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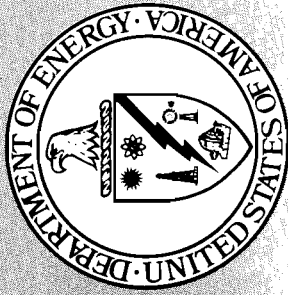
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A Facsimile Report

Karst in Evaporites in Southeastern New Mexico[*]

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Abstract

Permian evaporites in southeastern New Mexico include gypsum, anhydrite, and salt, which are subject to both blanket and local, selective dissolution. Dissolution has produced many hundreds of individual karst features including collapse sinks, karst valleys, blind valleys, karst plains, caves, and breccia pipes. Dissolution began within some formations during Permian time and has been intermittent but continual ever since. Karst features other than blanket deposits of breccia are not preserved from the early episodes of dissolution, but some karst features preserved today—such as breccia pipes—are remnants of karst activity that was active at least as early as mid-Pleistocene time. Rainfall was much more abundant during Late Pleistocene time, and many features visible today may have been formed then. The drainage history of the Pecos River is related to extensive karstification of the Pecos Valley during mid-Pleistocene time. Large-scale stream piracy and dissolution of salt in the subsurface resulted in major shifts and excavations in the channel. In spite of intensive groundwater studies that have been carried out in the region, major problems in groundwater in near-surface evaporite karst remain to be solved. Among these are determination of recharge areas and time of recharge.

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Acknowledgments

Foreword

This report is an account of the field work that was, in the author's words, "the culmination of his geological work, undertaken almost continuously since 1973 in support of the search for a site to store radioactive waste in southeastern New Mexico." The studies herein described rely heavily on the previous work, but entail a large amount of new field work, log interpretation, and review of the karst literature.

The "karst" issue is not a new one. Karstic features in southeastern New Mexico have been known from surficial observations that have been formally documented as early as 1925. Exposures of vast expanses of gypsum on the Gypsum Plain of Texas (the Castile Formation) and Burton Flat, Clayton Basin, and Nash Draw (Rustler Formation) contain a large variety of dissolution features that clearly indicate the importance of water in their formation. Sinkholes, blind valleys, reorganized drainages, and caves in gypsum observable at the surface are all reminiscent of analogous features such as limestone caves in the nearby Guadalupe Mountains. In the carbonate context, it requires but a small step in logic to contemplate the deep subsurface manifestations of cavern formation in the massive Capitan Limestone, which are well known in the drilling history of the Delaware Basin. As this report acknowledges, it is reasonably well accepted that cavern formation in carbonates takes place under phreatic, water-table conditions involving reaction of carbonic (and in some cases, sulfuric) acid with carbonate. The report also outlines the difficulties in the direct applicability to an evaporite terrain of descriptive (and unavoidably genetic) terminology of a substantial karst literature based on processes in carbonates. The very existence and preservation of evaporite terrain at the surface indicates that the amount of water available to a karstic groundwater system is very limited. In addition, the mechanical properties of evaporites do not lend themselves to the preservation of large volumes of open space deep underground to the same extent as brittle carbonates.

This report serves to bound the geological, climatic, and hydrological conditions under which karstic features in evaporites might be formed and preserved. An important consequence of the study is the emphasis on the observability of such features. In the extensive history of exploratory drilling associated with the past 10 years of the WIPP project, there was not found a large open cavity in evaporites (even gypsum) beneath otherwise flat terrain. In each case of anomalous subsurface stratigraphy associated with evaporite dissolution, there was an anomalous surficial feature. The converse, however, was not the case, and proved to be very misleading; geomorphic features alone do not prove the existence of a stratigraphic or structural anomaly associated with dissolution at depth. The drilling at WIPP-13, WIPP-14, WIPP-32, and WIPP-34 are cases in which a geomorphic feature (or a geophysical anomaly) was associated with no missing evaporite section at depth. These findings have resulted in a working hypothesis difficult for some to accept: that evaporite dissolution at depth has surficial manifestations. The coincidence of a gravity anomaly (reported by Barrows et al., 1983), an area of collapse, and gypsum missing in the Rustler Formation at depth at WIPP-33 is documentation of the importance of the presence of a surface manifestation. An active collapse feature will be preserved; if collapse ceases, the feature will become filled (as a consequence of the "fill" aspect of W. T. Lee's 1925 concept of erosion by solution-and-fill) and buried by drifting sand. It therefore becomes irrelevant to the local groundwater system. In the case of the feature at WIPP-33, Barrows et al., (1983, p 63) said, "There is no reason to distinguish this negative anomaly from those detected on the main WIPP site survey; all are assumed to have a common origin." In view of the established geological observability of features associated with dissolution, this may be an unwarranted assumption.

Finally, the "implications of this karst [allegedly developed in the Dewey Lake and Rustler Formations] to the WIPP geohydrology . . . carefully assessed" by Barrows (1982, unpublished memo to

W. D. Weart) must be evaluated in view of all the geologic, hydrologic, and geochemical data. Although this report does not interpret the last two types of available data in detail, the geologic data (bed thickness, depth, and degree of connectedness with water sources) place formidable constraints on the extent of possible karst development in the soluble beds in Ochoan evaporites, especially the deeply buried Rustler Formation at the WIPP site.

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Karst in Evaporites in Southeastern New Mexico

Introduction

This study is part of a series of investigations into the basic geologic setting of the Waste Isolation Pilot Plant (WIPP), a proposed facility for the storage of nuclear waste in underground beds of salt in southeastern New Mexico. Owing to the paramount importance of the continuing integrity of the salt beds in which the waste will be stored, much of the effort of these investigations has been directed towards understanding the natural processes, history, and products of corrosion of salt and associated rocks.

One result of corrosion of soluble rocks is a set of characteristic land forms called karst. The term "karst" has been applied in the literature to various types of terrain and processes (Jennings, 1971; Sweeting, 1973). The imprecision of the term is reason to define it here as it will be used in this report.

Karst is here applied to terrain whose topography and subsurface features are dependent on the dissolution of underlying soluble rocks. Depending on its stage of development, karst may or may not have surface expression. Such terrain is characterized by caverns, sinkholes, valleys without stream channels, and subterranean drainage. The scale of karst features ranges from local collapse of the surface to regional subsidence and from minute etching of rock surfaces to caverns with many connecting rooms.

Although dissolution of evaporite beds has not produced recognizable karst features at the WIPP site, dissolution has been active over a long period of geologic time along the Pecos River and its tributaries—the regional geologic setting of the site. For this reason a study of karst features has been undertaken to understand better the history and processes of their development and to formulate criteria by which karst in evaporites may be recognized.

Many karst areas and features observed today developed under past physiographic and climatic conditions that are greatly different from the present. In these areas karst processes and cycles of erosion that may have been accelerated by periods of increased rainfall are responsible for the dissolution which created the features visible today. At some places even the dependence of surface topography on underground drainage and dissolution is not now apparent. These areas are indicated only by remnants and residues of rock units.

Commonly ancient karst features have been masked by later episodes of erosion and deposition. The term "paleokarst" is applied to these relict features. In a region such as southeastern New Mexico where karst processes have been active for long periods of geologic time, modern karst features may merge almost imperceptibly with paleokarst. Both paleoclimate and regional geology are discussed in this report to provide a setting for these ancient conditions.

A vast literature in several languages treats karst in limestone and dolomite. The terminology of this literature has been summarized in glossaries (Fenelon, 1968; Monroe, 1970; UNESCO, 1972). Much of the discussion of karst has been limited to carbonate rocks, limestone, and dolomite, because of their economic importance in many parts of the world as hosts for supplies of potable groundwater and petroleum products. In addition to their economic importance, a myriad of cave formations as well as the unique environment of caves has attracted a group of specialized explorers, called speleologists.

Although karst in evaporites such as salt and gypsum has long been recognized, less emphasis has been placed on the study of these features than on karst in carbonates. Only a few attempts have been made to synthesize the world-wide occurrence of karst in evaporites (Herak and Stringfield, 1972; Nicod, 1976). Instead, interest has been centered around the problems presented by the dissolved salts that charge the groundwater in evaporite rocks and the hazards to construction posed by the dissolution of these rocks (Soyer, 1962).

Karst processes in evaporites are somewhat different from those in limestone and dolomite, but most of the geomorphic features resulting from both processes are analogous (Reams, 1964). Terminology applied to limestone and dolomite karst is generally adequate for evaporites. Those few features for which terms are not available are defined in this report.

Physiography, Vegetation, and Climate

The WIPP site is located ~25 mi (40 km) east of Carlsbad, Eddy County, NM. The area pertinent to the site and discussed in this report includes the southern Pecos Valley in New Mexico and adjacent portions of western Texas (Figure 1). It is situated in the northern part of the Chihuahuan Desert and the Pecos Valley section of the Great Plains physiographic province (Fenneman, 1962).

Physiography

The Pecos Valley section is marked by flood plain deposits and rolling hills partially covered by windblown sand. The Pecos River and its tributaries is the major drainage system. Surface drainage is southerly and generally parallel to the strike of the easterly dipping bedrock. The Pecos Valley is bounded on the west by the Sacramento section of the Basin and Range province and on the east by the High Plains section of the Great Plains province.
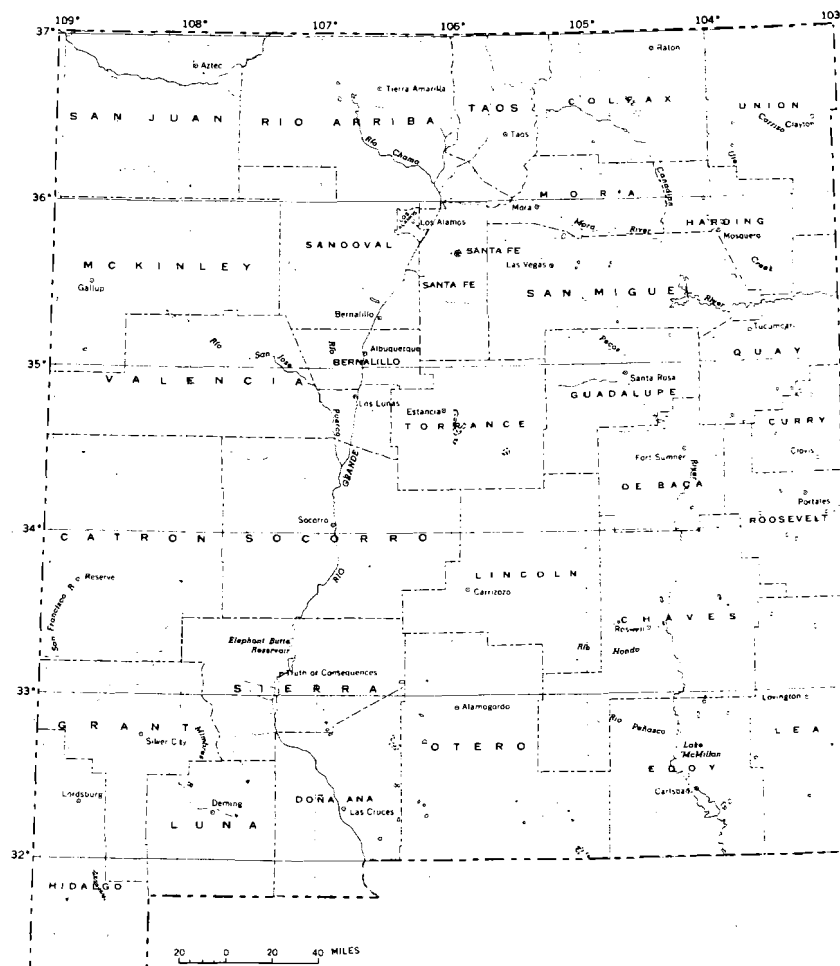
Figure 1. Index map of new Mexico showing the area of Figure 2, the area discussed in this report *(Figure 2 is a foldout on page 75 of this report)*

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Bedrock in the Pecos Valley drainage consists principally of gray limestones, gray-to-white evaporites, and reddish sands, silts, and clays. Evaporites at the surface are represented by gypsum. Salt, anhydrite, and associated minerals are common in the subsurface. Both the limestones and the evaporites are readily dissolved by fresh water, with the result that sinks and caves are abundant along the course of the river. Some rolling hills along the margins of the valley are the result of differential regional subsidence due to dissolution of evaporites or limestones in the subsurface.

Vegetation

Vegetation along the Pecos Valley flood plain consists of scattered trees and shrubs such as valley cottonwood (*Populus wizlizenii*), willow (*Salix* spp.), and salt cedar (*Tamarix* spp.). Where sufficient soil and water are available for irrigated crops, alfalfa and cotton are grown commercially. In the rolling hills and sand dunes away from the flood plain, vegetation is desert scrub, which includes such plants as mesquite (*Prosopis juliflora*) and creosote bush (*Larrea divaricata*). This latter plant association is used by some botanists to define the limits of the Chihuahuan Desert (Mabry et al., 1977). Other plants in this association include white thorn (*Acacia constricta*), bear grass (*Nolina microcarpa*), and allthorn (*Koeberlinia spinosa*).

