional exposures in the western part of the basin and west of the line, "edge of upper Salado dissolution," do isopachs consistently fall below 2000 ft. In most of the study area, composite thicknesses are between 2000 and 2200 ft. Gradients of local variations (mostly thickenings) are ~100 ft/mi, as observed for the Castile. Thus, throughout the study area, except in areas of known dissolution (Nash Draw and west of the line defined by Brokaw et al, 1972), there is no evidence of uncompensated thinning in the selected interval. Anderson's (1978, 1981) line, "edge of lower Salado salt," is shown in the extreme southwestern part of the study area. With respect to this line, the isopach study shows no anomalous thinning northeast of Anderson's line, and hence no preferential removal of any salt horizon in post-Permian time, aside from the dissolution in the Rustler and upper Salado. Anderson's line is coincident with the edge of features described by Maley and Huffington (1953), which are discussed next.

Coincidence of Fill and "Missing" Halite

Anderson (1981) inferred a genetic relationship between the sediment-filled troughs of Maley and Huffington (1953) (Chapter VI; Figure X-1) and the western limit of lower Salado halite (Figure VII-1). In the Balmorhea-Loving Trough (Figure X-1), Maley and Huffington contoured 1300 to 1400 ft of detrital(?) fill and <200 ft of salt in the combined Castile and Salado Formations. The basis for the salt thicknesses at the control points (boreholes) has never been fully explained. Just west of the trough the salt thickness rises abruptly to 900 ft and gradually thins to nothing 15 to 20 mi west of the Pecos. Maley and Huffington defined the fill and salt isopachs of the

Balmorhea-Loving Trough on the basis of fewer than 30 boreholes in a 1000-sq-mi area surrounding the trough. North and east of the trough, isopach contours show that the salt thickens to a maximum of ~2700 ft in the northeastern part of the Delaware Basin with an extremely uniform regional gradient of 136 ft/mi.

Thicker salt accumulations surround the Balmorhea-Loving Trough, suggesting that it is an incised feature. Also, the logs presented by Anderson (1981) do indeed show a thinning of Salado salt in the troughs, by whatever mechanism, although no preference is apparent for salt thinning restricted to any particular horizon. (The individual marker beds are not identified in the log correlation diagrams.)

Nowhere west of the Balmorhea-Loving Trough does the salt achieve a thickness of >1000 ft (Maley and Huffington, 1953), so the east edge of the trough marks the western limit to many salt beds (including the lower) in the Salado. Consequently, preferential dissolution of the lower Salado is a threat to the northern Delaware Basin only if it can be shown that it actively extends eastward and northward from the trough. A search for such a feature (which was not found) was described in the previous section of this work. The possible role of a filled depression and how the importance of that role might be tested are discussed in Chapter X. The presence or absence of undermining dissolution notwithstanding, the Balmorhea-Loving Trough lies between the intact sections of Ochoan evaporites in the north-central Delaware Basin and the halite-free gypsified exposures of the Ochoan at the western basin margin.

Ochoan Depositional Variations

It was long assumed by many workers that the Ochoan evaporites were precipitated from a solution of primary marine origin, identical to modern ocean water. Adams (1944) accepted this hypothesis "without question." Much later, Dean and Anderson (1978) demonstrated characteristics of the Ochoan evaporites "for which there is no modern analogue." One of the implications of evaporite mother-liquor identical to modern seawater is that if all primary evaporites were preserved, the ratio of amounts of the dominant minerals, anhydrite and halite, should be identical to that obtained by evaporating modern seawater.

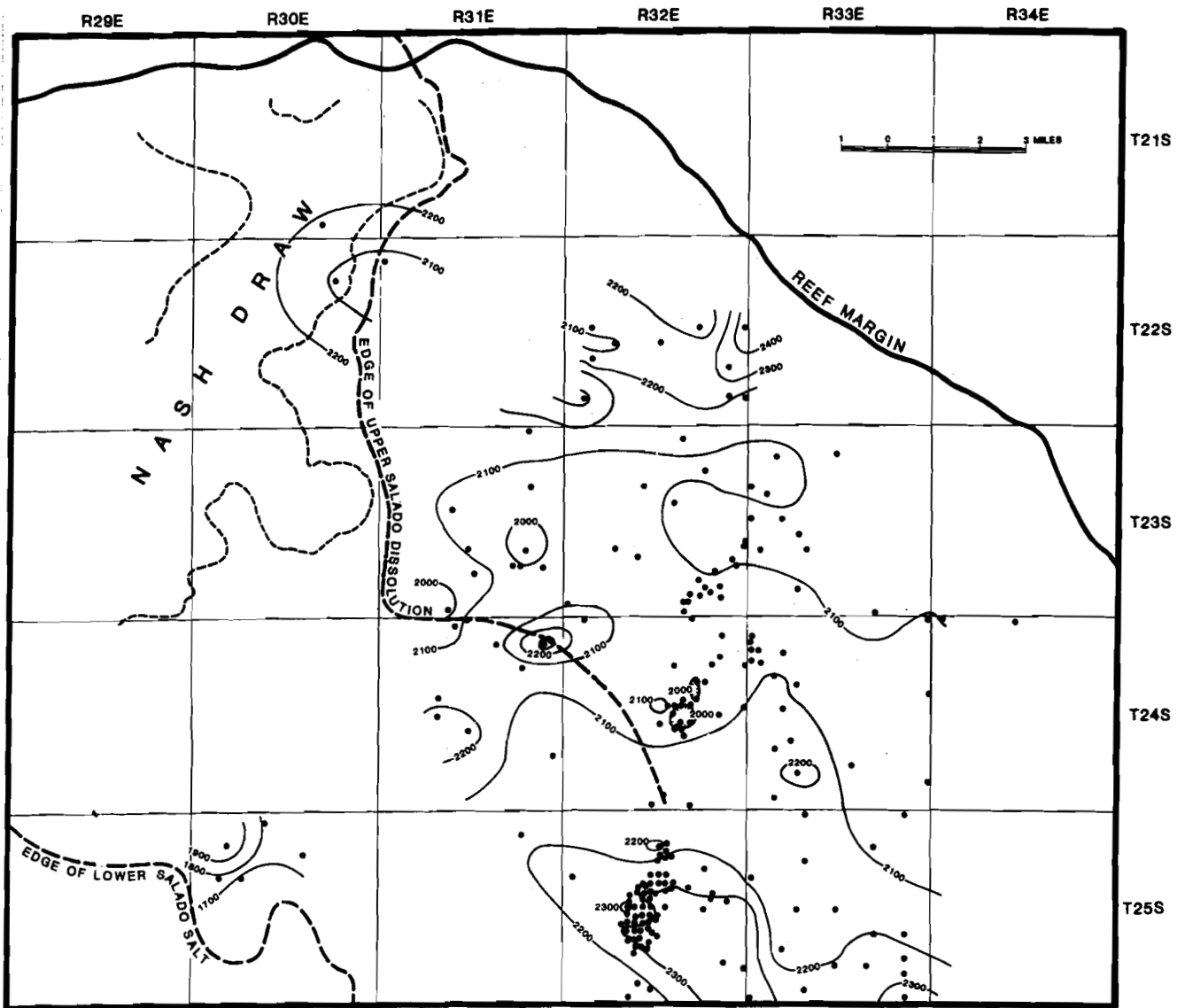


Figure VII-6. Isopach Map of the Castile and Lower Salado Formations in the Northern Delaware Basin. This map, like Figure VII-4, is based on interpretation of 348 borehole logs in a 5-by-6 township area. Contour interval is 100 ft, drawn on the composite thicknesses of the Castile Formation and the Salado Formation beneath marker bed 136. "Reef margin" is the same as in Figure VII-4. "Edge of upper Salado dissolution" was defined by Brokaw et al, 1972, as the western limit of Rustler halite, and the eastern limit of halite removal in the Salado. "Edge of lower Salado salt" is from Figure 16 of Anderson (1978).

Dean and Anderson (1978) showed that each calcite lamination precipitated on an anhydrite lamination represented a basin-wide freshening of the water during deposition of the thick anhydrite units of the Castile Formation. Laminations of calcite and anhydrite are easy to rationalize in terms of Usiglio's experiment (Table VII-1). At 81% evaporation by volume (0.190 L remaining) the last calcite is precipitating and giving way to the onset of calcium sulfate precipitation. (The problem of primary gypsum versus anhydrite is discussed in Chapter III and is not relevant to the present discussion.) At some point during anhydrite deposition, influx of fresher water (presumably carrying a new supply of CaCO_3 in solution) enters the basin, and the cycle begins anew with calcium carbonate precipitation, the water in the basin no longer saturated with calcium sulfate. During the formation of alternating carbonate-sulfate laminae, the solution never achieves saturation in sodium chloride, except perhaps at the culmination of a salinity cycle (Dean and Anderson, 1978). The accumulation of thick, pure, anhydrite-free halite in much of the Castile and Salado Formations is thus problematic: the state of the seawater solution must be such that either no calcium sulfate remains in solution, or calcium sulfate saturation is not exceeded, lest anhydrite be unavoidably intermixed with the halite. In the former condition, the inflowing solution must have been depleted in calcium sulfate before the nonsaturated NaCl solution enters the basin to deposit more pure NaCl. In the latter condition, NaCl-saturated solution that is calcium-free could obtain some calcium sulfate by partial dissolution of old sulfatic precipitate before entering the basin. In both cases, the solution must stay depleted or undersaturated in calcium sulfate throughout deposition of pure halite. Some factor external to the basin must exert exceedingly tight control over the solution composition to keep the anhydrite/halite ratio low. Such a consistently low ratio (<1/100 by weight) requires a seawater solution that has been evaporated at least 96% by volume before it ever enters the basin. This unlikely situation sustained over several hundred to several thousand years would give rise to several tens to several hundreds of feet of relatively pure halite as observed today.

Table VII-1. Precipitation of Calcite, Gypsum, and Halite From Seawater at 40°C
(Weights in grams precipitated from an original volume of 1 L)

Volume (L)	CaCO_3	$\text{CaSO}_4 \cdot 2\text{H}_2\text{O}$	NaCl
1.000			
0.533	0.0642		
0.316	trace		
0.245	trace		
0.190	0.0530	0.5600	
0.1445		0.5620	
0.131		0.1840	
0.112		0.1600	
0.095		0.0508	3.2614
0.064		0.1476	9.6500
0.039		0.0700	7.8960
0.0302		0.0144	2.6240
0.023			2.2720
0.0162			1.4040

Results of experiments of Usiglio, adapted from the description of Krauskopf (1967).

It is difficult to imagine the foregoing circumstances as a quiet depositional environment containing primary seawater that precipitates a thick accumulation of anhydrite-calcite laminae, followed in uninterrupted succession by a thick accumulation of relatively pure halite. Whereas the geological conditions of anhydrite deposition are relatively straightforward, those of halite deposition are not. An alternative easier to envision, and perhaps quite plausible, follows the conclusion of Adams (1969), who contended that the bromide deficiency in the Salado Formation with respect to known primary seawater evaporites indicated dissolution of older (Guadalupian?) evaporites outside the basin and transport of the NaCl-enriched dissolution brine into the basin, to be precipitated as "second-cycle" halite, even more depleted in bromide than the original.

Sodium chloride in solution entering the Delaware Basin (not as seawater, but as a dissolution brine to be

evaporated) could account for a nonuniform, discontinuous deposition of halite independent of anhydrite, both in terms of geographic distribution and origin mechanisms.

Another process that can give rise to nonuniform thicknesses and discontinuous distributions of halite beds is nonuniform anhydrite deposition. Whereas the calcite-anhydrite laminae couplets in the Castile can be correlated from point to point across tens of kilometers, Figure VII-3 shows that anhydrite beds are not uniformly thick throughout the basin. Dean and Anderson (1978) and Anderson et al (1972) have illustrated discontinuities in the laminae sequence. Dean and Anderson (1978) describe the occurrence of nodular anhydrite in the later parts of salinity cycles. The "dissolution breccia" identified by Anderson et al (1978) is very similar to that described by Schlager and Bolz (1977), who described intraformational breccias in the Zechstein very similar in appearance to those described by Anderson et al (1972). The Zechstein, otherwise similar to the Delaware in depositional geometry, was proposed to contain clastic accumulations of anhydrite in deep water, as well as the evaporitic thinly laminated anhydrite. Some of the Zechstein clastic anhydrites were interpreted by Schlager and Bolz (1977) as turbidites, others as "breccias, folds and sliding planes, probably all formed by syndepositional slumping." Thus, syndepositional, mechanical processes can give rise to (a) local variations in thickness ("mounds") in anhydrite, (b) a corresponding thinning of overlying halite, and (c) development of intraformational features that have previously been described as dissolution breccias of residual anhydrite after halite. Resultant thinnings of halite overlying thickenings in anhydrite were thus improperly interpreted as thinning by dissolution or epigenetic tectonics involving significant amounts of creep in brittle anhydrite. There is, in fact, evidence of discontinuities in salt beds in the upper Castile/lower Salado that are of depositional nature in response to basin geometry (e.g., the "infra-Cowden" halite, and probably much of the lower Salado).