Modern climate in the vicinity of Carlsbad, NM, is semiarid. Annual rainfall is ~11 in. (280 mm), and the mean annual temperature is 60°F (15.6°C) (Visher, 1954). Annual surface evaporation is ~98 in. (2500 mm) at Lake Avalon near Carlsbad and ~108 in. (2700 mm) at Red Bluff Reservoir on the Pecos River near the New Mexico-Texas State Line (New Mexico State Engineer, 1956, pp 264-267).

Paleoclimate

Karst processes are dependent on the interaction of climate, especially rainfall, and soluble rocks. In regions where climate has been relatively constant for long periods, karstification is a continuous process. In regions where climate has undergone drastic fluctuations through geologic time, the rates of karst processes are variable.

Although the climate of the Pecos Valley in southeastern New Mexico is now semiarid, the present climate is not precisely representative of the past. Extrapolation of present climate as the condition for the dissolution of evaporites and the formation of all karst features visible in the region today would be a misleading assumption. In the remote past, in geologic time more than 190 million years ago (Triassic Period), the climate varied from arid to humid (McGowen et al., 1979, p 15). During humid phases, the low-lying terrain supported a tropical-to-subtropical flora (Dorf, 1970; Ash, 1972). Streams were active agents of erosion and deposition. The present regional distribution of rock units indicates that, although streams flowed in a more easterly direction and drainage systems were different from those of today, some dissolution of evaporites occurred in the vicinity of the present Pecos Valley at that time. The region remained above sea level during Jurassic time. Deposits of Jurassic age are not found in southeastern New Mexico, but deposits in central and northern New Mexico suggest that the region was arid during at least part of Jurassic time. Owing to extensive erosion since deposition, the stratigraphic record of Cretaceous time is poorly preserved. Regional distribution of Cretaceous rocks indicates that shallow seas transgressed across southeastern New Mexico. Again, the climate was moist and subtropical. Forests grew along the swampy margins of Late Cretaceous seas in west-central New Mexico.

The stratigraphic record of early Tertiary time is meager throughout the south-central United States. The Rocky Mountains were being uplifted and the region was subjected to erosion. However, the Ogallala Formation of late Tertiary (Miocene and Pliocene) age is well-preserved on the High Plains to the east of the Pecos Valley. Ogallala deposition in some areas was largely by easterly flowing streams whose courses were controlled by pre-Ogallala erosion channels and collapse basins formed by dissolution of underlying Permian evaporites (Seni, 1980, p 5).

Climate in some areas of Ogallala deposition was generally more humid than at present (Seni, 1980, p 33). However, in southeastern New Mexico large areas of windblown sand of Ogallala age indicate relatively low rainfall (Bachman, 1980) and the deterioration of climate towards more aridity postulated by Frye and Leonard, 1957 for the end of Ogallala time.

During Pleistocene time, within the past 1.8 million years, climate over the entire United States fluctuated widely. It varied from periods of moist, cold continental glaciation to relatively warm and dry intervals. During parts of Pleistocene time, climate in southeastern New Mexico was pluvial with runoff much greater than at present. During Middle Pleistocene time ~500,000 years ago, rainfall and water surplus after evaporation were sufficient to feed a powerful Pecos River with far greater transporting power than at any time since. Some paleokarst features described in this report formed at that time. This unusual climate was followed for several hundred thousand years by somewhat less rainfall. Thick calcareous soils formed that could have been deposited only if the average annual rainfall was no more than 25 in. (625 mm) for extended periods, assuming temperature conditions (and evaporation) as at present.

During parts of Late Pleistocene time (Wisconsinan) ~75,000 to 10,000 years ago, rainfall was more abundant and temperatures were lower than at present. Lakes were present in many parts of New Mexico. In central New Mexico, Lake Estancia covered an area of ~450 square miles (Meinzer, 1911) and could have existed as a perennial lake only if temperatures were lower, rainfall higher, and evaporation less than at present (Leopold, 1951; Antevs, 1954). In southeastern and eastern New Mexico, many perennial ponds and lakes occupied closed basins 18,000 to 13,000 years ago. They were maintained by increased precipitation and/or decreased evaporation (Leonard and Frye, 1975). Many karst features now preserved in southeastern New Mexico formed during this period of increased rainfall.

Geology

The presence of oil, gas, potash minerals, and sulfur in commercial quantities; the attraction of natural features such as Carlsbad Caverns; problems with groundwater quality and quantity; and interest in the WIPP site have generated many detailed reports on the geology of southeastern New Mexico and adjacent areas in Texas. Most of these reports are readily available. For this reason only a brief outline of the regional geology and characteristics of rock types pertinent to this discussion are presented here. This outline is intended to show which rock units may be susceptible to dissolution and the geologic time during which dissolution occurred.

Table 1 summarizes the stratigraphic terminology used in this report, and Figure 2 shows the geographic distribution of rock units. Figure 2, a foldout on page 75, is placed out of sequence at the end of this report so that it can be kept extended for easy and continuous reference as the reader follows the text.

Table 1. Major Stratigraphic and Time Divisions, Southeastern New Mexico

(Time divisions from Berggren, 1972, in part)

Era	System	Series	Formation	Age Estimate
	Quaternary	Holocene Pleistocene	Windblown sand Mescalero caliche Gatuña Formation	~500,000 yr ~600,000+ yr
Cenozoic		Pliocene Miocene	Ogallala Formation	5 million yr 26 million yr
	Tertiary	Oligocene Eocene Paleocene	Absent Southeastern New Mexico	65 million yr
	Cretaceous	Upper (Late) Lower (Early)	Absent SE New Mexico Detritus preserved	
Mesozoic	Jurassic		Absent SE New Mexico	136 million yr 190 to 195 million yr
	Triassic	Upper (Late) Lower	Dockum Group Absent SE New Mexico	225 million yr
		Ochoan	Dewey Lake Red Beds Rustler Formation Salado Formation Castile Formation	
Paleozoic	Permian	Guadalupian	Capitan Limestone and Bell Canyon Formation	
		Leonardian Wolfcampian	Present but not discussed in this report	280 million yr

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Permian Rocks

Rocks of Permian age in southeastern New Mexico are divided into four series: in ascending order, the Wolfcampian, Leonardian, Guadalupian, and Ochoan. Wolfcampian rocks record the transgression of seas across older Paleozoic or Precambrian terrain. A relatively continuous marine, or epicontinental, environment followed this transgression throughout Permian time. Sand, clay, carbonates, and evaporites were deposited. Most of the evaporites were deposited late in Leonardian through Ochoan time.

Guadalupian and Ochoan Series

During Guadalupian time a massive limestone reef, the Capitan Limestone, encircled parts of southeastern New Mexico and western Texas (Figure 3). The area within the reef, the Delaware Basin, was preserved as a depositional basin and received fine-grained clastic sediments, carbonates, and evaporites throughout Ochoan time. These evaporites included thick beds of halite and anhydrite of the Castile and Salado Formations. A shelf area with many playas and pans where thin beds of sand, clay, dolomite, anhydrite, gypsum, and some salt were deposited lay outside the reef to the north. These deposits interfinger and grade into one another with complex stratigraphic relationships (Figure 4).
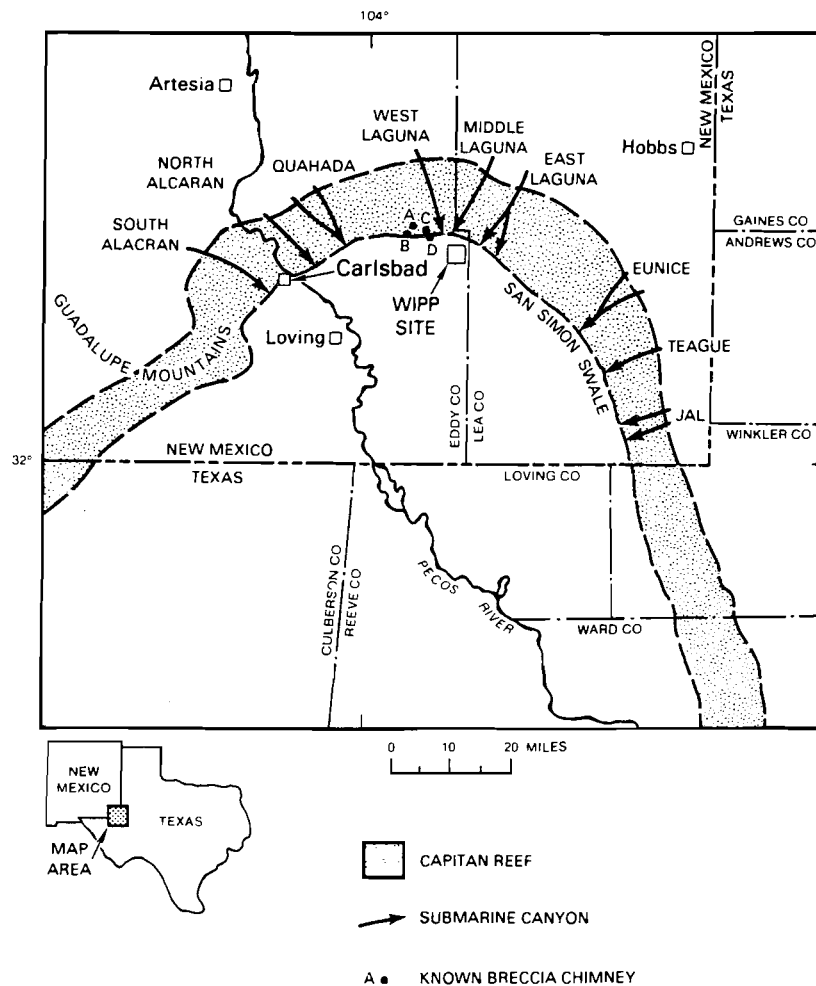
Figure 3. Index map showing location of the Capitan reef and its submarine canyon deposits relative to political boundaries. (Generalized from Hiss, 1975, Figure 11.)
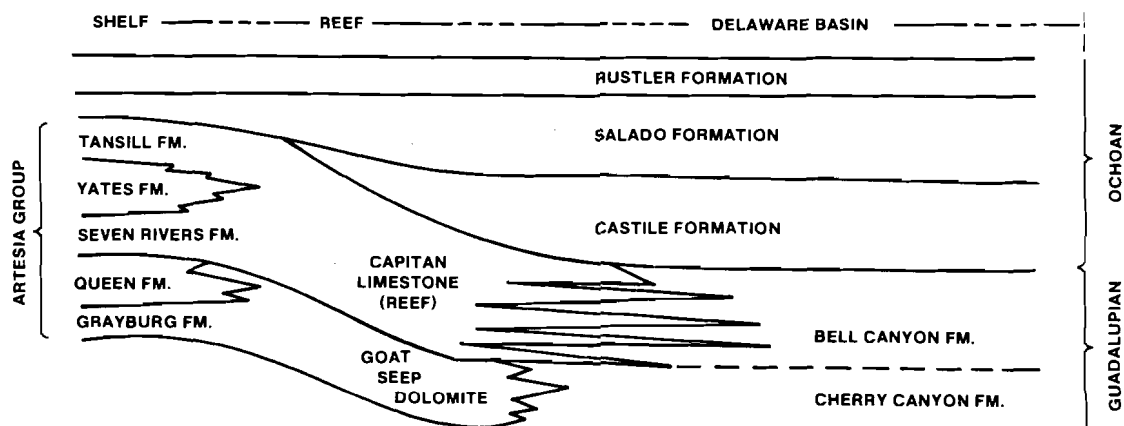
Figure 4. Diagram showing stratigraphic relationships of Permian rocks along the northern boundary of the Delaware Basin

Capitan Limestone and Associated Formations

The Capitan Limestone is a porous and permeable light-gray, fine-grained limestone. In many areas in the subsurface it is a major regional aquifer. Near the surface it is intricately jointed, and large caves have been dissolved in the formation. Both Carlsbad Caverns and New Cave in the Guadalupe Mountains are well-known examples that occur in the Capitan and adjacent Tansill Formation. However, the reef in the subsurface is not consistently permeable. During the building of the reef in Permian time, submarine canyons cut from the shelf area through the reef to the Delaware Basin (Figure 3). These are represented by stringers of fine-grained, carbonate-cemented sandstone that are much less permeable than the adjacent reef limestone. The submarine canyon deposits retard the free migration of groundwater (Hiss, 1975).

The Capitan grades laterally into the Bell Canyon Formation in the Delaware Basin. The Bell Canyon consists of interbedded limestone and sandstone and is well known for the production of oil and gas. In the shelf area away from the basin, the Capitan grades laterally into the upper three formations of the Artesia Group.

Castile Formation

The Castile Formation was deposited entirely within the Delaware Basin and is the basal formation of the Ochoan Series. In the subsurface it consists of laminated-to-massive beds of anhydrite and gypsum with interbeds of halite. The halite and anhydrite units interfinger at several places in the subsurface. Informal members of the Castile Formation have been described and designated by Roman numerals as (in ascending order) the Basal limestone, Anhydrite I, Halite I, Anhydrite II, Halite II, Anhydrite III, Halite III, and Anhydrite IV (Anderson et al., 1972). Later work has shown that Anhydrite IV is further bisected by an additional halite, Halite IV. This results in a unit designated Anhydrite V (Bachman, 1984).

The Castile Formation is exposed at the surface on the Gypsum Plain where it forms the bed rock. There it consists of laminated to massive gypsum (Figures 5 and 6). At the surface it is highly fractured by irregular joint sets. The Castile is as much as 2100 ft (640 m) thick in complete stratigraphic sections in the subsurface.

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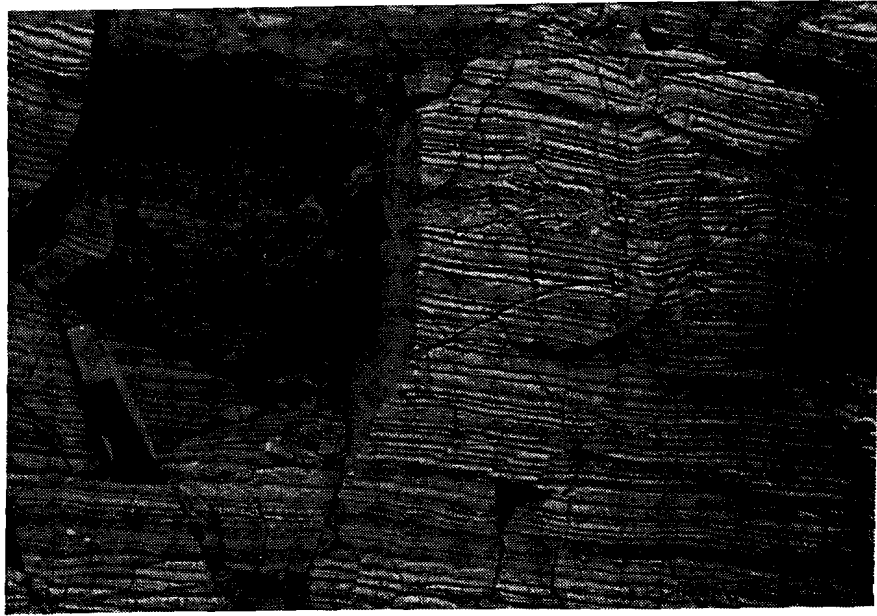
Figure 5. View of laminated gypsum in road cut, Castile Formation, Culberson County, TX. (The open fractures are typical of relatively unweathered near-surface gypsum. In nearby weathered exposures, fractures are filled with earthy gypsite.)

Figure 6. Fluting in horizontal beds of massive gypsum, Castile Formation, Gypsum Plain, Culberson County, TX. (The exposure is in the wall of an arroyo that drains into a cave.)

Salado Formation

Most of the Salado Formation was deposited within the Delaware Basin, although some units transgressed across the reef and were deposited on the shelf to the north. It is known mainly in the subsurface where it is dominantly halite with interbeds of potash minerals (Jones, 1954; 1978) and thin interbeds of anhydrite. Some of these beds of anhydrite and polyhalite are persistent and have been designated numerically as "marker beds" (Jones et al., 1960). Many marker beds are recognizable over much of the Delaware Basin and are indispensable for correlating potash and other horizons within the Salado. They are particularly valuable in determining the presence, or absence, of dissolution within the formation.

The Salado Formation ranges from a knife edge to ~2400 ft (732 m) thick in the Delaware Basin. Some variation in thickness is the result of dissolution. Where exposed at the surface, the Salado is a dissolution breccia composed of chaotic blocks of varicolored gypsum and reddish insoluble residue of clay. Dissolution of the Salado has resulted in the subsidence of broad areas in southeastern New Mexico. The formation has been traced northward outside the Delaware Basin into the shelf area at least as far as the latitude of Artesia, NM. On the geologic map (Figure 2), it is mapped with the Castile Formation in western Texas.

Rustler Formation

The Rustler Formation was deposited over parts of the shelf area far to the north of the Delaware Basin as well as across the basin itself. The formation is divisible into five members (Vine, 1963). In ascending order, these are an unnamed member, the Culebra Dolomite Member, the Tamarisk Member, the Magenta Dolomite Member, and the Forty-niner Member.

The basal unnamed member of the Rustler consists of reddish-to-gray siltstone, gypsum, and anhydrite. It ranges from ~90 ft (27 m) to ~120 ft (37 m) thick.

The Culebra Dolomite is a distinctive marker in the Rustler. It is 25 to 30 ft (7.7 to 9.2 m) thick, brownish-gray, and thinly bedded. Many layers contain abundant vugs that are ~2 to 10 mm in diameter. The vug rims are brownish and may contain minute crystals of selenite. These distinctive rocks are often well preserved in collapse breccia.

The Tamarisk Member is a nondescript massive gray gypsum where exposed on the surface. In the subsurface where dissolution has not destroyed its character, it includes halite, anhydrite, and traces of polyhalite. It is ~180 ft (55 m) thick in complete stratigraphic sections.

The Magenta Dolomite Member is very thinly bedded dolomite laminated with anhydrite or gypsum. The laminae are generally <10 mm thick and are undulatory at surface exposures. The Magenta is light gray to reddish-brown or purplish. It ranges from ~20 to 30 ft (6 to 9 m) thick.

The Forty-niner Member includes gray gypsum and reddish siltstone on the surface. In the subsurface it is composed of anhydrite, siltstone, and halite. It is ~25 ft (8 m) thick.

The depositional thicknesses of members indicated here are considered to be averages. In the central portion of the Delaware Basin the total depositional thickness of the complete Rustler Formation is >400 ft (123 m).

Dissolution affects various portions of the Rustler Formation at many places. The distinctive Culebra and Magenta Members are useful for estimating the amount of dissolution, or hydration, within the formation. For example, in the subsurface where the Rustler approaches its maximum thickness, the Tamarisk Member separates the Culebra and Magenta by ~180 ft (55 m). On the surface in the southern part of Nash Draw and at places on Crow Flat at the latitude of Artesia, the Culebra and Magenta are separated by no more than 5 ft (1.5 m) of insoluble residue. At these localities the Tamarisk Member has been almost completely removed by dissolution (Figure 7).
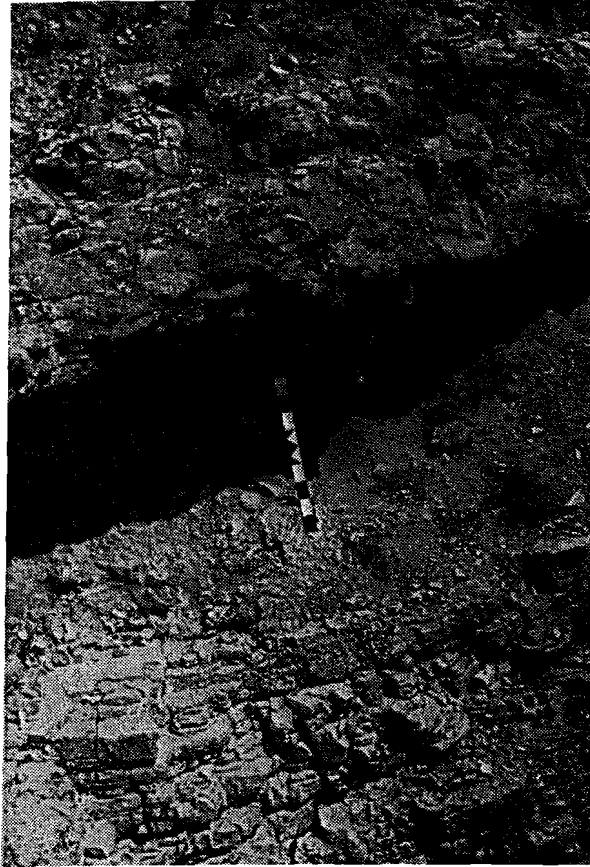
Figure 7. Photograph showing the resistant beds of the Culebra and Magenta Members of the Rustler Formation in outcrops ~12 mi (20 km) east of Artesia, Eddy County, NM (NE 1/4 Sec. 15, T.17S., R.28E.). (At this locality the intervening Tamarisk Member is represented by <5 ft (1.5 m) of insoluble residue (clay and silt). In complete undissolved stratigraphic sections in the subsurface to the southeast, the Tamarisk averages 80 to 100 ft (25 to 30 m) in thickness and consists of anhydrite or gypsum, halite, and clay. The scale is 1 m.)

Dewey Lake Red Beds

The Dewey Lake Red Beds rest conformably on the Rustler Formation and consist of alternating thin, even beds of reddish-brown siltstone and fine-grained sandstone. Small-scale cross-laminations and ripple marks are not uncommon. Many beds are mottled by greenish-gray reduction spots that are circular in outline on flat surfaces. Well-sorted quartz grains make up most of the rock. Selenite, clay, and small amounts of carbonate cement the grains.

Rocks of Triassic age rest unconformably on, and lap across, Dewey Lake Red Beds. This stratigraphic relationship, as well as the thickness and distribution of the Dewey Lake, suggests that moderate erosion with dissolution of some portions of the underlying evaporites may have occurred during the interval after the deposition of the Dewey Lake and before deposition of the Late Triassic sediments.

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The Dewey Lake in the subsurface in eastern Eddy and western Lea Counties is ~560 ft (172 m) thick. It thins towards the west and is absent at many exposures where Triassic rocks lap across the westernmost line of outcrop to rest on the underlying Rustler Formation.

Back-Reef Formations

San Andres Formation

The San Andres Formation is a widespread bedded back-reef limestone in central New Mexico. It forms the eastward-dipping back slope of the Sacramento Mountains and is in the subsurface along the Pecos River Valley northward from the vicinity of Carlsbad. In the subsurface to the east of the Pecos Valley it includes gray limestone, anhydrite, gypsum, and salt. North of the Delaware Basin and east of the Pecos River, the San Andres is at least 1150 to 1180 ft (348 to 358 m) thick in the subsurface. Along the Pecos River in the vicinity of Roswell, as much as 400 to 600 ft (123 to 185 m) of evaporites have been dissolved from the upper part of the formation in the subsurface; that interval consists of a dissolution breccia (Welder, 1983, pp 6-8).

Artesia Group

The Artesia Group (Tait et al., 1962) is the shelf, or back reef, facies of the Guadalupian Series. It includes five formations—in ascending order, the Grayburg, Queen, Seven Rivers, Yates, and Tansill. The formations consist of interbedded dolomite, limestone, reddish brown-to-yellowish siltstone, sandstone, and evaporites. Gypsum is the dominant evaporitic rock at the surface, but anhydrite and salt are present in the subsurface.

In the vicinity of the Capitan reef the carbonate facies is dominant. The evaporite facies dominates in an area ~100 mi (160 km) shelfward from the reef, and a clastic facies covers the remaining part of the shelf (Tait et al., 1962, p 515). Although the clastic facies is dominant in the sequence farthest from the reef, some thin beds of gypsum are present in the Artesia Group in surface exposures at least 200 mi (325 km) north of the reef. The group is ~1700 ft (523 m) thick near the northern end of the Guadalupe Mountains and thins to a wedge-edge far to the north in the Sangre de Cristo Mountains.

Some marker beds persist in the subsurface over broad areas, but at many places it is difficult to subdivide the Artesia Group into its component formations. This difficulty is the result of intricate interfingering of rock types. For this reason the broader term "Artesia Group" is used in much of this discussion to include one, or all, of its formations.

Karst features, including collapse sinks and karst valleys, are prominent in the Artesia Group along the Pecos Valley northward from Carlsbad. The Bottomless Lakes in the vicinity of Roswell are collapse sinks in the Artesia Group.

Rocks of Triassic Age

Rocks of Triassic age have been called the Santa Rosa and Chinle Formations in southeastern New Mexico, with little justification. Many exposures of Triassic rocks in that area are erosional remnants that cannot be traced into their type localities. For this reason the terms Dockum Group, or rocks of Triassic age, undivided, are used in this report. There is a considerable gap in time between deposition of Ochoan rocks and those of Triassic age, and the Dockum Group was not deposited until Late Triassic time.