Conclusions and Implications

Missing halite in (especially) the lower Salado Formation cannot be uniquely attributed to removal

by Cenozoic dissolution, as hypothesized by Anderson (1981). Other processes that can result in nonuniform thickness of halite beds are: (a) removal of halite by erosion in certain portions of the evaporites subaerially exposed shortly after deposition during the Ochoan, and (b) nonuniform deposition of halite during the Ochoan, resulting in missing halite interbeds among the anhydrite beds in the lower Salado. In the absence of characteristic geophysical log signatures indicating basin-wide intraformational dissolution breccia, and in the absence of core from such breccia, the proposal that "wedges" of halite in the lower Salado are missing because of "deep" dissolution in the Cenozoic is unsupported. Subaerial erosion of halite in the Ochoan, or nondeposition of halite in certain areas could also account for irregularities in halite distribution. Maley and Huffington (1953) suggested that the incision of a subdentritic drainage pattern "left its imprint upon the underlying evaporite deposits," creating dissolution channels of surficial origin that later received detrital fill. The features they described may thus be formed as a consequence of erosion rather than undermining of the evaporite section by preferential dissolution in the lower Salado Formation. The clay accumulations near the Castile/Salado boundary in ERDA 10 have not been established as dissolution residue, for example, nor were breccia beds, permeability, or dissolution brine observed in that hole.

The presence of the erosional unconformity, resulting in an irregular distribution of Halite III, opens the possibility of nonuniform, discontinuous halite deposition throughout the Delaware Basin during more of the Ochoan than previously indicated. Anderson et al (1972) have alluded to an eastward depositional thickening of Castile anhydrite beds, and a northward depositional thickening of Castile halite beds. This suggests, as they observed, that the processes governing deposition of anhydrite versus halite may significantly differ, together with direction of supply of evaporite mother-liquor. Adams (1972) has suggested that introduction of saline water to the Basin during Castile time was not at regular intervals. Adams (1969) proposed on the basis of bromine profiles that Salado halite was not a primary seawater precipitate, but second-cycle, periodically dissolved and washed in from the "shelf" evaporites of, say, the Tansill Formation. Bachman's (1980) interpretations

of geophysical logs resulted in a succinct statement of an alternative to "dissolution wedge" removal of halite:

"Instead, interbedded halite and anhydrite in the Castile interfinger and are discontinuous. These evaporites were deposited in individual and subordinate depositional pans within the Castile depositional basin. Beds of halite thin and wedge out towards the western edge of the basin and it appears that thick beds of halite were never deposited near the margins of the Delaware Basin during the Castile time."

Additional evidence for the discontinuous nature of deposition in the Ochoan comes from Jones' (1972) description of the McNutt potash zone in the middle Salado:

"The uncertainty concerning the number of [cyclical sedimentation] units is related in part to the overlap and pinchout of units and in part to the presence of many corrosion surfaces and intraformational unconformities along which unknown, and probably different, thicknesses of sediments were removed during the periods of flooding that interrupted evaporite deposition from time to time."

Examples of such flooding are probably preserved as clastic material in the Salado Formation, such as the La Huerta silt, locally in the lower Salado, and the Vaca Triste siltstone, at the top of the McNutt potash zone (Adams, 1944).

Summary

Through examination of available and inferred stratigraphic information and detailed discussion, it has been shown that

1. Variations in thickness of individual beds of the Castile Formation are generally depositional features, not products of dissolution. The total thickness of the lower and middle Castile remains relatively uniform except near the western erosional exposures in the basin.
2. The Anderson (1978, 1981) interpretations of geophysical logs to show preferential thinning of the lower Salado are questionable, due in large part to a failure of the selected borehole logs to allow clear identification of key marker beds and dissolution residue.
3. Sediment-filled troughs described by Maley and Huffington (1953) are incised into the western portion of the Ochoan sequence, in an otherwise uniform westward-thinning of the Ochoan in response to increasing westward exposure of originally deeper rocks. There is no basis for constraining their maximum age to "late Cenozoic," nor has a genetic relationship been established between the Balmorhea-Loving trough and the thinning of the Ochoan to the west.
4. East of the line of upper Salado dissolution (partly coincident with Nash Draw), the basinal Ochoan sequence shows only subtle variations in total thickness between the bottom of the Castile and marker bed 136. Thus, there is no evidence of preferential, uncompensated thinning of halite beds by dissolution in that area.
5. Depositional variations in the Ochoan can be rationalized not on the basis of post-Permian dissolution, but by different factors governing supply of anhydrite versus halite to the evaporite basin. Thus, while the deposition of anhydrite in the Castile may have been uniform throughout the basin, the deposition of halite in the Salado may not have been at all uniform.
6. Interpretation of thinning of individual halite beds as dissolution without consideration of adjacent beds is unwarranted.

Chapter VIII. Evidence of Interaction Between Evaporites and Groundwaters

Introduction

In any study of the regional extent of subsurface evaporite dissolution, it is important to be able to recognize the products of the process and to relate the products and the process to local and regional geology. This chapter describes subsurface samples, both rocks and water solutions, that have participated in dissolution to various degrees. First, an example is presented in which stratigraphic relationships are thought to exercise control over dissolution (Nash Draw). It is seen that dissolution at a point source, even over aquifers, is not a predominant mechanism in observable active dissolution. Solute and isotopic constituents of waters are then shown to reflect the kinds of mineral assemblages with which they have been in contact. Finally, examples of waters are presented whose constituents are not simple solutions of evaporite minerals, and evidence is presented against their having flowed through the evaporites from a local surface- or subsurface-supplied source.

Shallow-Seated Dissolution Associated With Nash Draw

A series of six core holes (WIPP 25, 26, 27, 28, 29, and 30; Sandia Laboratories and United States Geological Survey, 1979a, 1979b, 1979c, 1979d, 1979e, 1980a) have penetrated the Permian Dewey Lake Red Beds, Rustler, and uppermost Salado Formations in or near Nash Draw, a depression 5 to 10 mi wide and ~250 ft deep (Figure VIII-1). Nash Draw was thought to be a product of dissolution of the Ochoan evaporites at shallow depths and subsequent overburden subsidence. The processes of evaporite dissolution and erosion by solution-and-fill (Lee, 1925) contributed to formation of the depression since times at least as distant as the middle Pleistocene (Bachman, 1981). The draw is bounded by a series of gentle ridges locally

developed into steep escarpments in which are exposed the Dewey Lake Red Beds and portions of the Rustler Formation as well as the Triassic Santa Rosa sandstone and younger rocks (Vine, 1963; Bachman, 1981). Core holes and other exploratory holes were drilled in the northern Delaware Basin evaporite section to support geological investigations for the WIPP. Many of these holes, located near Nash Draw, have provided stratigraphic data and cores for petrographic examination and various geochemical analyses (Figure VIII-1).

Holes in the "W" series (WIPP 25-29) were drilled in Nash Draw expressly to investigate Nash Draw. Other holes (potash assay, hydrologic test, stratigraphic) are shown to the east, and the P-series have generally less detailed stratigraphic control due to incomplete coring. Trends of two bent cross sections, with stratigraphic control supplied by the holes, are shown by lines connecting the well locations. One section (Figure VIII-2A) runs across the draw, over Livingston Ridge to the east, and out onto the rolling sand-dune-covered plain. The other (Figure VIII-2B) runs approximately along the longitudinal trend of the draw. The sections illustrate the general topography, the stratigraphy, and zones of rock/water interaction that led to dissolution (notably of halite), and hydration (notably of anhydrite to gypsum).

The transverse section (Figure VIII-2A) shows that in the eastern part of Los Medaños the stratigraphic section is virtually complete, including a full complement of halite in the Rustler Formation. The uppermost occurrence of halite migrates down-section toward the west into Nash Draw. The uppermost halite boundary occurs below the Rustler-Salado contact only west of Livingston Ridge.

The uppermost unit of the Rustler Formation is anhydrite, and east of Livingston Ridge the top of nongypsified anhydrite in most places coincides with the top of the Rustler Formation. This indicates that the hydration of anhydrite by waters from above the Rustler has been minimal in the Los Medaños area.

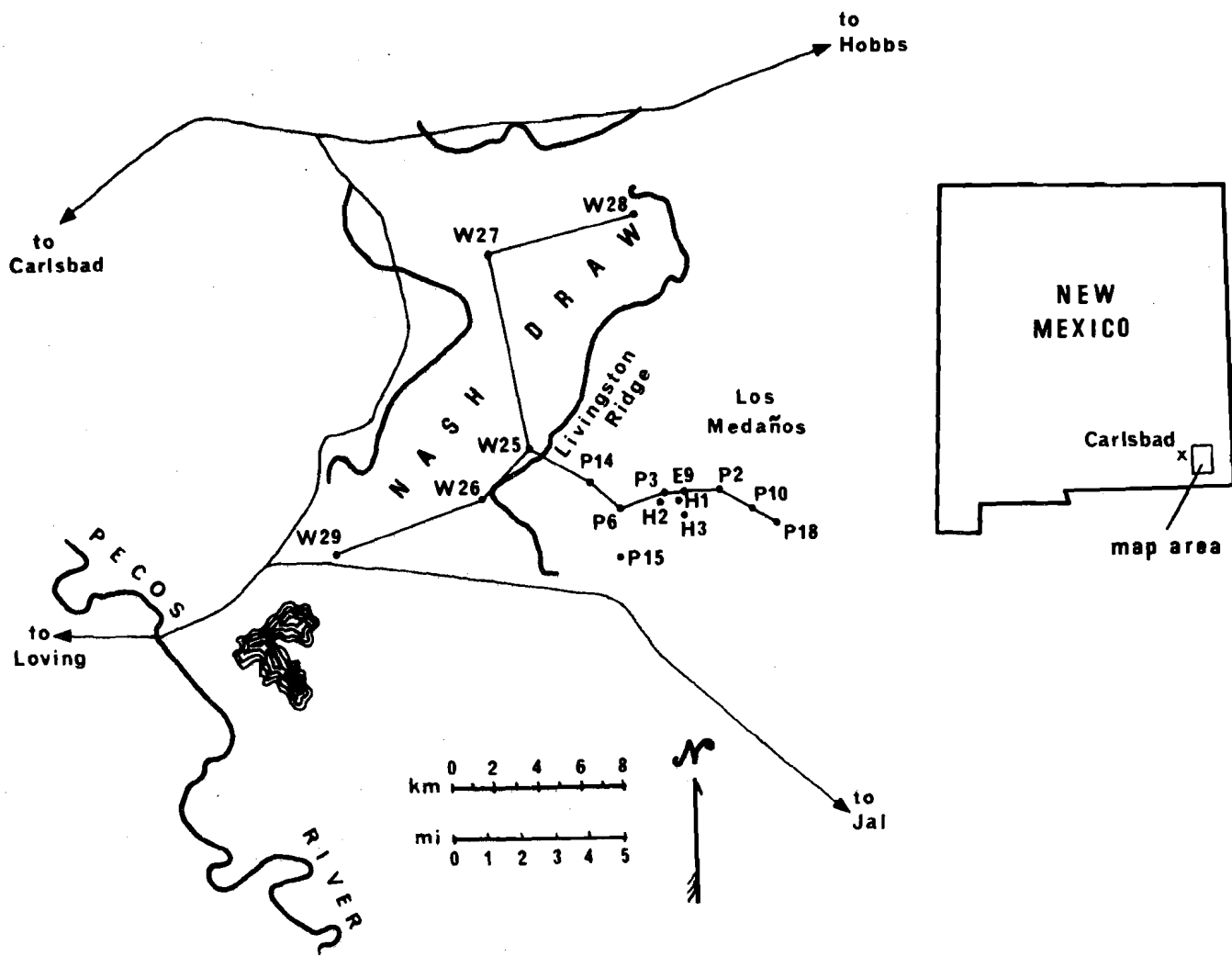
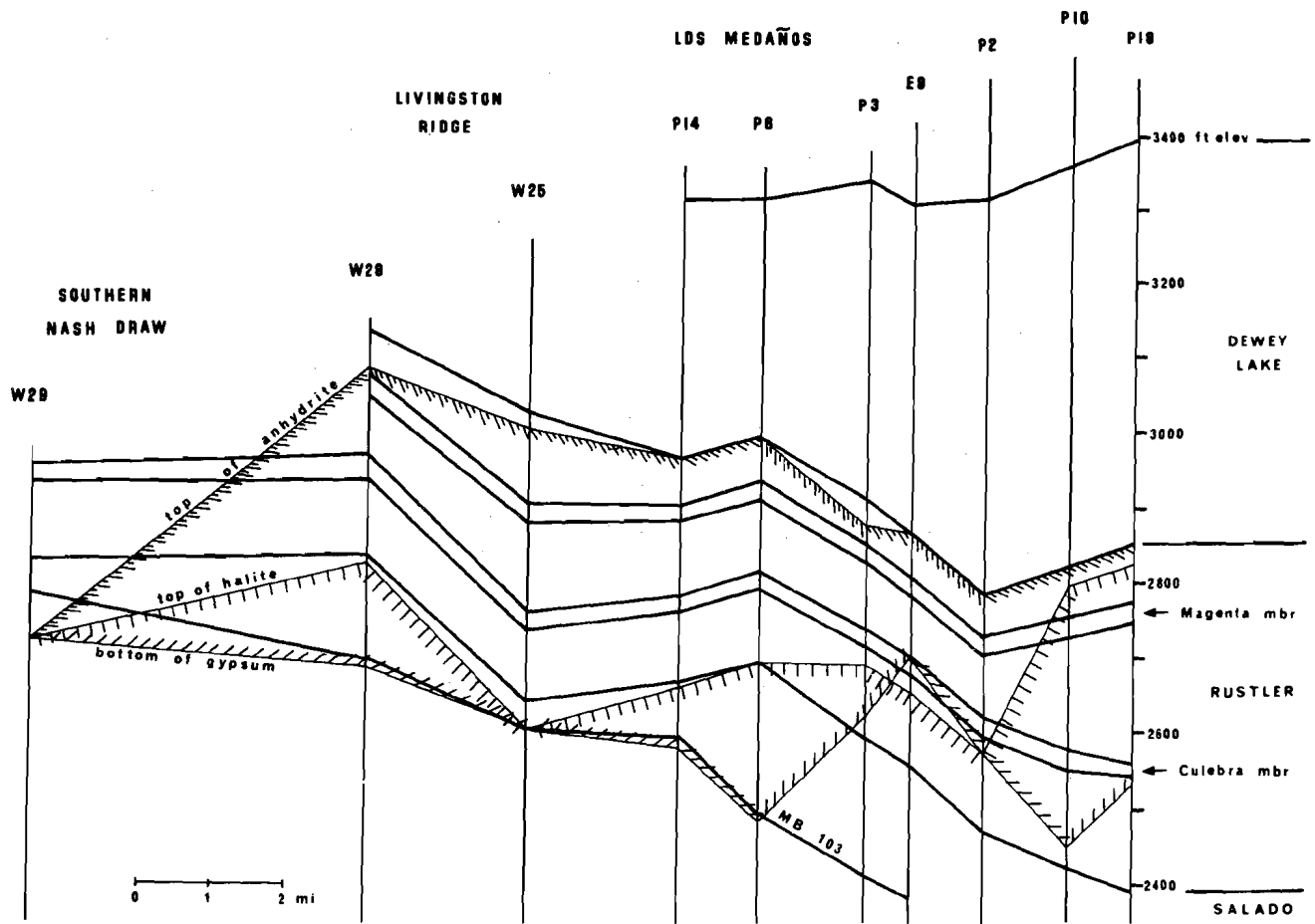
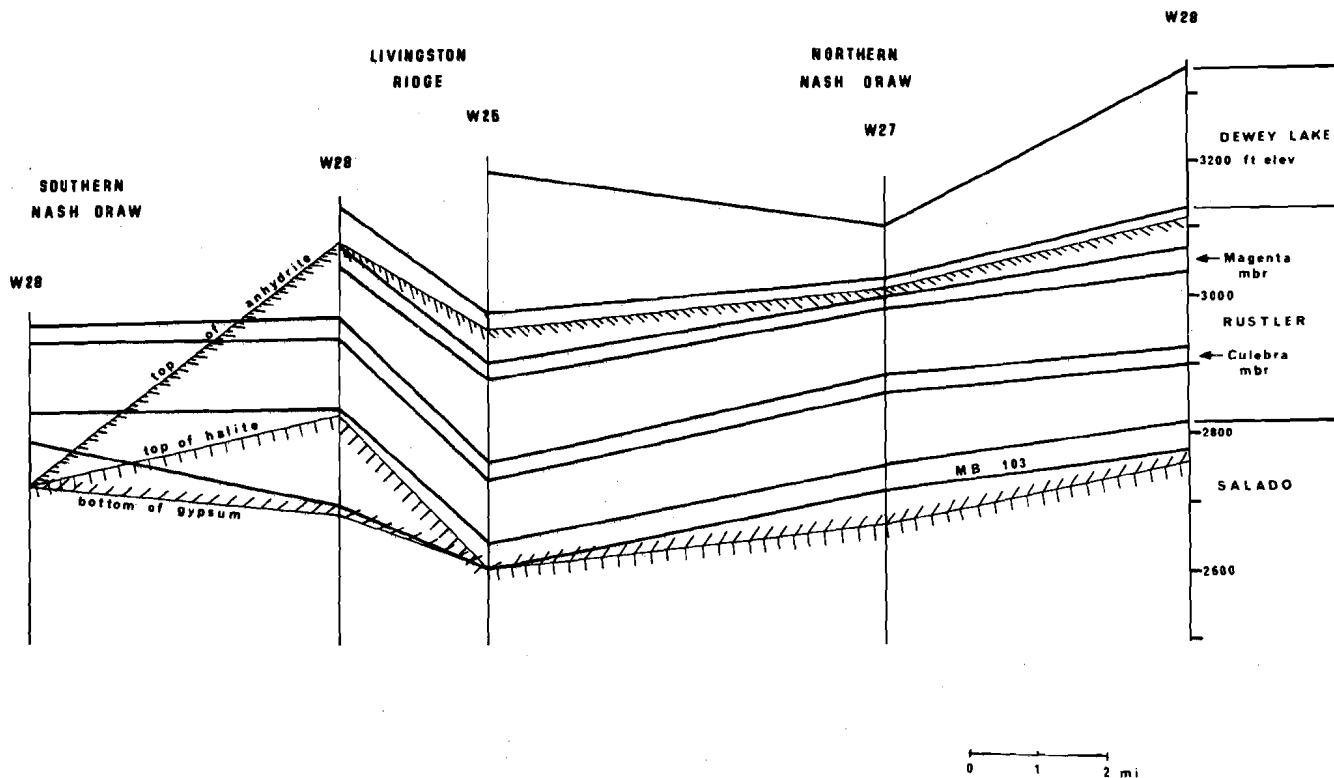