Rocks of the Dockum Group contrast with the underlying clastic rocks of the Ochoan Group and are readily distinguished on the basis of color, grain size, texture, and habits of exposure. The Dockum includes conglomerate, coarse sandstone, and shale that are generally dark reddish-brown. Large portions of the rock may show areas of greenish-gray reduction, but they lack the mottled appearance of the underlying Dewey Lake beds. Cross-laminations in the Dockum Group are large-scale and may show torrential cross-bedding. The sandstones are poorly sorted and contain conspicuous ferromagnesian minerals. Although salt hoppers may be present in some beds of the Dockum Group, there are no beds of evaporites.

The Dockum group was deposited as a complex of fluvial-deltaic-lacustrine systems; the depositional wedge-edge was along a line near the present Guadalupe Mountains and near the present limits of its preservation (McGowen et al., 1979). Its source area was west of this wedge-edge, and it is absent over a broad area in south-central New Mexico (Bachman, 1976).

Along the east side of Nash Draw the Dockum group is ~75 ft (23 m) thick. Eastward in the subsurface in western Lea County it is as much as 1500 ft (460 m) thick. This abrupt thickening within a distance of <20 mi (32 km) suggests that the gradient of the Triassic streams could have been as much as 75 ft/mi (14 m/km), which is improbable. Although there is torrential cross-bedding in some beds of Triassic sandstone, there is no evidence in the sediments that the gradients of Triassic streams were of this magnitude. The abrupt thickening of Triassic rocks eastward from their wedge-edge is here attributed to mild tilting of the depositional basin to the east during Triassic time and to post-Triassic erosion along the western margin of the basin. The tilting would account for erosion of the Dewey Lake Red Beds along the western edge of their present exposures and suggests that physiographic conditions were propitious for dissolution of some Permian evaporites at that time.

Brecciated Triassic rocks are also present in isolated collapse sink deposits along the Pecos River drainage ~5 mi (8 km) northeast of Carlsbad and along the east side of Red Bluff Lake. Triassic rocks have not been observed west of the Pecos River except to the north of Roswell. A large collapse block of Triassic rocks at least 1/2 mi (0.8 km) long is present on Five Mile Creek ~27 mi (43 km) north of Roswell. This occurrence was discovered recently by R. L. Borton (personal communication) and contributes to the history of dissolution and the history of the Pecos River in that area. Farther north, ~18 mi (29 km) south of Fort Sumner, the Dockum Group strikes westerly across the Pecos, where it merges with a normal sequence of exposures (Figure 2).

Rocks of Cretaceous Age

There is a major hiatus between rocks of Late Triassic age and those of Early Cretaceous age in southeastern New Mexico. The region was above sea level throughout intervening Jurassic time. Jurassic rocks are not present in the southern half of New Mexico, and stratigraphic relationships in adjacent regions indicate that rocks of this age were never deposited. There is no positive evidence of dissolution of Permian evaporites during Jurassic time, but the fact that the region was above sea level suggests that dissolution may have occurred then.

Although Early Cretaceous seas were present across southeastern New Mexico, their deposits have since been largely eroded. Rocks of Early Cretaceous age are preserved at only a few localities in the Pecos drainage system as collapse debris in areas of dissolution. Fossiliferous sandy limestone debris of Early Cretaceous age occurs at two localities, ~6 mi (9.6 km) and 7.8 mi (12.5 km), respectively, southwest of Whites City. The first of these appears to rest in a collapse sink ~200 ft (60 m) in diameter in the Castile Formation. Another occurrence of Cretaceous rocks and associated fossils is ~5 mi (8 km) east of Carlsbad.

All three of these exposures are keys to interpreting the present distribution of Permian, Triassic, and Cretaceous rocks. At the two localities southwest of Whites City, the fragmentary Cretaceous rocks and fossils rest on gypsum of the Castile Formation. Dolomite of the Culebra Member of the Rustler Formation is in close proximity to these occurrences, but both the Dewey Lake Red Beds and rocks of the Dockum Group are absent. These relationships indicate that the Dewey Lake and Dockum either were never deposited in the vicinity of these localities or that they had been eroded before Cretaceous time. They also indicate that intervening strata between the Castile and the Culebra have been removed either by erosion or dissolution, or that they were never deposited. At the locality northeast of Carlsbad, Cretaceous rocks are mingled with debris of Culebra Dolomite and Dockum conglomerate. At this locality rocks of the Dewey Lake Red Beds are absent and are presumed to have eroded before Triassic time.

Outliers of marine Early Cretaceous rocks are present in Lea County, NM, in the headwaters of the Texas Colorado River drainage east of the modern Pecos system (Ash and Clebsch, 1961). These rocks

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provide further evidence of distribution of Cretaceous seas. In places these rocks occur as randomly oriented blocks of limestone in closed depressions, indicating subsidence after lithification. Undisturbed late Tertiary strata overlie the Cretaceous rocks, which indicates that subsidence occurred before, or during, Late Cenozoic (Ogallala) time.

All these localities indicate that in southeastern New Mexico Permian evaporites were near the surface and available for dissolution by surface or groundwater during geologic time preceding the invasion of Cretaceous seas, as well as later in Cenozoic time. The present fragmentary nature of these exposures indicates much erosion and dissolution since Cretaceous time.

Cenozoic Rocks

A major hiatus marks earliest Tertiary time in southeastern New Mexico; and, other than a few thin igneous dikes, rocks of Early Tertiary age are not preserved. The only sedimentary rocks of Early Cenozoic age in the vicinity are in the Sierra Blanca region ~110 mi (176 km) northwest of Carlsbad where continental rocks of probable Paleocene or Eocene age are preserved. Owing to the absence of other deposits and regional evidence of continental uplift of the western United States, it is assumed that all of southeastern New Mexico has been above sea level and subject to erosion since Cretaceous time.

Upper Tertiary and Quaternary formations are the only Cenozoic rocks preserved in southeastern New Mexico. These include the Ogallala Formation of Miocene and Pliocene age, the Gatuña Formation of Pleistocene age, and various surficial deposits.

Ogallala Formation

The Ogallala formation of Miocene and Pliocene age is the earliest record of Cenozoic depositional and climatic history preserved in southeastern New Mexico. It includes alluvial, eolian, and lacustrine deposits, and it forms the "caprock" underlying the southern High Plains. It is well-exposed along Mescalero Ridge, Hat Mesa, Grama Ridge, and The Divide, where it rests on rocks of Triassic age. The Ogallala Formation has been definitely recognized west of San Simon Swale only in the relatively thin exposures at The Divide. However, some of the thick sedimentary fill in collapse sinks in the present Pecos drainage may be of Ogallala age.

The Ogallala Formation was derived from the west, where sediments were eroded from the newly uplifted Rocky Mountains and fault blocks of the southern Basin and Range. The formation was deposited on an irregular early Cenozoic erosion surface as a complex sequence of alluvial fill, fans, inset valley deposits, and windblown sand. Some irregularity of the early Cenozoic erosion surface was the result of dissolution of underlying evaporites and collapse of the surface.

At the end of Ogallala deposition, a pedogenic caliche, the Ogallala "climax soil" (Frye, 1970), was deposited on the High Plains surface, and the well-known "caliche caprock" was formed as part of this soil process. This caliche has been desiccated, brecciated, and recemented through many generations until its upper surface is a complex mixture of laminar deposits and pisoliths. Its texture and structure are characteristic of very ancient pedogenic caliche (Bachman and Machette, 1977).

In southeastern New Mexico, the Ogallala Formation underlying the caprock consists largely of well-sorted windblown sand. Some poorly sorted stream deposits and local carbonate pans are included in the formation. The Ogallala is ~26 ft (8 m) thick at The Divide, thickening to ~400 ft in the vicinity of Mescalero Ridge.

During the Tertiary Period, before Ogallala time, the course of drainage systems can only be surmised, but they are presumed to have been easterly away from the newly uplifted Rocky Mountains. During Ogallala time, drainage was easterly at least as far south as the latitude of Fort Sumner and possibly as far south as the latitude of Roswell and Artesia. From Roswell southward, however, the nature of the Ogallala drainage system is less clear. The Divide may have been a drainage divide even during Ogallala time, but the stream gravels at The Divide give no clues as to the direction of the main drainage. Thick deposits of alluvial fill in the Pecos drainage to the south of the New Mexico-Texas

State Line (Maley and Huffington, 1953) and in the Roswell Artesian Basin (Welder, 1983) may include some sediments of Ogallala age. These deposits of alluvial fill are of particular interest to the present study because they are preserved in collapse sinks that are the result of dissolution of Permian evaporites in the subsurface.

Gatuña Formation

The Gatuña Formation was named for exposures in Gatuña Canyon on the east side of Clayton Basin (Robinson and Lang, 1938, pp 84-85). A type stratigraphic section was designated and described by Bachman (1976, p 140). The formation is distributed intermittently over a broad area in the Pecos drainage system. It is recognized in areas east of the Pecos at least as far north as Roswell, Artesia, and Hagerman. It is present along the east side of Clayton Basin and Nash Draw and is especially well-exposed in Pierce Canyon and its tributaries. Along the New Mexico-Texas State Line, a gravel unit is assumed to be part of the Gatuña Formation.

Far to the north in the vicinity of Santa Rosa a sequence of shale, sand, and gravel may be at least partially equivalent to the Gatuña Formation.

The Gatuña Formation in the vicinity of Nash Draw and Pierce Canyon consists of pale reddish-brown and yellowish sand, sandy clay, and lenticular beds of gravel (Figure 8). The gravel contains pebbles derived from a variety of sources. They include reworked quartzite pebbles from the Dockum Group, pebbles from Tertiary igneous masses far to the northwest in the Sierra Blanca and Capitan Mountains, and clasts of pisolitic caliche presumed to be eroded from the Ogallala caprock to the east. At the reference section in Gatuña Canyon, more than one-third of the clasts in the conglomeratic gravel were pisolitic caliche.
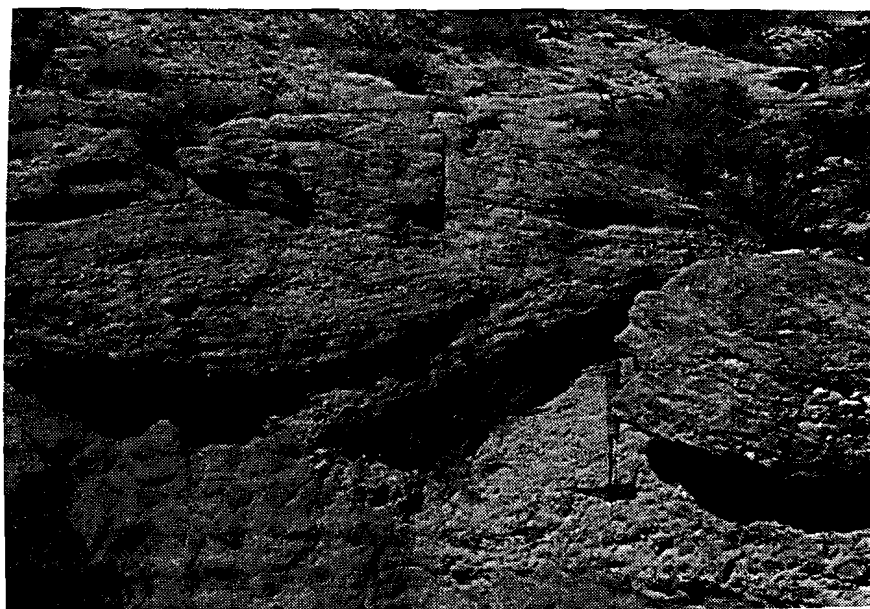
Figure 8. Conglomeratic, cross-bedded sandstone filling a channel at the base of the Gatuña Formation, Nash Draw, Eddy County, NM. (This sandstone was deposited by a stream that flowed southwesterly across the area now occupied by Nash Draw. Late in Gatuña time the stream cut into evaporites in the Rustler Formation, and Nash Draw began to form as a series of collapse sinks aligned along the strike of the Rustler. The stream was disrupted, and the sandstone is now suspended on the east slope ~100 ft (30 m) above the floor of Nash Draw.)

A bed of volcanic ash 1 m thick is present in the Gatuña Formation on the east side of Nash Draw. This has been determined, by potassium-argon and fission track dating of zircon, to be the Lava Creek B ash derived from the Yellowstone Park region and is ~620,000 years old (Izett and Wilcox, 1982). In a roadcut west of Fort Sumner in beds that may be equivalent to a basal part of the Gatuña Formation, a bed of volcanic ash 0.6 m thick is preserved. It is the Guaje ash derived from the Jemez Mountains and is ~1.4 million years old (Izett et al., 1981) based on K-Ar dating. It is equivalent to early Pleistocene deposits in other parts of North America. The sequence west of Fort Sumner was deposited in a different drainage system and is much older than the Gatuña in Nash Draw, but it may be equivalent to some part of the alluvial fill in deep solution basins along the Pecos River in Southern Eddy County, NM.

Soils

Soil accumulation results from processes that are extremely sensitive to climatic conditions and landscape stability. If rates of erosion, accumulation of sediments, or disturbance of land surface exceed the rate of soil formation, soil cannot be deposited. Owing to the value of soils as indicators of climate and surficial stability, they were examined as part of the stratigraphic sequence in this study. Paleosols, ancient soils, are of particular interest because they may record climatic conditions that were considerably different from the present.

Soils are a product of the interaction of weathering of bedrock, infiltration of moisture, animal and plant activity, and atmospheric dust and gases. These processes result in a mineralogic zonation, or profile, that differs from the underlying parent material and reflects local climatic conditions and environment.

Soil profiles are divided into three mineral horizons designated, in descending order, as A, B, and C (US Department of Agriculture, 1975). The A horizon at the surface is characterized by organic material, but it is not preserved in ancient soils and commonly is poorly developed in modern soils in semiarid regions such as southeastern New Mexico. The B horizon underlies the A and includes alluvial concentrations of clay or iron compounds. The C horizon underlies the B and is the weathered part of the profile overlying bedrock where soil-derived cements accumulate. These cementing substances include calcium carbonate, which is a conspicuous light gray-to-white deposit capping stable geomorphic surfaces in many parts of the semiarid world.

The carbonate deposit in the C horizon has received many provincial and imprecise names such as caliche, calcrete, kunkar, and croute calcaire. It forms the "caprock" on the High Plains and caps various broad surfaces in southern New Mexico. This carbonate has been distinguished from other C horizons by the designation "K horizon" (Gile et al., 1965). Although it is recognized that the term caliche is imprecise, in this report the term caliche is used to designate calcium carbonate deposits, the K horizon, that are generally parallel to the topography, and have a distinctive and predictable morphology. The topography on which the caliche is deposited may vary from relatively flat to rolling.

Mescalero Caliche

The Mescalero caliche is a paleosol and an informal stratigraphic unit named for the Mescalero Plain, a broad geomorphic surface, that lies east of the Pecos River and west of the High Plains in southeastern New Mexico (Bachman, 1976, p 141). The Mescalero caliche was deposited in an aggrading eolian environment from windblown sand, dust, and rainwater during an interval of climatic and tectonic stability that followed Gatuña time. Small quantities of calcium carbonate were leached from sand and dust and deposited in underlying soil horizons by downward-percolating soil solutions. During dry periods the sand and dust are reworked and moved about on the surface, and a new source of calcium carbonate is introduced seasonally into the region. This process has been studied in south-central New Mexico by Gile et al., (1981, pp 66-71).

Modern accumulations of pedogenic caliche do not occur where the rainfall over extended periods exceeds ~25 in. (635 mm) per year or where rainfall is less than ~2 in. (50 mm) per year. This relationship to precipitation provides a key to the climatic history of some regions.

On the basis of morphologic development, the Mescalero caliche has been correlated with the Middle Pleistocene caliche on the Llano de Albuquerque surface in the Rio Grande Valley (Hawley et al., 1976, pp 244-245). Uranium series disequilibrium measurements indicate that the Mescalero caliche began to form ~510,000 yr ago (J. N. Rosholt, written communication, 1979).

Berino Soil

The Berino soil is an informal stratigraphic unit that is here applied to the "Berino series" as used by the US Department of Agriculture for mapping soils in Eddy County, NM (Chugg et al., 1971). As used in this report, the Berino is a dark-red, sandy, clayey paleosol that overlies the Mescalero caliche at some places in the vicinity of the WIPP site. The Berino is usually overlain by windblown sand, but it is exposed at construction sites and was examined at many places in hand auger cuttings during this study. Owing to desiccation and erosion, the Berino varies in thickness but is rarely more than 1 m thick.

The Berino is noncalcareous and probably represents the remnant of an ancient B horizon. Uranium series disequilibrium studies indicate that the Berino, as it is presently preserved, began to form ~350,000 (±60,000) years ago (J. N. Rosholt, written communication, 1979).

Spring Deposits

Soft, earthy, light-gray gypsite is inset along the eastern side of Nash Draw as spring mounds and evaporated runoff deposits. Six low mounds ~5 to 10 ft (1.5 to 3 m) high are aligned northeasterly along the crest of a low anticline in Dewey Lake Red Beds (E 1/2 Sec. 15, T.22S., R.30E.). These deposits are interpreted as the result of evaporation of sulfate-bearing water that circulated through and dissolved parts of the Rustler Formation underlying Livingston Ridge and the eastern edge of Nash Draw and reached the surface at the position of the spring mounds. The runoff deposits contain fossils of horse, camel, and related Pleistocene forms and are believed to be Late Pleistocene in age.

Windblown Sand and Alluvium

Windblown sand is common and has accumulated in dune fields in southeastern New Mexico. Some dunes are presently active, but most are partially stabilized coppice dunes. The Mescalero sands east of Roswell and the Los Medaños field in the vicinity of the WIPP site are the largest areas of windblown sand accumulation in the region. Floodplains with large accumulations of sand are not characteristic of the Pecos drainage and are not potential sources of windblown sand. Large accumulations of windblown sand are presumed to be recycled by erosion from sands in the Ogallala Formation.

Alluvium, including silt, sand, and gravel, is present locally along the Pecos River and its tributaries. Deposition and erosion of alluvium in karst areas are important gauges of karst processes and will be discussed in other parts of this report.

Geologic Structure

After development of the Delaware Basin as a depositional basin and its filling by sediments during Permian time, the structural evolution of southeastern New Mexico has been relatively simple and has consisted mainly of regional warping (epeirogenic activity). Uplift of the Sacramento Mountains and igneous intrusion of the Capitan and Sierra Blanca uplifts to the west produced some local folding and faulting during Cenozoic time, but these movements are reflected in southeastern New Mexico mainly by gentle easterly dips in the bedrock. In the Delaware Basin these dips average ~75 to 100 ft/mi (14 to 20 m/km).

Although the eastern scarp of the Guadalupe Mountains is abrupt, there is no evidence of faulting along the front (Hayes and Bachman, 1979). The steep scarp is an erosional remnant supported by the dense Capitan reef and associated carbonate beds.

Joint sets are fractures in rocks at, or near, the surface. These rocks were formerly buried under younger sediments, compressed, and compacted. During regional uplift the overlying rocks were removed by erosion, which decreased both the gravitational pressure and lateral stresses. Joints are the result of this release of stress (Price, 1959). Most of the unloading of bedrock by erosion and denudation was during Cenozoic uplift in southeastern New Mexico. It is therefore assumed that major joints sets, accompanied by dissolution, were formed during Cenozoic time.

In southeastern New Mexico, intricate joint sets are conspicuous surficial structural features that have allowed surface water to infiltrate and have contributed to the formation of karst in evaporites (Figures 9, 10, and 11). The joints allow water to infiltrate from the surface and, with time, are widened by dissolution (Lattman and Olive, 1955). These solution-widened joints become pathways for water to penetrate underground and dissolve evaporites along bedding planes, producing caverns and collapse sinks.

Figure 9. Aerial view of joint sets in gypsum, Gypsum Plain, Culberson County, TX. (Some individual joints are traceable for more than 1 mi (1.6 km). Vegetation grows in joint partings.)

Figure 10. Aerial view of fracture (joint) pattern in gypsum on surface adjacent to collapse sink ~2 mi (3 km) south of Bottomless Lakes State Park, Chaves County, NM. (The fractures are ~3 to 10 ft (1 to 3 m) apart. Grasses grow in the fractures.)

Figure 11. Photograph showing characteristic fracture pattern in massive gypsum, Castile Formation, Gypsum Plain, Culberson County, TX. (Fractures are filled with earthy gypsite. See Figure 10 for contrast.)

Evaporites

The major evaporitic rock types in southeastern New Mexico are gypsum ($CaSO_4 \cdot 2H_2O$), anhydrite ($CaSO_4$), and rock salt (NaCl). Potassium-bearing minerals such as sylvite (KCl), carnallite ($MgKCl_3 \cdot 6H_2O$), and langbeinite ($Mg_2K_2SO_4)_3$, which are interbedded within the rock salt, have been mined for potash minerals near Carlsbad, NM, since 1931.

These rocks were deposited originally by evaporation of sea water in shallow lagoons, pans, and basins. At times the depositional environment may have been analogous to playas, or sabkhas, along the present coasts of western Mexico or of the Trucial States adjacent to the Persian Gulf. However, much deposition was within the deeper part of the Delaware Basin in southeastern New Mexico and western Texas. Deposition of these evaporites occurred during Late Permian time more than 225 million years ago.

Rock salt and its associated potassium-bearing minerals are very soluble in fresh water. For this reason they are readily dissolved by groundwater and rarely appear in surface outcrops. Where the stratigraphic interval of the rock salt is present at the surface, it is represented only by a residual dissolution breccia consisting of a chaotic mass of gypsum fragments, siltstone, and clay (Figure 12). Knowledge of the normal, undissolved stratigraphic sequence of the salt and potash beds in the subsurface has been gained from mining operations, drill holes for petroleum, and exploratory drill holes to study the WIPP site.

Although anhydrite occurs widely in the subsurface in southeastern New Mexico, this form of calcium sulfate is not stable under surface conditions in that area. Instead, calcium sulfate occurs in the hydrated form as gypsum, a widespread rock type in surface outcrops. Dissolution of rock salt and the hydration of anhydrite to gypsum contribute to the formation of karst and are responsible for extensive modifications in the subsurface, but most karst features observed at the surface occur in bedded gypsum.

Bedded gypsum is widely distributed at, or near, the surface in southeastern New Mexico and adjacent areas in Texas in various formations. Bedrock in the Gypsum Plain and Yeso Hills is gypsum of the Castile Formation. Gypsum in the Rustler Formation is the bedrock in much of Nash Draw, Clayton Basin, and Burton Flat. It is a significant part of the Artesia Group that forms the bedrock along the Pecos Valley northward from Carlsbad to Roswell and almost as far north as Fort Sumner. The total area of exposure of gypsum and associated sediments in this region is estimated to be at least 1400 mi^2 ($\sim$3,600 km^2). Anhydrite and salt are a part of the stratigraphic section in the subsurface in the Delaware Basin and along the Pecos Valley.
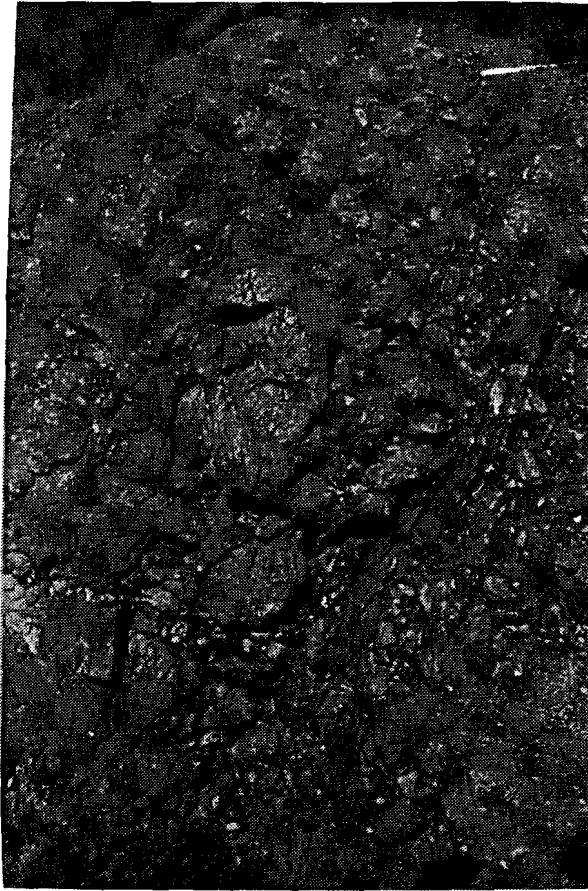
Figure 12. Dissolution breccia of the Salado Formation, Eddy County, NM

Dissolution and Karst

Abrasion is part of the erosive process in the formation of karst topography, but dissolution is the most dynamic process. The dissolution of constituents from rocks increases their permeability, which in turn enhances further dissolution. Mineral constituents (salts) are carried away from the karst region in solution. These solutions are highly mobile and may flavor many environments on their way to the sea.