Figure VIII-1. Map of Nash Draw and Vicinity, Southeastern New Mexico, With Locations of Boreholes



A. Transverse Cross Section

Figure VIII-2. Cross Sections of Nash Draw From Core Data. Section lines are given in Figure VIII-1.



B. Longitudinal Cross Section
 Figure VIII-2. (cont)

Gypsification has occurred within the Rustler Formation. In the vicinity of P3, E9, and P2, the uppermost occurrence of halite and the lowermost occurrence of gypsum nearly coincide with the Culebra dolomite member of the Rustler Formation, the most consistently productive aquifer in the region. Farther to the west (P6 to P14), note two phenomena: the "bottom of gypsum" abruptly becomes associated with the anhydrite of the marker bed 103; the "top of halite" line crosses into the Salado, representing the complete loss of Rustler halite.

The interaction of rock and water leading to gypsification of anhydrite has resulted in a succession of interfingered layers of gypsum and anhydrite. The preservation of anhydrite within dominantly gypsic intervals indicates that the water for conversion of anhydrite to gypsum did not originate as a body

migrating vertically across bedding planes (Figure VIII-3), converting all anhydrite in its path to gypsum. The residual anhydrite appears to occur at discrete stratigraphic horizons; gypsum develops most commonly in close proximity to the Magenta and Culebra dolomites (under Los Medaños) and to the 103 marker bed (under Nash Draw). These observations suggest that gypsification is related to phenomena associated with bedding planes. The occurrence of gypsum below the top of halite (particularly in marker bed 103) shows that gypsification can occur in anhydrite, even though adjacent overlying halite was locally isolated from undersaturated solutions. Further, the principal source of "fresh" water underlying Nash Draw need not be the "brine aquifer" at the top of halite, as previously identified (Hale et al, 1954).

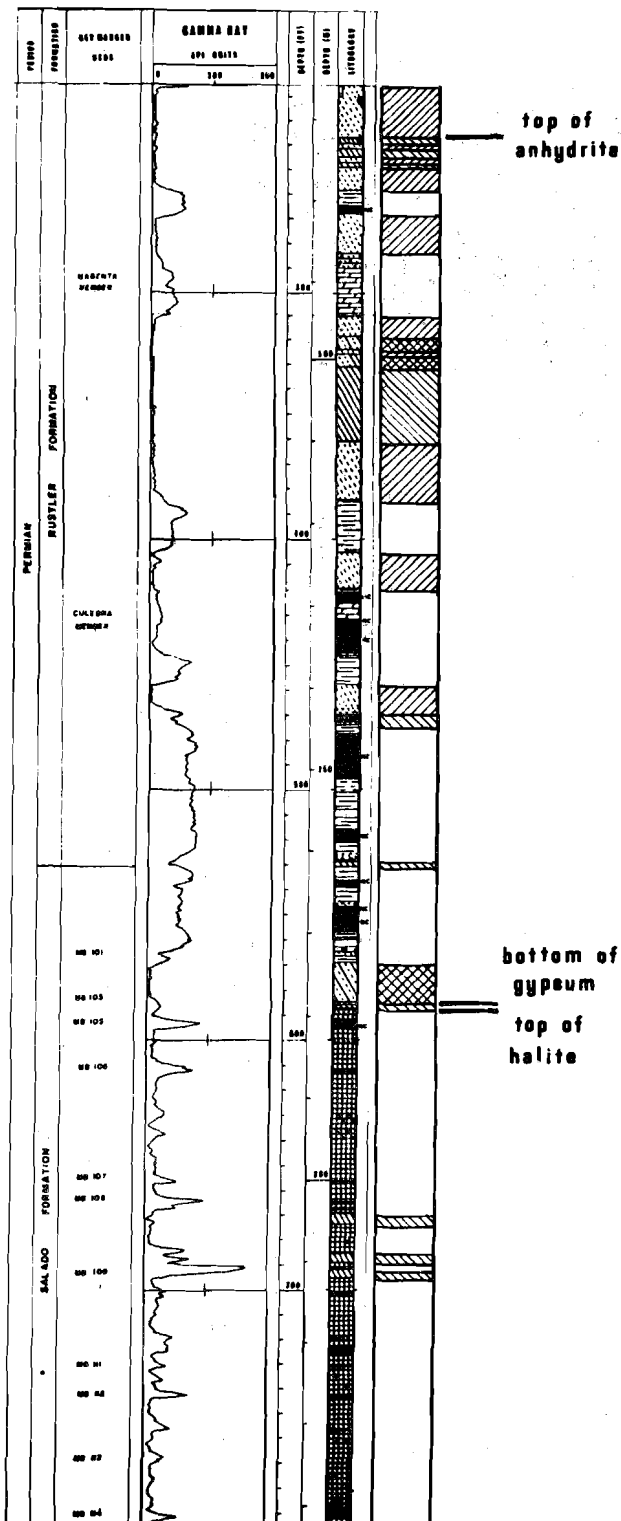


Figure VIII-3. Stratigraphic Section, WIPP 28, Showing Zones of Gypsification. (Upper-left to lower-right solid ha-chures indicate anhydrite)

Stratigraphic control points in the longitudinal section along Nash Draw (Figure VIII-2B) include three holes from the previous section (W25, 26, and 29). Along the longitudinal section, except in WIPP 29 (representing a very "mature" portion of Nash Draw, where near-surface dissolution effects are manifest as deep as the upper part of the McNutt potash zone), the "top of anhydrite" does not drop below the Magenta dolomite. Throughout the longitudinal section, except near Livingston Ridge, the top of halite coincides with the bottom of gypsum, both below the top of the 103 marker bed. This indicates that the interaction of rock and water was concentrated along these coincidental boundaries, and that the water was undersaturated enough to allow both removal of halite and hydration of anhydrite, but there is no gypsification of anhydrite beds within the underlying halite. In the very "mature" portion of southern Nash Draw (W29), there is no anhydrite above the top of halite. Thus, W29 may represent the closest approach to "brine aquifer" conditions, but the "brine aquifer" need not be ubiquitously, uniformly developed throughout Nash Draw (Robinson and Lang, 1938), or constrained to the Rustler/Salado contact.

Figure VIII-4 illustrates the interaction of rock and water within a competent rock layer, adjacent to which there is no evidence of such interaction. The core is a sample of marker bed 103 (anhydrite). Note the fractures having boundaries that can be "fit" back together again. The fractures are filled with halite, indicating the interaction of rock with water, but it did not involve gypsification or water sufficiently undersaturated to instigate gypsification.

The data of both geology and geochemistry in the Delaware Basin are interrelated. Because of affinity of the Rustler Formation waters with the meteoric field (Figure VIII-5), it can be concluded that their oxygen and hydrogen isotope compositions prove a definite meteoric origin for many of them. Some other waters, many from the Rustler, have isotopic compositions outside the worldwide meteoric field, but can be traced to having originally meteoric affinities. These results indicate an interaction of rock with water involving mutual exchange of both hydrogen and oxygen. In the stable isotope literature, the amount of perturbation of isotopic compositions (say $\delta^{18}\text{O}$) from local meteoric values (during interactions between rock and water) is inversely proportional to the rock-to-water ratio, as supported by the data in Figure VIII-6.

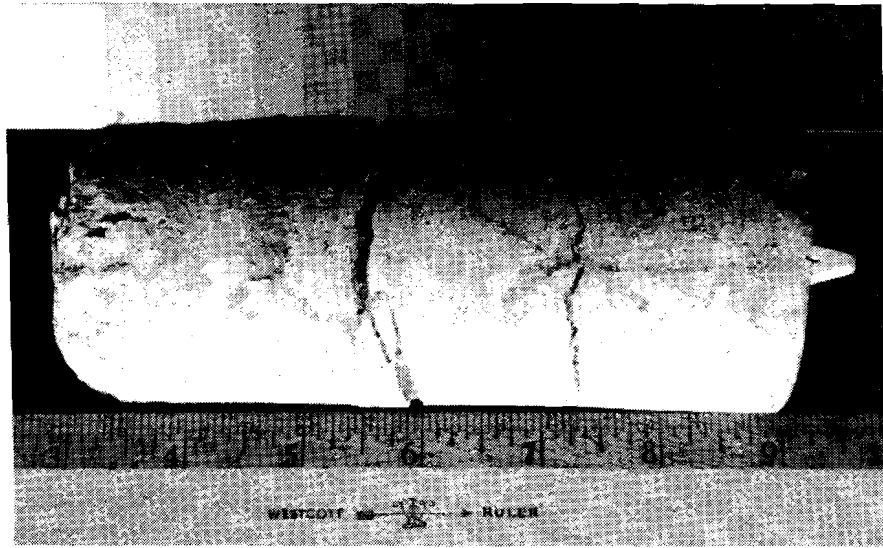


Figure VIII-4. Photograph, Core From Marker Bed 103 (WIPP 25) Containing Halite-Filled Fractures

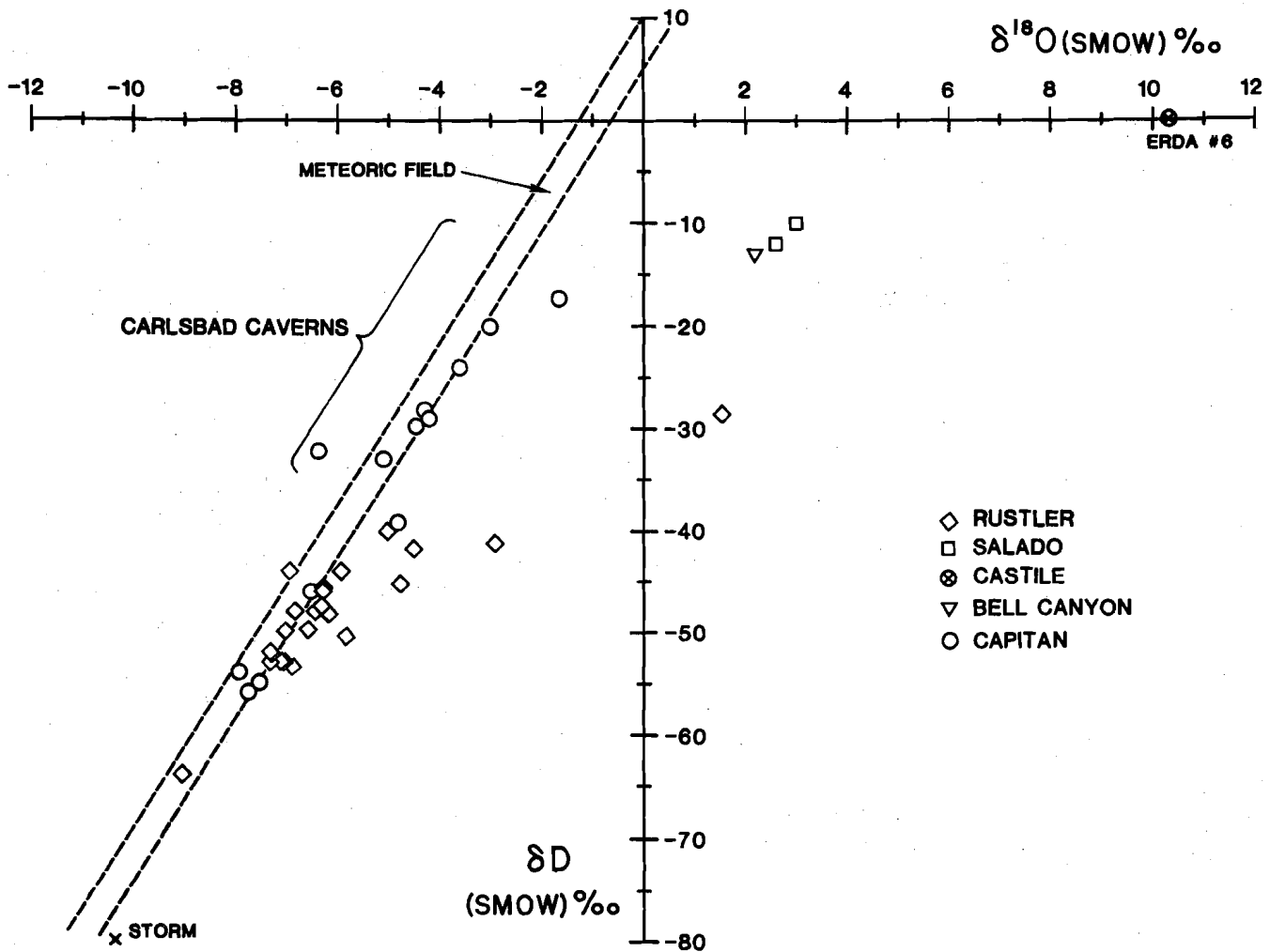


Figure VIII-5. Stable-Isotope Relationships for Some Delaware Basin Groundwaters (δD vs $\delta^{18}O$). "Meteoric field" is delineated based on the work of Epstein et al, 1965, 1970, and Craig, 1961.

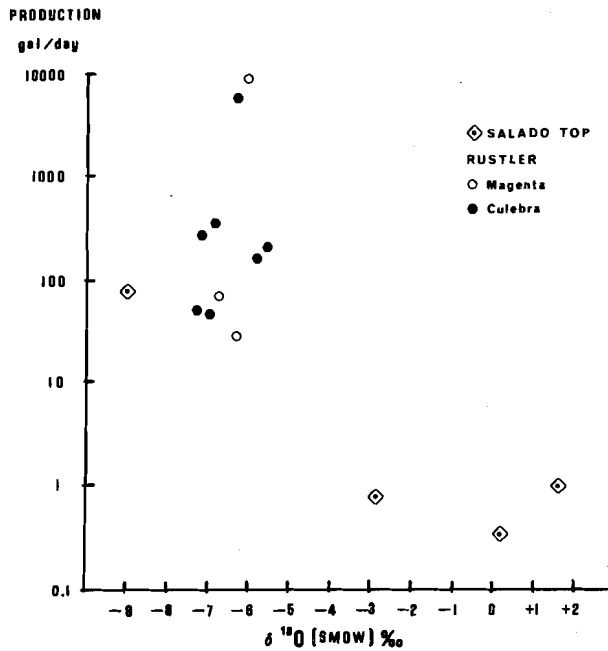


Figure VIII-6. Relationship Between Production Rate and $\delta^{18}\text{O}$ Value, Rustler Formation Waters

As seen in Figure VIII-6, there is a relationship between $\delta^{18}\text{O}$ value and production rate. It is not linear, but rather L-shaped. Some isotopic values (at higher production rates) are closely clustered at the expected local meteoric water values ($\delta^{18}\text{O} \sim -7\text{‰}$, $\delta\text{D} \sim -50\text{‰}$). Note the gap in the data, between production-rate values of 1 and 10 gal/d. At production rates less than the "gap" range, deviations of $\delta^{18}\text{O}$ values from meteoric values are significant. This relationship supports the argument that the oxygen isotope shift represents a reduced water-to-rock ratio. Also, note that above a certain production rate, in this case 1 to 10 gal/d, $\delta^{18}\text{O}$ is independent of production rate. Variations in production rates probably reflect actual variations in permeability, since the test holes providing the data were designed to make all hole conditions similar.

The oxygen isotope shift cannot be the result of fractionation by partial evaporation from a free water surface; the δD vs $\delta^{18}\text{O}$ trajectory slope of 3 for the Rustler waters (Figure VIII-5) is not sufficiently steep (Allison, 1982). Further, the isotopically "shifted" portion of the groundwater system is not exposed to the atmosphere any more than is the isotopically "meteoric" portion. The "diamonds" in Figure VIII-6 all represent water of the basal Rustler Formation, which was

previously identified with the "brine aquifer" as an active dissolution zone. The high productivity diamond in Figure VIII-6 is from P14, the farthest west in the Los Medaños series of holes. Note that this water has a meteoric $\delta^{18}\text{O}$ value as well as nontrivial production rate, indicating that the water-to-rock ratio is not so small as it is farther east. The water of P14 is probably more characteristic of the Nash Draw, rather than the Los Medaños hydrologic system (see Chapter V).

Figure VIII-7 shows a relationship between total dissolved solids (TDS) and $\delta^{18}\text{O}$ value. The Rustler waters are all meteoric, with no dependence of TDS on $\delta^{18}\text{O}$ value; saturation with NaCl was not achieved even within a factor of 3.

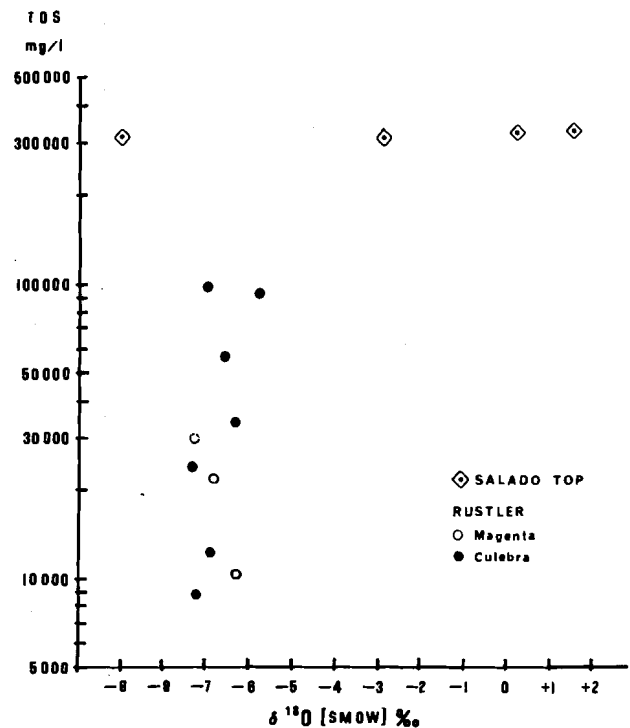


Figure VIII-7. Relationship Between Total Dissolved Solids and $\delta^{18}\text{O}$ Value, Rustler Formation Waters

Waters near the Rustler/Salado contact (diamonds) are all nearly saturated with NaCl, regardless of $\delta^{18}\text{O}$ value. Again, the left-most point represents the westernmost hole (P14) in the Los Medaños series. Its $\delta^{18}\text{O}$ value indicates no oxygen isotope shift away from a meteoric value of $< -5\text{‰}$, that water is abundant here; its TDS indicates a saturated solution. The other

saturated solutions are in much shorter supply, as described before, and show an oxygen isotope shift ($\delta^{18}\text{O} > -4\text{‰}$) away from the local meteoric range as a result of interaction in which the rock-to-water ratio was relatively large. The saturated solutions have no capacity for additional uptake of halite, and therefore have no further potential either for halite dissolution or for hydration of anhydrite to gypsum (see Chapter III). Because of its more abundant supply, the P14 water (having come from a meteoric supply) probably in the immediate past participated in the removal of halite. The Rustler waters, in spite of being meteoric, have limited access to halite, most likely adjacent to the Magenta and Culebra dolomites. Examination of cores reveals some removal of halite as well as some degree of gypsification near the dolomites in the Rustler.

From these discussions it can be inferred that only waters in abundant enough supply to retain meteoric stable isotope values participate in active halite dissolution. Other nonmeteoric waters in Guadalupian rocks (e.g., the Bell Canyon water in Figure VIII-5 and Table VIII-1) are undersaturated in solutes, but the nature of their solutes does not indicate simple uptake of soluble components of host and adjacent rocks. These waters contain abundant calcium chloride and sodium chloride, a feature that together with the isotope shift is characteristic of many oil-field waters that were shown to be stagnant and small in supply with respect to amount of rock. Thus, these waters were profoundly influenced by rock, rather than vice versa, and cannot be actively moving, continuously recharged dissolution brines.

Stratigraphic, mineralogical, and geochemical observations of dissolution products, both rock and water, allow formulation of an integrated hypothesis for the mechanism of dissolution of evaporites. First, dissolution of halite and gypsification of anhydrite in the Rustler and Salado Formations is taking place only where there is a continuous supply of meteorically derived water. Second, the necessary supply of water is carried in fractures in rocks more competent than rock salt, such as the Magenta and Culebra

dolomite members of the Rustler Formation and probably marker bed 103 of the Salado Formation. Excursions of water from the fractured brittle rocks into sub- and superjacent rock initiates dissolution above and below the water-bearing horizon, with resultant removal of halite. The subsequent collapse opens fractures in brittle rock, locally increasing the permeability. After some halite has been removed, polyhalite (in the lower Rustler and upper Salado) becomes exposed to the solutions, but few additional changes are expected to occur at a given locality until nearly all halite has been removed, so that salinity is no longer maintainable at the saturation concentration. As the salinity drops to a certain level (at present unknown), polyhalite begins to dissolve incongruently, leaving a residue of calcium sulfate (hydrated or not). As the solute content drops further, hydration of anhydrite to gypsum takes place, since the thermodynamic activity of water has locally risen to the appropriate level (see Chapter III). Finally, the residual gypsum is dissolved and carried away.

Except in places where erosion by solution-and-fill (Bachman, 1981) has created well-developed gypsum karst topography (complete with sinkholes), anhydrite beds are preserved sandwiched between gypsum layers. This attests to the laterally migrating nature of a water-bearing dissolution zone with limited upward or downward excursion from a stratabound zone, and there is no evidence of a single "dissolution surface" migrating either downward or upward. No profound effects of water moving through long vertical fractures were discovered, as in the case of "breccia pipes" and sinks (cf Anderson, 1978) in otherwise-undisturbed evaporites.

As results of laboratory studies of mineralogies and textures of the original and altered evaporite core become available, together with reliable characterization of solutes and isotopic constituents of waters recovered from Nash Draw, mineralogical alterations accompanying dissolution can be documented in more detail. Such documentation will be the subjects of additional studies, independent of this report.