Limestone ($CaCO_3$) and dolomite ($CaCO_3 \cdot MgCO_3$) form the rocks best known for their role in the formation of classic karst. These carbonate rocks are dissolved by the simple combination of carbon dioxide in the atmosphere with groundwater to form carbonic acid:

$$CO_2 + H_2O \rightleftarrows H_2CO_3 .$$
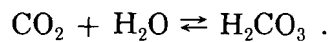
Carbonic acid readily attacks carbonates (limestone) to form highly soluble calcium bicarbonate:

$$H_2CO_3 + CaCO_3 \rightleftarrows Ca(HCO_3)_2 .$$
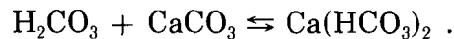
The degree of saturation of carbon dioxide in the solution controls the amount of carbonate dissolved. An increase of carbon dioxide in water heightens its dissolving power. Reduction of partial pressure of carbon dioxide in cave atmosphere reverses the process and leads to precipitation of carbonate to form stalactites, stalagmites, travertine, and related formations (Jacuks, 1977, p 28). The chemistry of the decomposition of carbonates has been discussed extensively in the literature and summarized by Jacuks (1977, pp 26-77) and Jennings (1971, pp 23-30).

Although the dissolution of evaporites is a simpler process, the formation of karst in evaporites has attracted less attention. Dissolution of these rocks requires only water to act as the solvent. Dissolution becomes a continuous process where unsaturated water flows through the evaporite in an open system that allows the solute to be carried away. Dissolution ceases where the system becomes closed or where the solution becomes saturated.

Three general types of dissolution occur in the Pecos River drainage system:

1. *Local dissolution near the surface in the vadose zone (above the water table).* Sinkholes, solution-widened fractures, and some caves form in this zone.

2. *Regional, or bulk, dissolution that may occur either in the vadose or shallow phreatic zone (just below the water table or the zone of saturation).* Some caves and solution breccias that develop regionally at a stratigraphic horizon are in this category.

3. *Deep-seated dissolution that occurs in the deep phreatic zone.* Some caves, such as Carlsbad Caverns, may undergo their initial development in this zone. It is notable that the dissolution of soluble rocks is not restricted to the base level of erosion. Moneymaker (1941) reported cavities in limestones at depths >100 ft (30 m) below the zone of saturation in the Tennessee Valley. Morgan (1941, p 782) reported cavities in the San Andres Limestone near Lake McMillan 1700 ft (515 m) below the surface.

These types of dissolution commonly overlap, and at places it is difficult to assign a given feature to a single environment. If possible, it is more desirable to learn the history of a feature.

In addition to dissolution, anhydrite may hydrate changing composition and crystal form to become gypsum:

$$CaSO_4 + H_2O \rightarrow CaSO_4 \cdot H_2O$$
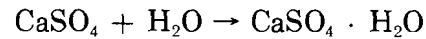
anhydrite + water $\rightarrow$ gypsum .

This reaction results in an expansion in volume of $\sim$36.5% (Jacuks, 1977, p 79) and has led to the argument that "in a gypsum karst no deep passage for water can form . . . since only the top few meters consist of gypsum, the rest of the deposit below being anhydrite. . . ." (Jacuks. 1977, p 80). Theoretically, fractures in anhydrite should be self-healing because of this expansion on conversion to gypsum. However, field observations indicate that many joints in evaporites are long-standing and do allow relatively deep infiltration of water from the surface.

Snyder (1985, p 10) has observed that the removal of halite in an evaporite sequence allows anhydrite to "settle and crack." Groundwater then flows more freely through the sequence, and the anhydrite hydrates to gypsum. This hydration tends to thicken the formation. Mutual interaction of hydration and dissolution results in erratic thickening and thinning of the sequence.

Karst Features

Many attempts have been made to classify karst features. Most of these classifications are genetic and are based on interpretations of the processes that appear to be forming the karst features at the present time. Consideration of genesis cannot be avoided, but genetic classifications are often misleading because it is impossible to observe processes in operation over long spans of geologic time. To avoid the circular reasoning that often pervades genetic classifications in geology, we find it is more practical to classify on the basis of direct observation of physical features where possible and to interpret genesis from these observations.

The processes governing the formation of karst are limited in number. The major problems in classification of karst are the variables arising from fluctuations in rates of dissolution under changing hydrologic conditions through geologic time. Quinlan (1968) examined these and other variables and presented an extensive classification that is used in parts of this discussion. The features that result from these variables may be divided artificially into (1) those which presently are at, or near, the surface, and (2) those underground. Bogli (1980) has approached the problem of classification in this manner by distinguishing between exokarst (surficial karst phenomena) and endokarst (underground karst phenomena). Although it is recognized that these two phenomena overlap, an approach similar to that of Bogli is used here.

Karst terminology is derived from several languages, which results in many synonymous terms and fine distinctions. This complex terminology is fitting for diverse geographic and geomorphic conditions. The following discussion is an attempt to define some of the more common terms and to simplify them for provincial usage.

Surficial Karst

Surficial karst includes features that form at or near the surface in the vadose zone. These include dolines, collapse sinks, solution-subsidence valleys, and small-scale surficial etched features such as solution runnels. Portions of large-scale features such as caverns and breccia pipes appear at the surface as surficial karst, but they have a long geologic history, and their roots are deep underground.

Dolines

A doline is a "basin- or funnel-shaped hollow in limestone, ranging in diameter from a few meters to a kilometer and in depth from a few to several hundred meters. Some dolines are gentle grassy hollows; others are rocky cliff-bounded basins. A distinction may be made between those formed mainly by direct solution of the limestone surface zone, solution dolines, and those formed by collapse over a cave, collapse dolines. . . . In America most dolines are referred to as sinks or sinkholes." (Monroe, 1970, p 7.)

This definition points to one problem in the terminology applied to limestone, or carbonate, karst as opposed to karst in evaporites. Dolines in the sense of hollows formed by direct dissolution of the surface are rare and may not exist on gypsum surfaces in southeastern New Mexico. Basin- or funnel-shaped hollows are common on gypsum terrain, but these are collapse sinks, or collapse dolines, in some stage of development or infilling. Dolines in the strict sense of dissolution on a surface are common on the Ogallala caliche caprock of the High Plains. Some of these are aligned southeasterly for many miles over that surface. Smaller clusters of less spectacular dolines are present in places on the Mescalero caliche, but these have not been observed to be aligned.

In this discussion the term doline will be reserved for these shallow, surficial dissolution hollows. The term "collapse sink" will be used for hollows formed by collapse of the surface over caves.

Collapse Sinks

Collapse sinks are closed depressions formed by the collapse of the roof of a cave (Monroe, 1970, p 6). No attempt was made to count the number of collapse sinks in the Pecos drainage system. They may be numbered in the many hundreds. Some are presently active; others may be termed dormant, "extinct," or "fossil" and are represented only by surficial depressions filled with broken fragments of exotic rock. In most "extinct" collapse sinks, dissolution of underlying rocks appears to have ceased locally. At these places dissolution may be impaired by the insoluble debris that fills the underground cavities.

Active collapse sinks are present in gypsum on the Gypsum Plain, in Nash Draw, on Burton Flat, and along the Pecos drainage northward from Carlsbad (Figures 13 and 14). Collapse sinks at many places terminate dry arroyos and could be classified as swallow holes, or stream sinks, during periods of heavy runoff. Some collapse sinks serve to increase the gradient of arroyos and are instrumental in erosional headward cutting. Annular scarps form at the surface around active collapse sinks (Figure 15).

Figure 13. Gypsum and dolomite, Rustler Formation, in a collapse sink, Burton Flat, Eddy County, NM. (Rocks at the surface have collapsed into a cave that was dissolved along a fracture system. The collapse sink is linear in plan and may be followed on the surface for ~100 m. See Figure 14.)

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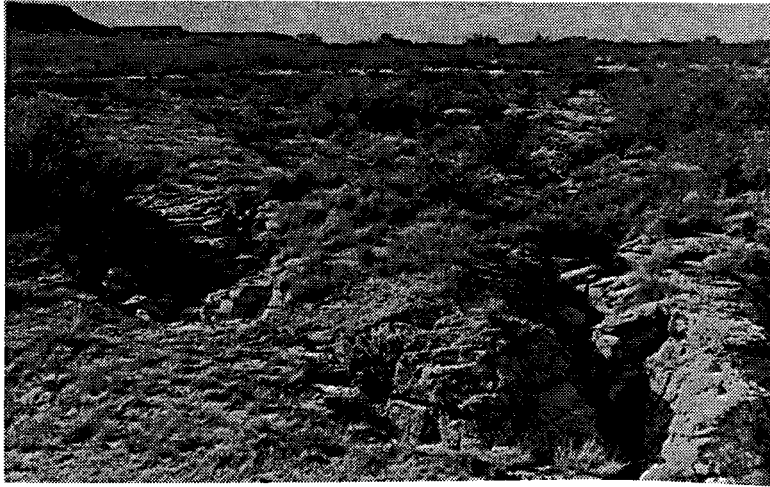
Figure 14. Collapse sinks along fractures, Nash Draw, Eddy County, NM. (Scarp in middle ground is on fan surface. See Figure 13). Livingston Ridge, capped by Mescalero caliche, is in background to left of the photograph).
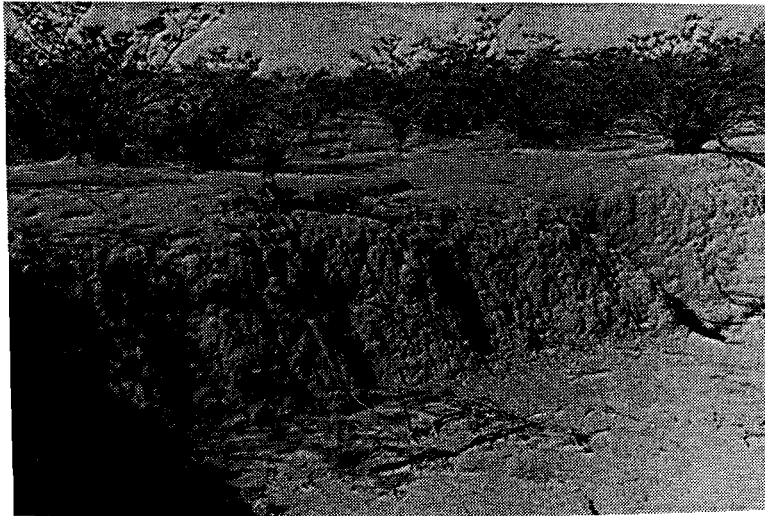
Figure 15. Scarp on a fan surface caused by subsidence around a collapse sink, Nash Draw. (The hammer rests on the downthrown side. Vegetation in the upper part of the photograph is mostly creosote bush (*Larrea*). See also Figure 13.)

The most common collapse sinks are small-scale features such as those in Nash Draw and Burton Flat. These have collapsed into cavities in underlying gypsum and are usually <50 ft (15 m) deep. San Simon Sink (Figure 16) in Lea County, NM, and Wink Sink in Winkler County, TX (Baumgardner et al., 1982) are examples of large-scale active collapse sinks. Both these sinks have their roots many hundreds of feet beneath the surface. They overlie the Capitan reef and are believed to result from dissolution of salt in the Salado Formation by water circulating in the reef aquifer. San Simon Sink is considered to be a modern analog of *breccia pipes*.
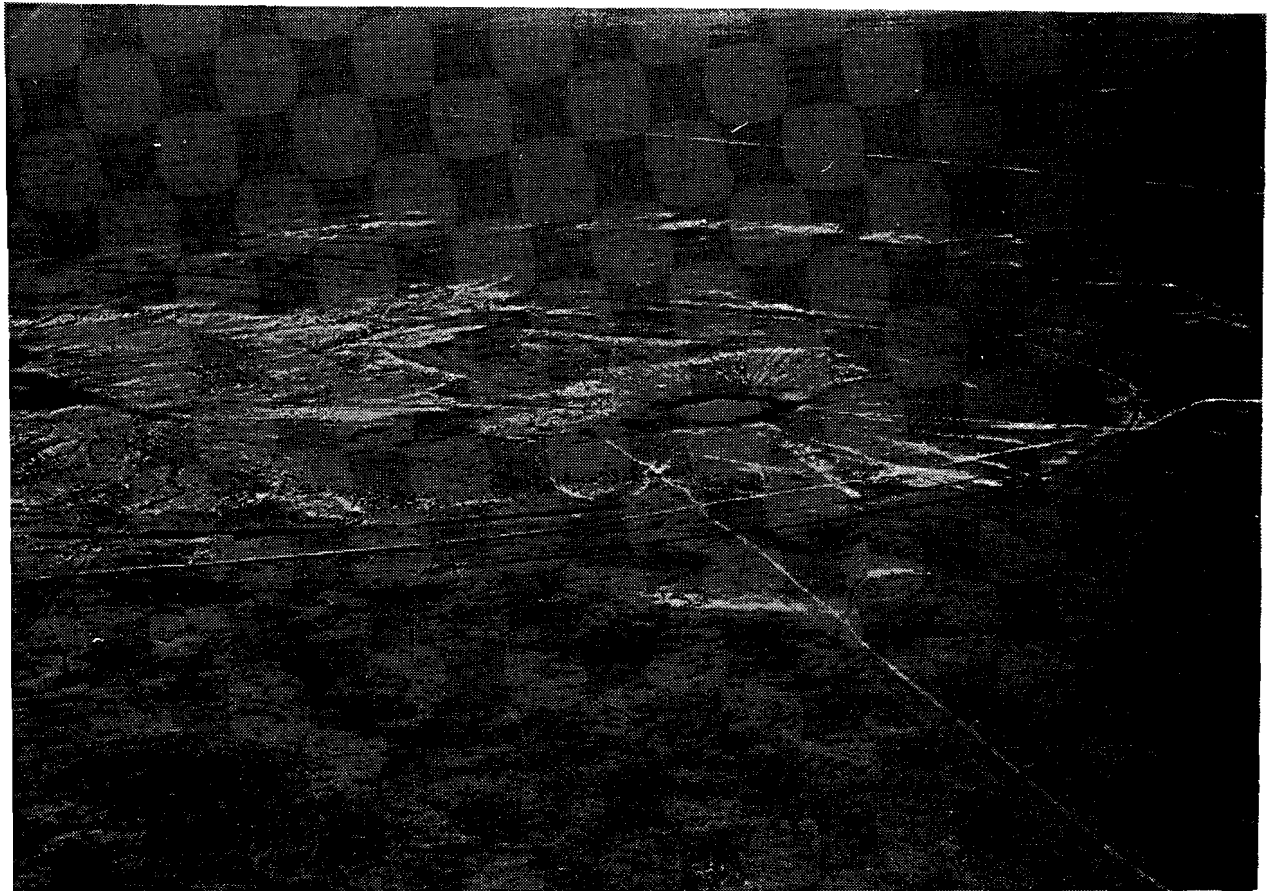
Figure 16. Aerial view of San Simon Sink, Lea County, NM. (Annular rings are visible cutting the surface around the sink. An exploratory hole (WIPP 15) was drilled to a depth of 810 ft (249 m) in the center of the sink. The upper 555 ft (168 m) consisted of Quaternary sands and clays. The lower 255 ft (78 m) was drilled in Triassic rocks).