TABLE VIII-1. Values of Solute Concentration for Some Groundwaters in the Delaware Basin (Concentration in mg/L)

Name	Unit	Sec-Twp S-Rge E Location	Ca ⁺²	Mg ⁺²	Na ⁺	K ⁺	HCO ₃ ⁻	SO ₄ ⁻²	Cl ⁻	Total	Na/Cl
Middleton	Capitan	21-19-32	1100	570	10600	330	300	3720	17000	33600	0.62
H3	Culebra	22-31-29	1500	670	19000	630	115	5700	29600	57200	0.64
P14	Culebra	24-22-30	3100	760	7600	600	357	1400	20000	33700	0.38
Duval Mine	Salado	Subsurface	640	55400	27500	30000	1090	3650	236500	355100	0.12
Miss. Chem. Mine	Salado	Subsurface	200	44200	43600	45800	not	12050	226200	372000	0.19
							measured				
ERDA 6	Castile	35-21-31	130	350	112000	5100	1310	16000	186100	321000	0.60
AEC 8	Bell Canyon	11-22-31	10000	2500	55000	860	420	240	120000	189000	0.46
AEC 7	Bell Canyon	31-21-32	9100	2350	57500	780	67	1280	114400	185500	0.50
ERDA 10*	Bell Canyon	34-23-30	5170	1410	84500	620	55	1770	149300	242800	0.56

*Values only relative; include probable contamination by drilling-mud filtrate

Delaware Basin Groundwaters

Connections were inferred between various occurrences of groundwaters in the Delaware Basin, generally in favor of certain models of dissolution (cf Anderson, 1981). Certain characteristics of the various waters, however, place constraints on the degree of connectedness of groundwaters. Some postulated connections are even disallowed by the geochemical data. The kinds of data relevant to establishing connections (or lack thereof) between groundwaters are (a) major (and, less commonly, minor) solutes, (b) natural deuterium and ^{18}O abundances (stable isotope values) in the water molecules themselves, and (c) the activity ratios of ^{234}U to ^{238}U .

The most complete survey of the geochemistry of Delaware Basin groundwaters to date is that by Lambert (1978). That work concluded that three kinds of interactions of rock with water were illustrated by brines (defined as solutions with $>30\,000$ mg/L total dissolved solids):

1. Simple uptake of solutes from host or adjacent rocks. This is the type of water expected to result from active dissolution of evaporites. The stable isotope values are those of local meteoric waters, indicating a continuously recharged surface source for the water molecules. Such waters lie within the "meteoric field" of Figure VIII-5 (Epstein et al, 1965, 1970; Craig, 1961). The major solutes (Na, Ca, Mg, K, SO_4 , and Cl) are in the same proportions as in common assemblages of evaporite minerals (dolomite, anhydrite, halite, kainite, carnallite, bischofite, sylvite, kieserite, to name a few).
2. Uptake of solutes *and* additional D or ^{18}O (or both) from host or adjacent rocks. This is the type of water that may have participated in active dissolution of evaporites at one time, but that became entrapped for long enough residence time in rock to allow changes in the water molecules themselves. The time required for isotopic exchange between minerals and water at low (say 30°C) temperatures is difficult to measure in laboratory studies, because such reactions are kinetically inhibited at low temperatures. "Accelerating aging" techniques are of no help, because such treatment alters the reaction mechanisms, making it impossible to

derive true rate constants for the conditions of interest. Further, laboratory kinetic studies of low-temperature isotopic exchange reactions cannot take account of such complex factors as natural catalysis by trace metals, natural diagenesis, microscopic surface effects, etc. Consequently, resort can be made to natural analogs, in which the reactions of interest took place over a long (geologic) time (Lambert and Epstein, 1980). That isotopic exchange takes place very slowly at low temperatures is demonstrable. Outcrops of igneous rocks maintain their igneous values of $\delta^{18}\text{O}$ in quartz, feldspar, and other minerals, in spite of surficial erosional exposure. Barr et al (1979) have shown that water in the Capitan has not come to isotopic equilibrium with marine limestone even after a residence time of 1.3 my, but water has preserved its meteoric value. In this classification, the mineral source of the additional $\delta^{18}\text{O}$ and D in the water was not precisely identified, due to lack of stable isotope studies of Delaware Basin rocks. The most likely source of exchangeable $\delta^{18}\text{O}$ and D, however, is clay minerals (Savin and Epstein, 1970).

3. Uptake of solutes *and* additional D or $\delta^{18}\text{O}$ (or both) *and* cation exchange between rock and water. This last kind of relationship indicates the most complex interaction of rock with water. The major solutes of these waters are in proportions *not* the same as in common evaporite minerals. By the method of calculating probable compounds (Collins, 1975) in the sequence of evaporite crystallization given by Braitsch (1971), the normative constituents of such waters can be such minerals as thenardite, tachyhydrite, and antarcticite. The stable isotope relationships of this class of brines are similar to those of class 2 above.

A distinction between class 1 (active dissolution) waters and the other two classes is readily apparent in Figure VIII-5. Waters whose $\delta^{18}\text{O}$ and δD values separate them from the meteoric field are not connected by unrestricted flow with a surface-recharged source. Thus, class 1 waters may include Rustler (Culebra) and Capitan (below the fresh/brine interface of Hiss, 1975) (Table VIII-1). Waters from the Salado, Castile, and Bell Canyon Formations cannot be participating in an actively moving groundwater system. That Salado (potash mine seep) waters may have at one time

originated from the surface (for example, by migration along bedding planes) is not impossible; there is, however, no direct evidence of connection between potash mine seeps and surface recharge. That they are extremely limited in volume is clearly demonstrated by the fact that mine seeps cease flowing in a matter of days to months. The stable isotope and solute relationships of Salado waters constrain them to class 2. Their solutes may be in equilibrium with their host (potash ore zone) rock, by virtue of various congruent and incongruent solubilities of host minerals. Their stable isotopes, however, are dominated not by meteoric water, but by the minerals in the rock, indicating an overwhelmingly large rock-to-water ratio prevalent in the reaction system. Whether the Salado water molecules are dominated by waters of crystallization in evaporite minerals (such as polyhalite) or by a source such as hydroxyl in clay minerals cannot be said at this time, owing to a lack of necessary mineralogical isotopic studies.

The Bell Canyon Formation was invoked by Anderson as both a major source (1978) and a major sink (1981) for waters dissolving evaporites. The geochemical data clearly show that neither can be the case. The Bell Canyon waters cannot fall into class 1, because of their nonmeteoric stable isotope values. In fact, there is indication from AEC 8 core that the formation water is in oxygen isotopic equilibrium with the calcite cement in the Bell Canyon (5% by weight) at a temperature of 130°F. The temperature measured in the Bell Canyon Formation in the AEC 8 and ERDA 10 boreholes is between 90°F and 128°F, in close agreement with the isotopically calculated temperature based on the temperature dependence of the $\delta^{18}\text{O}$ fractionation factor between calcite and water (O'Neil et al, 1969). Thus, the 5 wt% calcite, of marine origin (which we assume had an original $\delta^{18}\text{O}$ value of $\sim +30\text{‰}$ with respect to Standard Mean Ocean Water, SMOW), has reacted with water, resulting in the lowering of the marine $\delta^{18}\text{O}$ value to the observed value $+25.1\text{‰}$. If, similarly, we assume that the Bell Canyon water was of local meteoric origin ($\delta^{18}\text{O} \sim -7\text{‰}$), it was raised to $+2.2\text{‰}$ by reaction with the calcite to achieve isotopic equilibrium. By closed-system material balance calculation, the molar ratio of calcite to water in the Bell Canyon water-bearing zones would be at least 1.88, if the Bell Canyon water was of recent meteoric origin. The equivalent weight ratio of calcite to water is ~ 10.44 . Thus the amount of water in the water-bearing zones of the Bell Canyon Formation is indicated to be 0.5% by weight. This is

an absolute maximum, if the Bell Canyon calcite is inferred to be of marine origin, and all Bell Canyon water is inferred to be of modern meteoric origin.

It would be valuable to obtain uranium isotope ratios along the potentiometric gradient of the Bell Canyon Formation. Such measurements could be interpreted as rates of water movement (if any) and serve as indicators of rates of low-temperature isotopic exchange. Adequate sampling at such depths is costly and must be meticulous. The sampling and analysis was not done to the extent required for determining the rates of water movement in the Bell Canyon Formation.

Waters from the Bell Canyon Formation cannot fall into class 2, the "dead-ended" dissolution waters, by virtue of their major solute relationships. Three widely separated samples of Bell Canyon water (Figure VII-1) were analyzed (Table VIII-1). Calculating probable compounds (Collins, 1975) shows too much calcium to be contained entirely in such common minerals as calcite, dolomite, and anhydrite. Similarly, Bell Canyon water contains too much chloride to be contained entirely in halite. Thus, Bell Canyon waters are said to be "calcium chloride normative" (Lambert, 1978). Calcium chloride (equivalent to the mineral antarcticite, $\text{CaCl}_2 \cdot 6\text{H}_2\text{O}$) is not found as a congruently soluble evaporite mineral or as a product of incongruent dissolution of any evaporite mineral known to occur in the Delaware Basin. Thus, Bell Canyon waters are not solutions resulting from simple dissolution of evaporites.

Neither the proportions of solutes nor the stable isotope values of Bell Canyon waters support Anderson's hypothesis (1981) of "down-gradient movement of brine into conduits communicating with the basin aquifer." Further, the higher potentiometric surface of the Bell Canyon precludes flow from the Capitan through the evaporites into the Bell Canyon (see Chapter V). Thus, the Bell Canyon is not a sink for dissolution-derived brine. The degree to which the Bell Canyon may have acted as a source for water to dissolve evaporites is similarly limited. The high-calcium low-sulfate Bell Canyon solutions are not chemically compatible with high-sulfate environments of the Salado and lower Castile. To bring the two in contact would result in an instant precipitate of gypsum, with little or no additional dissolution of halite, owing to the already high NaCl content of Bell Canyon waters. In addition, undersaturation of Bell Canyon waters with respect to NaCl (compare chloride levels in Salado waters; Table VIII-1) does not

make them likely candidates for waters that percolated downward from the Castile or Salado.

Possible origins of waters rich in calcium chloride were addressed by Graf et al (1966), Lambert (1978), and Graf (1982). The physical processes by which Bell Canyon waters could serve as agents of dissolution of superjacent evaporites are the subject of calculations by Wood et al (1982).

Recently there has been a great deal of interest in "brine reservoirs" in the Castile. The work of Lambert (1978) has placed the ERDA 6 water (Table VIII-1) in class 3. The work of Barr et al (1979) has indicated a minimum age of isolation of the water from the nearest large potential source (Capitan) of 800 000 yr by the uranium-isotope disequilibrium method. Calculating probable compounds yielded normative sodium sulfate (as thenardite) in the ERDA 6 water, next in abundance to halite. Similar results were obtained for the WIPP 12 water. These waters' stable isotope compositions are unique, and demonstrate isolation from any active source of recharge from either the Capitan or Bell Canyon. The solutes in the Castile

water also make it incompatible with the water from the Bell Canyon; a high-sulfate (Castile) water and a high-calcium (Bell Canyon) water cannot be in direct connection without the precipitation of gypsum. The origin of excess sodium sulfate in Castile waters has not been uniquely determined. Sodium could have been enriched in the water at the expense of magnesium by cation exchange, or the sodium sulfate in the water may have arisen through incongruent dissolution of glauberite in the Castile. These questions cannot be answered by water analyses alone, but require extensive core studies as well. Additional information bearing on the origin and age of brine occurrences at ERDA 6 and WIPP 12 will be presented as data and interpretations become available.

Waters in the north-central portion of the Delaware Basin that are possibly active in dissolution include those in the Rustler and Capitan. Waters indicated not to be active in dissolution in this area are those now found in the Salado ("mine seeps"), Castile ("brine reservoirs"), and Bell Canyon Formations.

Chapter IX. Time-Dependence of Dissolution

Introduction

As indicated in previous discussions, dissolution of Ochoan evaporites may have taken place as early as the Ochoan. Evidence for this is manifest in such features as (a) the erosional unconformity that truncates Halite III in the Castile Formation, and (b) the local unconformity proposed by some for the Salado/Rustler boundary. For the Delaware Basin as a whole, two divergent schools of thought have emerged regarding timing of major evaporite dissolution. The Bachman view (1974, 1976, 1980) invokes several episodes of dissolution since Triassic time, each dominated by greater degrees of evaporite exhumation and wetter climate, interspersed with episodes of evaporite burial and/or drier climate. The Anderson (1978, 1981) view is that most of the dissolution results from a continuing process spawned by exhumation and the hydrologic conditions that existed in and since the late Cenozoic.

With either view in mind it is still not possible to calculate an average rate of halite removal from the Delaware Basin since the Permian. With the Bachman view, it is possible to determine a rate based on the age of certain geomorphological features, but only during the episode of dissolution defined by discrete time markers. The nature of the Anderson view (deep-seated undermining of the halite) does not allow for developing surficial expression of progressive dissolution, except at isolated points, all said to be post-Gatúña (Chapter VI). Anderson's surficial evidence for deep-seated dissolution consists of exhumed remnants of old processes, mostly from the erosionally exposed western and southwestern parts of the Delaware Basin. These features include dissolution breccias and castiles. Decreases in the halite/anhydrite ratio of certain zones between marker beds are taken by Anderson as evidence of deep-seated (undermining) dissolution. A more detailed direct evaluation of the Anderson hypothesis of lower Salado halite removal is limited by (a) the expense of coring the appropriate lower Salado intervals especially to look

for deep-seated dissolution breccias said to be equivalent to exposures of breccias at the western basin margin, (b) the restrictions on boreholes penetrating the evaporites in the known potash resource area (KPRA), an area that may be critical to the evaluation of the hypothesis (Cheeseman, 1978), (c) no economic motivation for potash industry holes to penetrate as deep as the Castile, and (d) no economic motivation for precompletion geophysical logging in the Ochoan in hydrocarbon industry holes, whose targets are typically Guadalupian and older, deeper rocks.

Anderson's chief objection to an episodic model of "deep-seated" dissolution through geologic time is that the deep-seated undermining does not necessarily leave discrete geomorphic time-markers as does shallow-seated (e.g., Nash Draw) dissolution. There is no prominent topographic expression of "wedging out" of the lower Salado. Generally, a scarp forms along the surface projection of the division between collapsed residue with overburden and undissolved evaporites (see Gustavson et al, 1980, and Johnson, 1981, for descriptions of dissolution-related phenomena in Texas and Oklahoma portions of the Permian Basin). Scarp formation appears to be insensitive to depth of halite removal, and appears to keep pace with halite removal, but erosion could conceivably keep pace to obliterate scarps as they form. Dissolution under Nash Draw has left such a scarp: Livingston Ridge (Figure VI-1). The absence of such a scarp weakens Anderson's hypothesis considerably, especially since Gustavson et al (1980) and Johnson (1981) have shown that salt-bed dissolution results in relatively abrupt termination, not regional thinning of the beds, and that scarp erosion is not expected to be rapid. If erosion destroys such a scarp or prevents it from forming, the dissolution rate is reasonably slow, and the erosion rate can be used to calculate threat on the time scale appropriate to WIPP. There is also evidence that large-scale halite dissolution leaving porosity in a dissolution breccia is followed closely by

gypsification of residual anhydrite (Chapter VIII). This relationship does not appear to be present in the east-central Delaware Basin in the upper Castile and lower Salado.

How much halite has actually been removed from the Ochoan evaporites has of itself little relevance to the long-term integrity of halite beds selected for storage of radioactive waste. Dissolution of most of the Ochoan halite since Gatuña time (600 000 yr ago) would imply a more severe threat to the survivability of the remaining halite than if dissolution of the same volume of halite had taken place over, say, 150 My. Anderson (1978) cites occurrences of collapsed Gatuña in various surficial features, but pays little heed to the significance of undisturbed Gatuña occurrences in the western exhumed portions of the Basin where there is evidence of pre-Gatuña post-Permian dissolution of halite.

"Major dissolution occurs while land masses are above sea level," according to Bachman (1980). In addition to local dissolution in the Castile before Salado time (Adams, 1944), there is evidence for dissolution after Salado and before Rustler time along the western part of the Basin (Adams, 1944). There is also evidence that the Delaware Basin was above sea level throughout the Triassic, Jurassic, Tertiary, and Quaternary. Not all of these times are separable in the geologic record of southeastern New Mexico.

Dissolution Episodes

Triassic

The upper Triassic Dockum Group is represented in southeastern New Mexico as stream deposits, whose depositional wedge-edge is near the present westernmost Dockum outcrops (McGowen et al, 1979). Triassic rocks resting on the Culebra member of the Rustler Formation in the Pecos Valley (with no intervening Magenta member) indicate that the Ochoan was exhumed in the western part of the Basin and eroded in late Permian or early or middle Triassic time (Bachman, 1980). Bachman also points out that occurrences of Dockum lapping across erosionally truncated Rustler rocks coincide with absences of halite in the subsurface, indicating that dissolution may have formed the dissolution breccias of Anderson et al (1978) as early as Triassic.

Jurassic

Geologic evidence throughout southern New Mexico indicates that erosion was dominant in the Jurassic. The most compelling evidence for this consists of occurrences of thinned Triassic or absent Triassic rocks, with nonbrecciated Cretaceous marine rocks resting on Permian. In other places, Cretaceous rocks lap across the wedgeout of Triassic rocks resting on the truncated Permian rocks (Bachman, 1976). The presence of Triassic rocks in the eastern part of the Basin limits the extent of eastward-progressing erosion of Permian rocks in the Basin, which when exposed to erosion would presumably be exposed to dissolution as in the modern analog of Nash Draw (see Chapter VIII).

The mere occurrence of Cretaceous rocks in sinkholes can probably not be used as reliable indicators of pre-Cenozoic dissolution (Anderson, 1981). Unless the younger rock (Cretaceous) can be shown as deposited in a sinkhole in Permian rock, the mere presence of (perhaps) jumbled Cretaceous rock does not place a lower limit on the age of collapse.

Tertiary (?)

In eastern Lea County, structure of the redbed surface (Triassic?) shows deposition of the Ogallala Formation on an uneven surface pockmarked with depressions (Nicholson and Clebsch, 1961). Many of the depressions have no topographic expression at the top of Ogallala or younger rocks. Since the Ogallala "was deposited on an irregular erosional surface as a series of complex alluvial fans" (Bachman, 1976), and inferred regular stream flow indicated initially moister conditions than at present, it is not unreasonable that some dissolution took place in the erosional period of the pre-Ogallala Tertiary. Perhaps some of these pre-Ogallala surface depressions are expressions of evaporite removal and subsequent collapse in (or before) the Tertiary.

Pleistocene

The regionally persistent Mescalero caliche, together with conglomerates and sandstones of the Gatuña Formation, provide stratigraphic markers for a relative chronology of surface-expressed karstic features indicative of dissolution during the Pleistocene. Subsurface dissolution and surface collapse both preceded and followed the formation of Mescalero caliche

at Nash Draw (see previous chapter) and Clayton Basin (Figure VI-1). The collapse feature at Crow Flats, however, preserves evidence for at least three episodes of dissolution and collapse: (1) after Triassic and before Gatuña time, (2) during or after Gatuña time, and (3) after Mescalero time (Bachman, 1976). At Crow Flats, indurated Triassic rocks have collapsed into sinkholes 50 ft deep in Rustler gypsum, and are unconformably overlain by the Gatuña Formation. There is not enough information to identify the timing of this episode of solution and collapse more precisely than post-Triassic and pre-Gatuña (> 600 000 yr). At some places at Crow Flats, sinkholes containing collapsed Gatuña are capped by relatively undisturbed Mescalero caliche. Elsewhere flat-lying Mescalero caliche truncates dipping beds of Gatuña. Displacements of the caliche in collapse features are as much as 110 ft at Crow Flats, 100 ft in Clayton Basin, and 180 ft in Nash Draw since Mescalero time.

Various estimates of middle Pleistocene climatic conditions have indicated that Gatuña time was more moist than the Holocene. Some workers (Bachman, 1980; Anderson, 1931) have reasoned that wetter climate accelerated the instantaneous rate of dissolution. One point of view (Anderson's) contends that virtually all dissolution of halite in the Ochoan section occurred since the Pleistocene [*sic*] uplift of the Guadalupe Mountains and east-northeastward tilting of the Delaware Basin, and that the maximum rate of dissolution, first established around Gatuña time, remains unabated. The Bachman point of view contends

that dissolution was episodic during the past 225 my, as a function of regional base level, climate, and overburden. Virtually all agree that it is unrealistic to apply a calculated average rate of dissolution, determined over 500 000 yr, to shorter periods, much less to extrapolate such a rate into the geologic future. Because of the impossibility of accurate identification of an isochronous "dissolution front" from existing subsurface datum points, and based on previous (now invalid) assumptions that all dissolution took place in the Quaternary, even previous rate estimates (Bachman et al, 1973; Bachman, 1974) are too high.

The occurrence of Gatuña sediments in the depressions described by Maley and Huffington (1953) cannot be used to place maximum age limits either on the onset of dissolution in the Delaware Basin or on the uplift of the Guadalupe Mountains and resultant tilting of the Delaware Basin. The tilting has not been precisely dated, but only estimated by Hayes (1964) as "late Cenozoic," in the total absence of geologic time markers. The occurrence of Gatuña at the top, exposed portions of the fill in the depressions establishes the minimum age of the fill as Gatuña-time. Neither faunal nor radiometric determinations (nor even lithologic characterization and correlation) have been performed to allow application of a maximum age to the base of the fill. Thus, it is not reasonable to constrain the entire accumulation of fill to any particular time interval. Thus Anderson's (1981) constraint of all dissolution in the Delaware Basin to "late Cenozoic" is, in view of the limitations imposed by the available data, unwarranted.

Chapter X. The Plausibility of a Dissolution Hypothesis

The Factors

A hypothetical flat-lying infinite tabular body of evaporites overlain and underlain by at least 300 m of rock with minimal vertical permeability will probably never dissolve. In the Delaware Basin several criteria for recognizing evaporite dissolution were met, according to geological observations (cf Johnson, 1981). Several factors must be operable in any model to account for observed dissolution:

- Trigger** There must be some perturbation in the structural, hydraulic, or stratigraphic relationships that have preserved the evaporites intact, so as to initiate conditions favorable to dissolution.
- Path** A throughgoing permeable conduit must be actively maintained to allow less concentrated solutions into contact with soluble rock and to allow more concentrated solutions to escape.
- Continuity** A body of soluble mineralogy must have some connected lateral or vertical extent, rather than be disseminated as isolated clots through rock of low connected porosity, or dissolution is quickly self-consuming and self-terminating.
- Source** A supply of solution undersaturated in soluble rock components must be brought through the path.
- Sink** Solution more concentrated in soluble rock components than source solution must escape, so as not to accumulate and stifle the dissolution process.

This exposition of the dissolution process in the Delaware Basin ends much as it began, with a review of the four proposed mechanisms of dissolution: solution-and-fill, phreatic dissolution, stratabound dissolution, and brine density flow. The application of each of the mechanisms is evaluated next for the plausibility of its application to explain various features in the

Delaware Basin described in this work. Each evaluation is made with an emphasis on the geologic evidence for effectiveness of each of the five factors governing dissolution: trigger, path, continuity, source, and sink.

Dissolution Mechanisms

Solution-and-Fill

Erosion by solution-and-fill is the easiest process to evaluate; all its factors are surface-controlled. Its trigger is erosional exposure of gypsum terrain, in which it is best developed. In the Delaware Basin the trigger was east-northeastward tilting of the evaporites, exposing edges of evaporite beds to weathering. The path for moving water is the interface between atmosphere and rock, which becomes extended through etching, giving rise to open channels, collapse, and gypsum karst topography. Continuity is determined by degree of areal subaerial exposure of soluble rock, and appears to be ensured on the observable outcrops of gypsum. Source is determined by rainfall and shallow infiltration. Sink is governed by patterns of surface and shallow subsurface drainage. Because the functions of the five factors can be verified by near-surface observations, it is not subject to serious doubt that solution-and-fill has been active in the formation of collapse structures in Nash Draw, Clayton Basin, Crow Flats, and probably the Gypsum Plain. It may legitimately be asked, then, why this mechanism is subject to evaluation, and what is its relation to halite removal, if it is most commonly developed in gypsum terrain. The importance of solution-and-fill lies in its role as a trigger for another mechanism, stratabound dissolution. The depth to which solution-and-fill is active (i.e., whether or not it may affect halite at depths of 100 to 300 m) depends

upon the degree to which associated rocks can maintain open channels to the surface. Rock salt itself cannot efficiently maintain open space in itself (Powers et al, 1978), and it has been shown that dissolution along bedding planes (a special case of stratabound dissolution) has been more efficient in removal of halite and in gypsification of anhydrite beneath (for example) the zone of direct surface influence in Nash Draw (Chapter VIII).

Phreatic Dissolution

Removal of halite by dissolution in open chambers resulting in larger chambers (Anderson and Kirkland, 1980) is a process difficult to reconcile with existing geological data. The trigger is postulated to be a fracture. The path must be either the triggering fracture or another fracture, since rock salt does not possess an inherent permeability large enough to carry the required amounts of water for the required length of time (5×10^{-8} D at 2000 psi confining pressure for a natural crystalline aggregate, 10^{-12} D for a single crystal; Powers et al, 1978). The inability of rock salt to maintain an open space even the size of a mine drift is demonstrated by the following extract from Brokaw et al (1972):

“During final mining most of the pillar support is removed and the worked-out areas gradually subside or cave . . . In those mines where mining has been completed, the subsidence is nearly 100 percent of the mined height. In most caved areas the subsidence is reflected in the overlying surface by the development of gentle depressions . . . Subsurface observations of the subsiding areas indicate that the rocks above the mined areas move slowly downward as a simple cohesive block after final mining is completed . . . The unique physical properties of the salt prohibit the development of open fractures. The fractures formed are tight or quickly sealed by flowage or recrystallization of the salt. The absence of water seepage into the mines in the areas of subsidence is indicative of the self-sealing character of the salt beds.”

Phreatic dissolution does not appear to have the requisite path development in rock salt, and thus this process is not efficient in rock salt.

Phreatic dissolution is more successful in brittle rocks such as limestone (e.g., Carlsbad Caverns). Collapse breached the insoluble Tansill dolomite by breakdown over a phreatic (or vadose) dissolution chamber in the underlying Capitan, thus creating the sinkhole known as the Natural Entrance (see Jagnow, 1979). Breakdown in the Fletcher anhydrite over a phreatic cavity in the Capitan, with collapse of overburden in a cylindrical body, is to date the most internally consistent process to account for the origin of Vine's Domes A and C (Chapter VI; Snyder and Gard, 1982). Phreatic dissolution in the Capitan has the required trigger (brittle fracture), path (again, brittle fracture), continuity of a relatively pure limestone body (with an average thickness of 2000 ft, occurring in an arcuate band 6 to 12 mi wide surrounding the Delaware Basin), source, and sink (water moving through old dissolution channels in the limestone with varying amounts of calcium carbonate in solution).