A collapse sink on Livingston Ridge near the east side of Nash Draw was drilled (WIPP-33, Sec. 13, T.22S., R.30E.) to determine the nature of unusually thick sedimentary fill at that place and the internal structure of the sink. The hole was drilled to a depth of 840 ft (254.5 m) and penetrated the following stratigraphic units (Snyder and Mcintyre, 1981):

	Ft	(m)
Holocene sedimentary fill (sand and silt)	44	(13.5)
Dewey Lake Red Beds	357	(110)
Rustler Formation	276	(85)
Salado Formation	163	(50)

(Hole bottomed in upper part of Salado Formation)

Much of the anhydrite in the Rustler Formation had been hydrated to gypsum, and some gypsum had been dissolved from the formation. The Magenta and Culebra Dolomite Members of the Rustler were present but fractured and partially dissolved. Open cavities were encountered in the Rustler Formation.

The sink appears to be the result of collapse of the surface into the cavities in the Rustler Formation. The cavities are the result of dissolution of evaporites (anhydrite, gypsum, and halite) and dolomite by circulating groundwater in an open system. The cavities were extended towards the surface by vertical stoping. The Mescalero caliche is disrupted at this place, which indicates that the collapse occurred after the Mescalero formed. At present a small arroyo drains into the surface depression, and the sink is being infilled by surficial sand and silt.

Karst Valleys

Karst valleys were defined in Indiana by Malott (1939) as minor valleys that have been cut through clastic rocks into underlying beds of limestone. Karst features have developed on the valley floors, and much of the drainage is underground. "The change from surface to underground drainage is marked by 'dry beds' and swallow hole features, which receive and carry storm waters only."

Many valleys in southeastern New Mexico in gypsum are analogous to the limestone valleys described by Malott. The intermittent drainage of the semiarid climate of New Mexico serves to emphasize the periodic activity of the swallow holes. Collapse sinks and caves become swallow holes during heavy rain storms.

The history of these valleys can be deciphered at some places. Nash Draw in eastern Eddy County is a specific example. Nash Draw is a complex karst valley ~16 mi (26 km) long and ranges from ~4 mi (6 km) to 12 mi (19 km) wide. It is situated ~15 mi (24 km) southeast of Carlsbad. It has been studied intensively over more than two decades, first as geologic background for Project Gnome (a project involving the underground detonation of a nuclear device) (Vine, 1960); and, later, as geologic background for the WIPP site (Bachman, 1980 – 1981; Mercer, 1983). Consequently, the geology and history of Nash Draw are better understood than are most other karst features in the region.

Nash Draw began to form during Pleistocene time. During Gatuña time a tributary drainage system flowed southwesterly across the area now known as Livingston Ridge and Nash Draw toward the main stem of the ancestral Pecos drainage. The Gatuña stream had sufficient carrying power and turbulence to create beds of cross-laminated pebble conglomerate (Figure 8).

As this drainage system eroded into bedrock, it encountered the updip edge of the Rustler Formation (Figure 17). Dissolution began along the strike of the Rustler beds and initiated the formation of collapse sinks in the evaporites. These sinks were roughly aligned along the strike of the Rustler beds, which resulted in the present alignment of the central portion of Nash Draw.
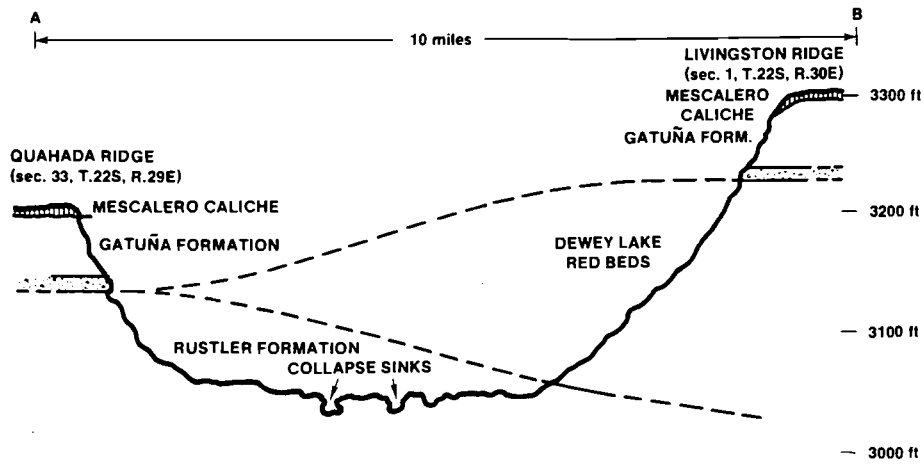
(a) Diagrammatic cross-section of Nash Draw showing stratigraphic relations of Gatuña Formation and evaporites in the Rustler Formation. (The vertical scale is greatly exaggerated).
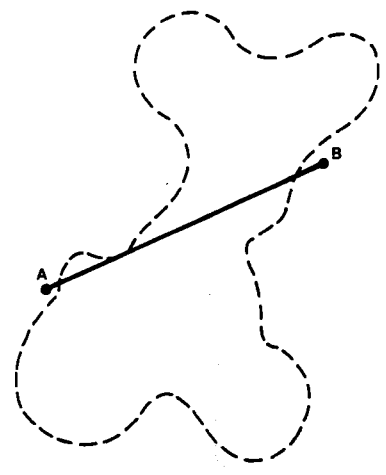
(b) Diagrammatic plan of Nash Draw showing location of cross section.

Figure 17. Diagrams of Nash Draw.

As the collapse sinks coalesced during the further development of Nash Draw, the Gatuña drainage system was disrupted. Today the eroded edges of stream gravels in the Gatuña are exposed on both sides of Nash Draw as much as 200 ft (61.5 m) above the floor of Nash Draw. This disruption occurred late in Gatuña time—after the fall of the Lava Creek B ash and before the Mescalero caliche was deposited. Thus Nash Draw was initiated as a karst valley ∼500,000 years ago and is a relatively young geologic feature.

Nash Draw assumed its present orientation as a result of the further coalescence of collapse sinks along the regional strike of evaporites in the Rustler Formation. Dozens of collapse sinks and caves are present in various parts of Nash Draw, and it continues to expand as a karst valley (Figures 18 and 19). Extensive headward cutting in the upper reaches of Nash Draw is the result of increased gradient by enlarging collapse sinks downstream in the floor of the draw. Steep-walled arroyos 20 to 25 ft (6 to 7.8 m) deep near the head of Nash Draw grade towards, and drain into, clusters of active collapse sinks in the central portion of the draw (SE 1/4 Sec. 33, T.21S., R.30E.). These interrelated processes of corrosion and corrasion are responsible for the continued expansion of Nash Draw.
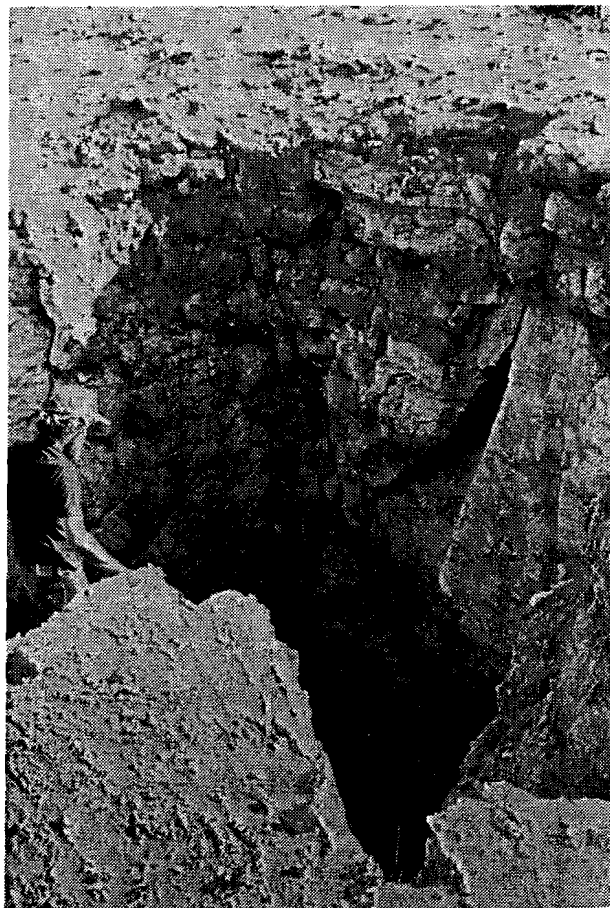
Figure 18. Collapse sink in upper part of the Rustler Formation, Nash Draw, Eddy County, NM, showing nearly vertical fractures in bedded gypsum. (More than 75 collapse sinks in various stages of formation have been observed in Nash Draw. Many of these sinks are entrances to caves.)

Figure 19. Cave in bottom of arroyo, Nash Draw, Eddy County, NM. (During heavy rainstorms, water drains into the cave and becomes a part of subsurface drainage. The arroyo appears to be the remnant of a collapse sink. Its walls are steep, and it terminates at the cave entrance. The height of the cave entrance is ~4 ft (1.2 m).)

Factors other than surface erosion may be responsible for the widening of Nash Draw. One factor is the development of collapse sinks along the margins of the valley floor above the level of the arroyos. These sinks do not appear to be controlled as much by surface erosion as by the action of groundwater on evaporites near the surface. A second factor includes the process of spring sapping; however, this process has apparently not been active since Late Pleistocene time. Spring deposits composed mainly of earthy gypsite carry a fauna of Pleistocene horse and camel on the eastern margin of Nash Draw (Sec. 15, T.22S., R.30E.). These deposits resulted from the dissolution of near-surface beds of gypsum and movement of groundwater to the surface. The gypsite was deposited by evaporation of the spring water.

The positions of individual springs are marked on the surface by low mounds of gypsite that are superimposed on sandstone of the Dewey Lake Red Beds and aligned parallel to the eastern margin of Nash Draw. Locally, at the alignment, there is a reversal of the regional eastward dip. Beds of the Dewey Lake dip westerly towards the axis of Nash Draw. It is presumed that this local reversal of dip was the result of collapse in the central part of Nash Draw. Tensional fractures along the crest of the reversal allowed groundwater to flow to the surface.

Swallow Holes

A swallow hole is defined as "a place where water disappears underground in a limestone region. A swallow hole generally implies water loss in a closed depression or blind valley; whereas a swallet may refer to water loss into alluvium at a streambed, even though there is no depression." (Monroe, 1970, p 17.) Swallow holes are conspicuous sumps in regions of prevalent perennial streams. In semiarid regions such as southeastern New Mexico, where most drainage is intermittent, swallow holes are visible only during rainstorms and heavy runoff. During dry periods the swallow holes appear as collapse sinks or as caves in the beds of dry arroyos.