Because of its position above the Capitan, subsidence in San Simon Sink might also be attributable to collapse into a phreatic cavity in the Capitan. That the Sink's origin may not be an initial consequence of halite removal is as consistent with available geologic data as is Anderson's (1978, 1981) inference of unspecified faulting near the basin margin followed by an upward-stoping chimney growth by brine density flow. Thus, if a modern example of an actively forming breccia chimney is sought, San Simon Sink appears to be a more likely target of study than either Bell Lake or Slick Sink. San Simon Sink has the element of catastrophic collapse, consistent with failure of the roof of an underlying phreatic cavity in brittle rock, analogous to the breccia pipes (Chapter VI).

Brine Density Flow

Regardless of the demonstration using laboratory apparatus conducted by Anderson and Kirkland (1980) to illustrate dissolution by brine density flow, there are few features in the Delaware Basin attributable to this process alone. Its more important possible application is as a triggering agent or sink for other mechanisms, such as stratabound dissolution (see below).

Some details of the dissolution mechanism proposed by Anderson (1981) to support his hypothesis of dissolution in the Delaware Basin are summarized as follows:

"... Moving, undersaturated waters have somehow found access to salt deep within the evaporites and selectively dissolved large quantities of salt from beneath overlying salt beds... [p 142] The mechanism suggested here to account for the selective removal of salt at the lower Salado horizon is the down-dip movement of undersaturated surface waters to a point of contact with salt and the down-gradient movement of brine into conduits communicating with the basin aquifer... [p 143] For some reason, the undersaturated surface waters moving downward through fractures and solution channels in the exposed gypsum up dip and to the west found an easier path of communication in their down-dip movement at the lower Salado horizon. A pre-Salado dissolution residue [marking the erosional unconformity on Castile Halite III] may have predisposed this horizon through increased permeability or, as observed by Hills (1979), the Salado is especially vulnerable to down-dip dissolution owing to the presence of numerous seams and beds of clay and sands [p 143]."

In the foregoing excerpts, the source is surface water, and the sink in this model is the Bell Canyon Formation. The path is, at various points, either the bedding planes in the Salado on the near-source end, or conduits into the Bell Canyon on the near-sink end. Trigger is not specifically identifiable, but Anderson (1981) suggests that the "Cenozoic-filled depressions" (Maley and Huffington, 1953), structure in the Castile (Anderson and Powers, 1978), and faults of unspecified orientation (Haigler, 1962) provided the necessary perturbations to the evaporite sequence during the Cenozoic tilting of the basin, so that the evaporites fell victim to the surrounding hydrologic regime.

The flaw in this model of dissolution is the identification of the Bell Canyon Formation as a sink. It has been shown (Lambert, 1978; also Chapter VIII) that the solute relationships in the very small total amount of Bell Canyon waters (Chapter V) cannot have arisen from simple evaporite dissolution. An extract from Johnson (1981) illustrates this difficulty:

"The Na/Cl [weight] ratio of brines formed by dissolution of salt in western Oklahoma is remarkably close to 0.64, regardless of whether the water is a low-salinity or a saturated brine (Leonard and Ward, 1962).

This is because salt (and very little else) is being dissolved from the nearby salt deposits, and the combining ratio of Na and Cl in pure crystals of halite is 0.65. Oil-field brines consistently have Na/Cl ratios of 0.55 or less, and the ratio decreases well below 0.50 as the salinity increases."

Bell Canyon waters (Chapter VIII) from AEC 7, AEC 8, and ERDA 10 (this last probably contaminated somewhat with drilling-mud filtrate) have Na/Cl ratios of 0.50, 0.46, and 0.56, respectively (Table VIII-1). Thus, we see that the Bell Canyon waters have closer affinity with oil-field brines than with dissolution brines, and the Bell Canyon Formation yields definitive evidence that it has not been a sink for dissolution brines. These measurements are all down-gradient from the postulated "leading edge of an eastward-advancing deep-seated dissolution process," where Anderson (1981) said the product brines would accumulate.

The presence of faults or fractures is not always a trigger. Faults within the Castile Formation have been inferred on the basis of seismic reflection studies of the WIPP site (Borns et al, 1983). These faults may connect the Bell Canyon and Salado Formations, yet no removal of Salado halite is evident beneath the WIPP site. The closest suggested phenomenon that might be related to halite thinning is the "124 marker bed low" 2 mi north of ERDA 9 (Powers et al, 1978). The feature is in the "disturbed zone" and is discussed in more detail by Borns et al (1983).

Stratabound Dissolution

How, then, does regional dissolution in the Delaware Basin bedded evaporites take place? The question is answered here by another working hypothesis that is consistent with all the geological observations.

If dissolution is to have its continuity requirement satisfied, dissolution must be constrained to discrete beds of halite, as was shown to be the case in Nash Draw. It is largely for this reason that stratabound dissolution is probably the most efficient mechanism for removing bedded halite. The chief difference between the stratabound model proposed here and the subset of the stratabound model proposed by Anderson (1981) is in the path and sink factors. The source factor is shared by all dissolution models: meteoric water. In the Delaware Basin and surroundings, the progressive decline of evaporite abundance from east to west is correlative with erosional exhumation of progressively deeper stratigraphic horizons, because

of the east-northeast tilting. Thus, the triggering agent for stratabound dissolution is updip surficial exposure of lateral equivalents to evaporite beds. We have seen that such erosional exposure has taken place many times since the Ochoan (Chapter IX). The path proposed for stratabound dissolution is the only rock type that can demonstrably preserve open space in the midst of subsurface rock salt: the anhydrite interbeds. The stratabound hypothesis also asserts that dissolution is self-perpetuating. New fractures would be developed in the anhydrite interbeds at the advancing local dissolution front, as halite removal leaves open space above and below the water-carrying fractured anhydrite interbed. Collapse into the open space below the anhydrite would create additional fractures in the anhydrite. Collapse above and below the anhydrite creates a fracture zone of increased permeability, and opens additional access channels to deliver fresher water to halite and remove the dissolution brine. Anhydrite is converted to gypsum, and a surface scarp may or may not be formed to mark the collapse.

There remains the most difficult factor to identify: sink. As was said before, there is insufficient volume and flow-rate capacity in the Bell Canyon Formation for it to serve as a sink for all the dissolution brine generated (if indeed it is now being generated to the extent proposed by Anderson, 1981). Further, even if it had the capacity, there is evidence that the Bell Canyon Formation has not served as such a sink. Another sink of sufficiently low hydraulic potential is required. The lack of a good candidate sink is a major reason why *all* schemes for dissolution at depth (that are postulated to be presently very active) must be considered suspect.

One should not disregard the possibility that the "Cenozoic-filled" depressions (Maley and Huffington, 1953) may function in part as a trigger or part of the path factor in dissolution, rather than wholly as an indicator of dissolution. It is likely that these depressions (as well as such features as San Simon Swale) are in part erosional features (suggested by the thinning of Dewey Lake or Triassic as well as older Permian evaporites in the depressions; Figure X-1). These localized permeable structures are at least as deep as 1800 ft as indicated by present depth of fill (Figure X-1), and in pre-Gatuña times may have been deeper. That they intersected evaporites at depth through a combination of deflation and erosion by solution-and-fill seems likely, as indicated by the structural depression of the Rustler in the areas of deep fill (Maley and Huffington, 1953, plate 3). A chain of debris-filled depressions leading southward and draining the dissolution brine from the nearby fractured anhydrite beds

that they intersect would adequately fulfill the requirements for a sink. The problems remain (a) to demonstrate such movement of water in fractured anhydrite beds, aside from Nash Draw, and (b) to find the outfall of dissolution brine in the southern Delaware Basin, as salt springs or seeps in west Texas, analogous to the occurrences at Malaga Bend.

Figure X-2 diagrammatically summarizes the elements of a plausible stratabound dissolution model for the Delaware Basin. The stratabound model functions as follows:

1. Rainwater infiltrates an exposed bed of gypsum, through channels (Olive, 1957) or residue of solution-and-fill (Lee, 1925).
2. Water migrates preferentially along fractures in gypsum (or anhydrite), making outward excursions into adjacent strata, which may be more or less permeable.
3. Where rocks adjacent to fractured water-carrying brittle rock are halite-bearing, they dissolve, resulting in collapse, new fractures in brittle rock, local enhancements in permeability, scarp formation at the surface, and gypsification.
4. Dissolution brine escapes through local "low" in potential level, leaving an abrupt termination of the halite beds, not gradual thinning.

The halite and anhydrite beds diagrammatically depicted in Figure X-2 cannot be arbitrarily identified as any particular evaporite beds in the Ochoan section, such as specific units of the Castile or Salado Formation. The processes depicted in Figure X-2 probably do represent the elements of dissolution in the Rustler and upper Salado Formations in Nash Draw, since geological evidence of dissolution in rocks and waters has been found in Nash Draw. The stratabound dissolution process similarly cannot be applied arbitrarily to units in the Salado and Castile Formations in the absence of direct geological evidence of dissolution (residues and brines). The only direct evidence of active stratabound dissolution is in the lower Rustler and upper Salado Formations in Nash Draw.

Various aspects of the stratabound working hypothesis for dissolution of evaporites in the Delaware Basin are testable. As yet, none of its aspects are inconsistent with available geologic data. The "Cenozoic-filled" depressions could simply be pre-Gatuña deeply developed equivalents of Nash Draw, whose ultimate configuration may at some time in the geologic future resemble the Balmorhea-Loving Trough (Hiss, 1975) just south of Nash Draw. Methods of testing the hypothesis are presented in Chapter XI.

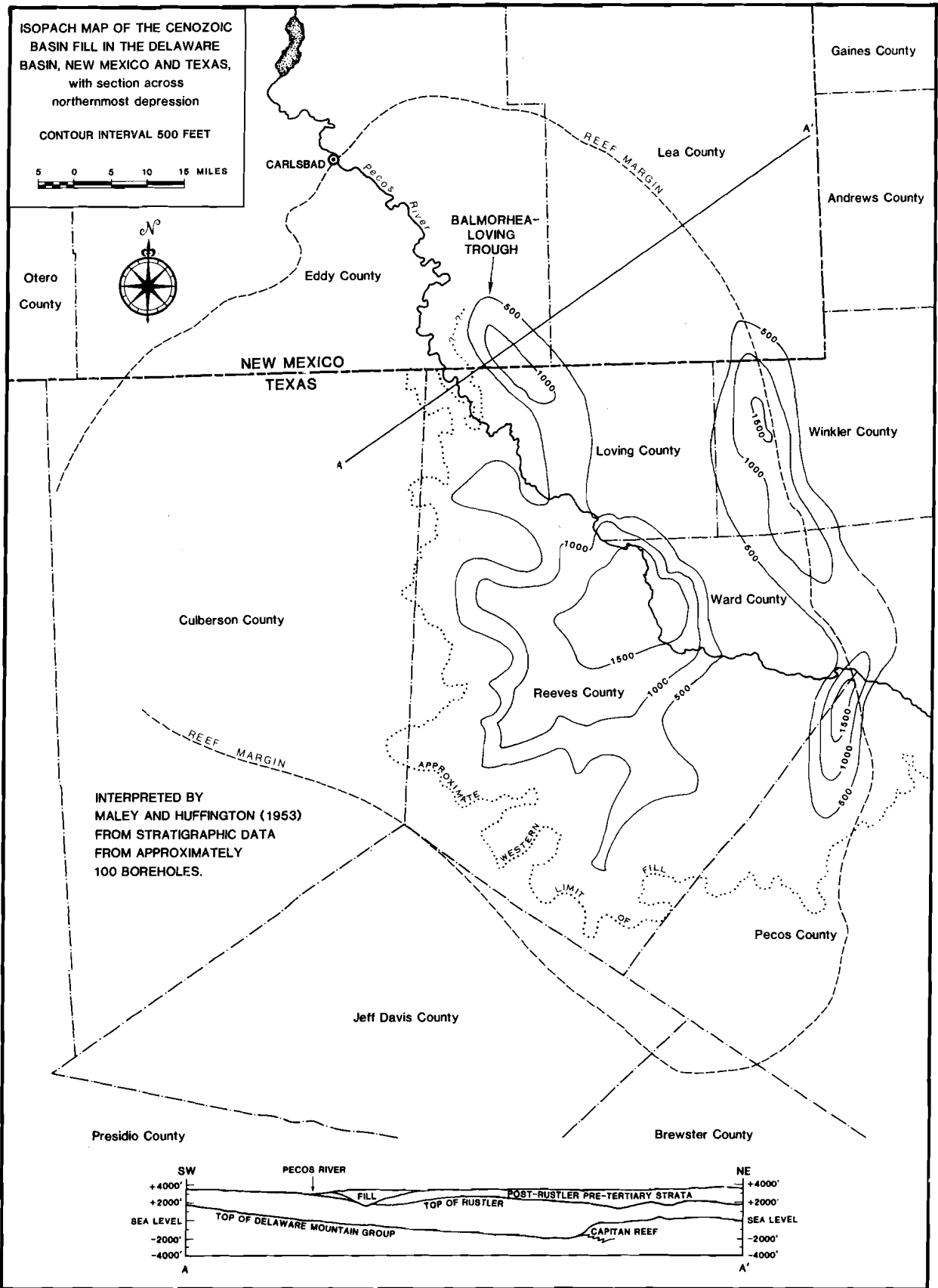


Figure X-1. Distribution, Depth, and Structure of Cenozoic-Filled Depressions in the Central Delaware Basin (after Maley and Huffington, 1953).

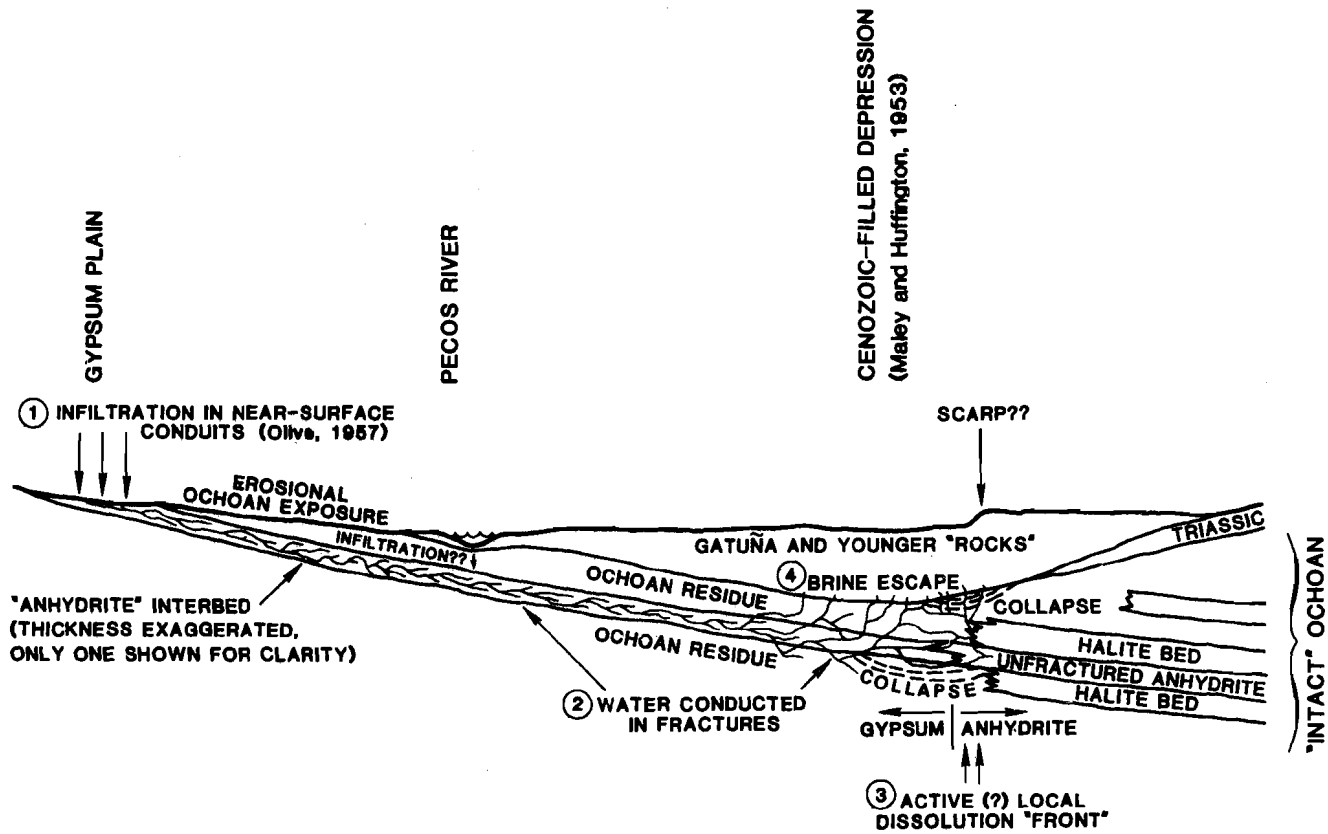


Figure X-2. Diagrammatic Illustration of a Hypothetical Process Leading to Stratabound Dissolution of Evaporites. Circled digits refer to sequential steps described in the text.

Chapter XI. Conclusions, Implications, and Recommendations

Conclusions

It is unlikely that the original extent of evaporites in the Delaware Basin at the end of Ochoan time was as large as recent estimates. Geological evidence suggests nondeposition of halite in parts of the Castile, particularly at the basin margin. Thus, missing halite commonly results in little or no departure from regional thickness of Castile evaporites. Stratabound dissolution at great depth is not a unique explanation for the breccias in the south and west portions of the basin that are laterally equivalent to halite beds in the Castile. The possibility remains that the "dissolution breccias" are in part as old as Permian.

Relationships among individual halite and anhydrite beds in the upper Castile and Lower Salado Formations are at present not understood with the same degree of confidence throughout the basin. The difficulty arises from inability to completely separate the effects of (a) depositional heterogeneities, (b) ambiguous identification of members or marker beds, (c) partial removal of halite by dissolution or subaerial erosion, and (d) perturbation of original bed thicknesses by localized deformation. These complications limit the confidence that can be placed in estimates of total amount of evaporite removed by dissolution. The distribution of Ochoan halite was originally not as uniform as that of anhydrite.

The habitats and quantities of water available to dissolve Delaware Basin evaporites are limited. The Bell Canyon, Castile "brine reservoir," and Salado "mine seep" occurrences are neither agents nor consequences of large-scale active dissolution in the north-central Delaware Basin. Dissolution of halite is indicated to be occurring now in the Rustler and upper Salado Formations, in association with meteorically derived waters in the Rustler Formation. Meteorically derived waters in the Capitan limestone are potential agents of evaporite dissolution near the Capitan. Water movement in the Bell Canyon, Castile, and Salado

Formations (except in Nash Draw or areas of subaerial exposure) in the north-central Delaware Basin is either imperceptibly slow or nonexistent. There is evidence of an absence of vertical hydraulic connections among the three units.

Several geomorphological features in the Delaware Basin are at least in part attributable to dissolution of halite or gypsum, or of both. Most of the surficial features of positive and negative relief are related to evaporite removal in the zone within ~1000 ft of the surface, including karst domes, karst mounds, many depressions associated with the Pecos River drainage, and perhaps the Gypsum Plain castiles. The only point-source features thus far found to have a "deep-seated" origin are the breccia chimneys, which result ultimately from collapse not in evaporites but in Capitan limestone.

Of the four models of dissolution (solution-and-fill, phreatic, brine density flow, and stratabound), stratabound would be the most efficient in removal of evaporites at depth. Brine density flow, while physically plausible, is geologically problematical. There is evidence to show that undermining of evaporites to initiate collapse and removal of halite has not occurred in relation to the Bell Canyon Formation or any other unit underlying the evaporites. The Bell Canyon hydrologically and geochemically can have acted as neither a source nor sink for liquids involved in dissolution of overlying or upgradient halite. The brine-density flow model fails in relying on the removal of dissolution brine through the Bell Canyon Formation. The Bell Canyon Formation exhibits insufficient volume or flow capacity, exhibits no saline stratification (cf Hiss, 1975), and its water has characteristics of a stagnant oil-field type brine not related to dissolution so much as to shale ultrafiltration (cf Coplen and Hanshaw, 1973), or perhaps osmotic processes (Graf, 1982).

Dissolution of Ochoan evaporites in some parts of the Delaware Basin (particularly the western part) took place episodically since the Permian. In addition, there is geological evidence for some dissolution in the Triassic, Jurassic, Tertiary, and Pleistocene. There is no evidence that an "undermining" type of dissolution inferred to be responsible for removal of the greater volume of evaporites need be confined to the late Cenozoic. No new geomorphological data are likely to become available to establish time limits on dissolution episodes better than those data presently available.

Dissolution along bedding planes, whose source of water could conceivably be traced to outcrops of Ochoan in the western Delaware Basin, may be (but is probably not) active at present in the west-central portion of the basin, removing evaporites within the Castile and Salado Formations. One limiting factor in the application of the stratabound dissolution model or any dissolution model at depths greater than the upper Salado is as yet the inability to identify an efficient sink for the disposal of saturated brine of dissolution origin.

Implications

Precise predictions are difficult to make for a rate of future dissolution according to any particular model without a consideration of the model's specific details. The episodic nature of dissolution does indeed establish time markers delineating the various discrete episodes. Previous estimates for rates of advance of a "dissolution front" along the WIPP storage horizon have been thought too high by Bachman (1980) and too low by Anderson (1981). The reason for the disparity is attributable to the difference in selection of dissolution model and inferences drawn from the model to calculate rate.

This work has shown that the most rapid dissolution model, regardless of depth, is one similar to the mechanism that removed halite in the Rustler and upper Salado Formations in Nash Draw. The rate of growth of Nash Draw is limited by hydrologic flow in the Rustler Formation. If the Rustler "aquifers" are taken as representative of water movement elsewhere (deeper) in the Ochoan evaporite section, the factors limiting the stratabound dissolution rate are path (a limited permeability), sink (removal of saturated brine by some process of finite rate), and source (the

amount of water available). Thus the anticipated threat to the evaporite horizon serving as the WIPP underground-facility host rock is no greater than that posed by a Nash Draw-type feature developing updip from WIPP in the WIPP horizon. Since Nash Draw has not engulfed Los Medaños in the draw's 600 000+ yr of existence, the threat of dissolution of the WIPP horizon at the WIPP location and depth during the next 250 000 yr is not great.

Nash Draw's rate of formation is calculated by using the dissection of Gatuña, but not penecontemporaneous caliche, 600 000 yr ago. If Nash Draw is a vertically subsiding feature, the development of 100 m of relief in 600 000 yr proceeds at an average rate of 2×10^{-4} m/yr. If Nash Draw is a depression that has migrated along the 1° dip from west to east, keeping pace with stratabound dissolution along marker bed 103, the average rate of 8×10^3 m eastward migration of such an active dissolution feature would be 10^{-2} m/yr. If Nash Draw is hypothetically permitted to grow to its logical culmination, a sediment-filled trough similar to those described by Maley and Huffington (1953), with no impediment to stratabound dissolution (such as the immense problem of brine disposal), in 1.8×10^6 yr the depression would have moved 17×10^3 m eastward, exposing the top of the Salado Formation directly over the WIPP site.

If stratabound dissolution were to begin to preferentially attack the equivalent WIPP horizon at its projected intersection with the surface near the Pecos River, at the sustained rate of 10^{-1} m/yr, $\sim 2.5 \times 10^6$ yr would be required for the dissolution zone to travel the 25 km eastward and downdip to breach the waste emplacement horizon, now at a depth of 650 m. To facilitate this, an active mechanism must continually remove the saturated dissolution brine. In addition, mechanisms to supply water for dissolution and to maintain the necessary open space at depth would have to be just as efficient as observed in Nash Draw, where the quantity and undersaturation of the solution is known to be favorable for dissolution in the Rustler and upper Salado. Many years of drilling experience have failed to reveal either the requisite open space or the solutions with potential to dissolve halite in the lower Salado or Castile Formations.

Even though several point-source geomorphic features are found in the Delaware Basin, development of a collapse structure at a point source near the base of the evaporite section, stopping upward to breach the WIPP facility, is unlikely. The probability of this in

the basin is lower than it would be over the Capitan limestone, since processes of origin would differ. It is unlikely that a large dissolution cavity in rock salt would survive so as to facilitate catastrophic collapse. The Bell Canyon Formation has virtually no potential for initiating collapse in the basin evaporites (Wood et al, 1982). Stratabound dissolution of evaporites produces collapse along a curvilinear trend (cf Nash Draw), not at a point source outlying the "dissolution front."

Recommendations

Several uncertainties remain to be resolved if it is considered necessary to completely and unambiguously understand dissolution processes in the Delaware Basin. If additional confidence is required in the foregoing conclusions for the future integrity of the WIPP facility, the following investigations could be undertaken:

1. The presence of active dissolution can probably be tested only through additional drilling. This would require drilling the appropriate intervals in the upper Castile and lower Salado Formations (to look for dissolution brines of quite different character than in ERDA 6, WIPP 12, etc, particularly in fractured anhydrites), and to core for demonstrable dissolution residue. In addition, a Cenozoic(?) sediment-filled depression would need to be completely cored at one or more locations to determine its origin and possible roles in dissolution.
2. Clay mineralogy studies may help to differentiate between depositional clay accumulations and dissolution residue. Bodine (1978) documented the differences between silicate mineral assemblages disseminated in halite and those deposited in discrete beds.
3. Stable isotope studies of mineral assemblages, including those inferred to be dissolution residues, would reveal the nature of the water, if any, that was last in contact with certain minerals whose oxygen and hydrogen are exchangeable. Such measurements would provide high confidence in calculation of rock/water ratios and in the origin of the water that interacted with the minerals, whether the water is meteoric, intergranular, or the water of crystallization. In many systems, stable isotope studies are the only way to detect such past interactions.

Summary

It was the intent of this work to show that

- Previous (Anderson's) hypotheses for intraformational dissolution have been tested and found to be unsupported by the available data
- There is no direct evidence of present or past preferential removal of lower Salado halite
- A potentially efficient mechanism for stratabound dissolution (more efficient than "brine density flow" involving Bell Canyon, Capitan or Castile) has been identified
- There is little evidence for active stratabound dissolution anywhere save in the Rustler and upper Salado Formations in the Nash Draw area
- If an efficient sink for brine disposal does not exist, there is *no* active dissolution, regardless of the postulated mechanism
- Intraformational dissolution, if it exists, is no more a threat to the WIPP than is the dissolution in Nash Draw
- If greater confidence in the foregoing conclusions is required, specific tests of the stratabound hypothesis could be made, for the hypothesis is testable

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