In semiarid and arid regions, the concept of the swallet is of even less value. In these regions it is not uncommon for water to be lost entirely by evaporation in the alluvium of stream beds. At these places, runoff dries to a trickle and water disappears into the alluvium without involving karst processes. Instead of dissolving the bedrock, the water evaporates, and solutes in the water may be deposited in the alluvium as a cement.

Solution-Subsidence Troughs

Solution-subsidence troughs were described on the gypsum plain in western Texas and southeastern New Mexico (Olive, 1957). These features are defined as "straight narrow shallow surface depressions formed by the collapse of rock into subterranean caverns. . . . They are generally <20 ft deep, a few hundred feet to more than a mile wide, and a quarter of a mile to 10 miles long. Characteristically, they have relatively flat bottoms, which occupy about two-thirds of the trough . . . Caverns, sinks, fissures, and small shallow undrained depressions, all of solution origin, are common in areas of trough development" (Olive, 1957, pp 351-353).

Solution-subsidence troughs resemble poljes in form, except that the latter are mainly of tectonic origin and begin as karst features in down-warped, or down-faulted, blocks of limestone. Poljes are generally aligned in the direction of major tectonic trends and may be modified by solution and by later tectonic movements (Sweeting, 1973, pp 193-198). Solution-subsidence troughs resemble karst valleys but are developed entirely in gypsum bedrock and lack the clastic overburden that flanks karst valleys. Solution-subsidence troughs are represented by many unnamed valleys on the Gypsum Plain and elsewhere in southeastern New Mexico (Figure 20).
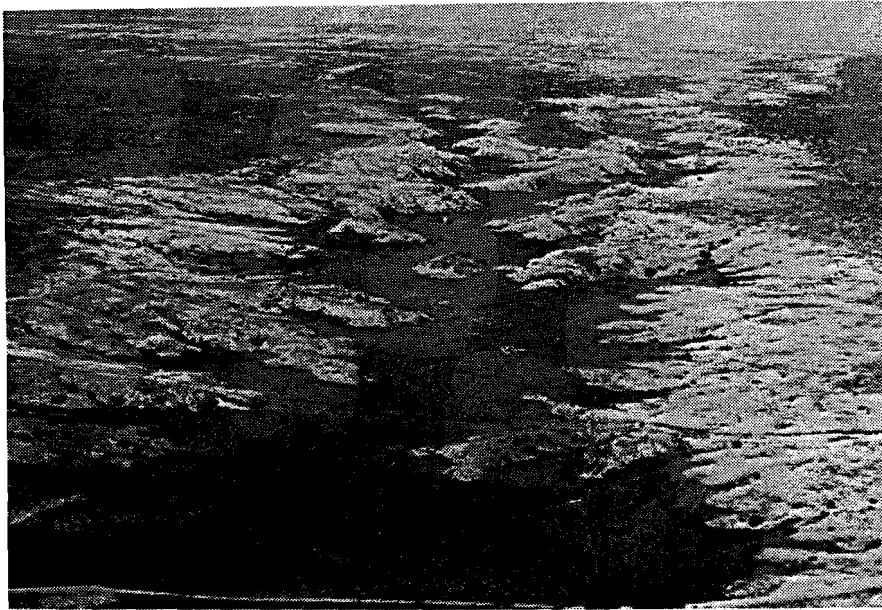

Figure 20. Aerial view of solution-subsidence trough, Gypsum Plain, southern Eddy County, NM. (This trough is ~2 mi (3.2 km) long and ~0.5 mi (0.8 km) wide at the widest part. The light areas on either side of the trough are massive gypsum of the Castile Formation.)

Blind Valleys

A blind valley is one that ends suddenly at the point where its stream disappears underground into a sink or cave. As water flows into the sink, turbulent whirlpools, swallow holes, or swallets may be created. Half-blind, or semiblind, valleys are blind valleys that overflow. In these valleys the sink cannot accept all the runoff during heavy rains, and water flows downstream on the surface.

In semiarid southeastern New Mexico, "blind arroyo" is a more descriptive term for these features. Blind arroyos are analogous to blind valleys, except that arroyos are small, deep, flat-bottomed gullies with intermittent flow occurring only after heavy rain. Usually a steep bank marks the termination of the arroyo at a collapse sink or cave (Figure 19). Blind arroyos are similar to half-blind valleys in that the discharge sink may overflow during heavy rain. However, in this event water merely backs up headward in the arroyo.

Headward cutting in the arroyo may occur as its gradient increases by dissolution of evaporites in the sink. At other times sinks may become filled with sediment as a result of erosion within the arroyo. Runoff is then diverted to a new escape route in the subsurface, or, by completely choking off the underground passages, water may flow for a time on the surface. When karstic features are infilled, their groundwater system is reorganized. These features may no longer contribute to the recharge of the regional groundwater system. This is part of the process described by Lee (1925) as "erosion by solution and fill."

Many karst valleys, as well as solution-subsidence troughs, are also blind arroyos. Remuda Basin in eastern Eddy County is a blind arroyo that terminates abruptly in a cave in gypsum in the Rustler Formation. Flood debris is visible at high-water marks in the arroyo where water has overflowed from the discharge sink.

Karst Plains

Karst plains are regions of horizontal, or nearly horizontal, strata on which small closed depressions, collapse sinks, subterranean drainage, and other karst features are developed. Major surface drainage cannot cross these plains because the sinks are very effective in collecting runoff. The major drainage is subterranean, and surface drainage is local.

Burton Flat northeast of Carlsbad is an example of a karst plain in evaporites. Major surficial drainage does not escape from this area. The Rustler Formation makes up the bedrock, and collapse sinks pit the surface and absorb the runoff. No attempt was made during this study to count individual collapse sinks. Some sinks are compound and aligned along fractures, with the result that a census would be misleading. It is estimated that collapse sinks on Burton Flat may be numbered in the many tens.

Some collapse sinks on Burton Flat are paleokarst features. One deposit of collapse breccia in the southern part of Burton Flat (Secs. 25-26, T.20S., R.28E.) includes Triassic conglomerate, Gatuña shales and sandstone, and is partially engulfed by Mescalero caliche. The Triassic rocks represent an outlier that is ~10 mi (16 km) west of the nearest Triassic exposures. Erosion removed the Triassic between the two localities after Triassic time. The presence of Gatuña rocks in the collapse debris suggests that the erosion may have occurred during Gatuña time. The presence of the Mescalero caliche as a cement across the surface indicates that the collapse occurred before Mescalero time. This feature is presumed to be one of several eroded breccia pipes, which are discussed below.

Minor Features

The most common small features associated with karst in evaporites are products of both erosion (corrasion) and dissolution (corrosion). These features include solution runnels and grooves.

Solution runnels are very common on the surface of exposed gypsum in southeastern New Mexico. They are elongate, parallel grooves that vary upwards of 10 mm deep, 10 to 25 mm wide, and may be as

much as 150 mm long. They occur on inclined surfaces and indicate the direction of flow of surface run-off. They are present on the surface of exposed gypsum in the walls of arroyos or at entrances to caves (Figure 6). Some are present on slightly inclined blocks of gypsum where rain washes across the surface.

Runnels are rare on gypsum in the floors of arroyos, but larger grooves are common. These are as much as 4 in. ($\sim$10 cm) to 12 in. ($\sim$30 cm) deep and of similar width. One of these grooves may mark the center of the channel and the direction of flow of the arroyo.

Scallops are irregular concavities carved in the surface of soluble rocks. They are common in regions of carbonate karst but are rare in the gypsum karst of southeastern New Mexico. Poorly developed scallops were observed in the floor of a small arroyo on Burton Flat (Figure 21).

Figure 21. Scallops on the top surface of gypsum in the Rustler Formation, Burton Flat, Eddy County, NM. (At this locality the beds are nearly horizontal.)

Solution runnels and grooves indicate the character of water flow. "The distance between their crests is inversely proportional to the flow velocity. They are the consequence of the interaction of fluid flow and rate of dissolution of a soluble surface." (Sweeting, 1973, p 140.) Flutes and scallops in caves are of interest to the study of flow velocities in vadose zones, but where they occur on the surface in gypsum karst they are of less significance. The flutes on the surface exposures of gypsum indicate dissolution by surface runoff during rainstorms. They indicate relatively gentle processes. The large-scale grooves in gypsum in the floors of arroyos are products of the raging torrents that fill these arroyos during intermittent flooding.

Karst mounds are erosional remnants of dissolution breccia (Figure 22). They are especially common along the east side of the Pecos River southward from Malaga.
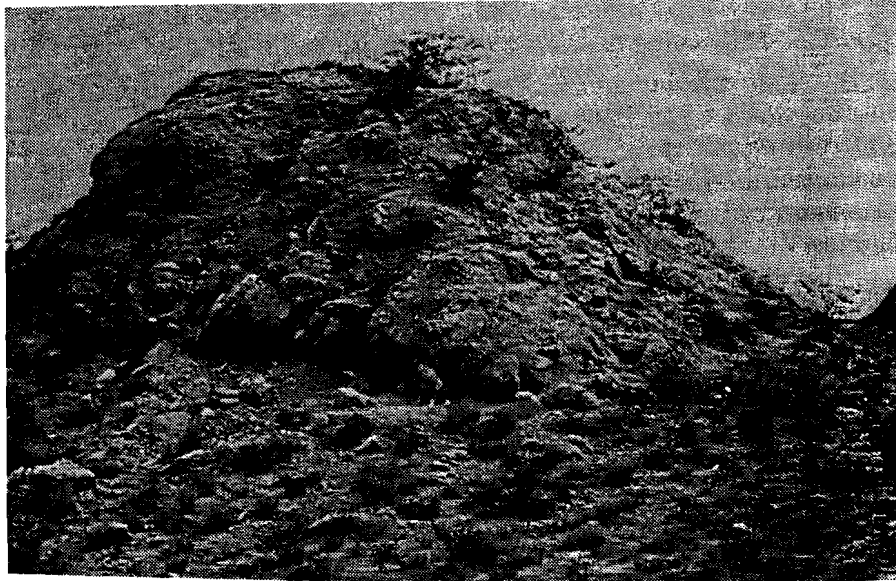
Figure 22. Karst mound. (Erosional remnant of chaotic blocks of gypsum and dolomite, intermingled with silt and clay. Mounds vary in form from conical to ovate. They are common features along the banks of the Pecos River in the southern part of Eddy County, NM, where the Rustler Formation is exposed. The mound in this photograph is ~20 ft (6 m) high and is one of several distinctly visible on aerial photographs.)

Underground Karst

If the matter of scale is disregarded, most underground karst may be classified as caves or parts of caves. Although some caves are in the phreatic zone below the water table, it is presumed that most caves have some hydrologic connection to the surface, which allows for the infiltration of surface water into the cave system. The connection to the surface causes some portion of the cave system to be classified as surficial karst. However, in this discussion features that appear to have in part a deep-seated origin, presumably within the phreatic zone, are classified as underground karst.

Caves

In the classification of caves, Davis (1930) considered that caves and cave formations are part of the cycle of erosion and result from subsurface hydrologic conditions. He recognized that caverns might form below (phreatic) or above (vadose) the water table. Bretz (1942) expanded and refined the concept of phreatic and vadose caverns by describing specific features and establishing these as criteria for recognizing the two types. Later studies have emphasized the formation of caves at or near the water table itself, which was advocated earlier by Swinnerton (1932).

These classifications and theories of cave origin are based mostly on observations of caves in relatively flat-lying limestone. Many well-known caves in southeastern New Mexico are in limestones in the Guadalupe Mountains, but the caves considered here are in relatively flat-lying gypsum. The history of these gypsum caves resembles that of caves in limestone in some particulars.

Caves in gypsum in Oklahoma and Texas have been observed to possess characteristics of both phreatic and vadose origin (Bretz, 1952; McGregor et al., 1963). In Alabaster Cave in Oklahoma, Bretz observed that vadose features are superimposed on phreatic features. The phreatic features include dome-shaped solution cavities in the ceilings, selenite plating on the walls, and dissolution features in the main chambers. The vadose features are more varied and include collapse of walls and ceilings, channeling of the floors and the ceiling above detrital debris, and fluting on the wall rock.

Caves in gypsum differ somewhat from those in limestone in that they may be formed and enlarged more commonly in the vadose zone than in the phreatic zone. Infiltration of surface water to underground systems resulting in the dissolution of gypsum is controlled by features as small as joint sets and as large as collapse sinks. Abundant joints provide permeability. Joints themselves are widened by dissolution (Lattman and Olive, 1955) and control the distribution of collapse sinks on the Gypsum Plain, Burton Flat, and at other places (Figures 9 and 10).

As mentioned above, in blind valleys flood waters flow underground into collapse sinks. The water introduced into collapse sinks in this manner is relatively fresh, but it may carry a sediment load that serves as a corrasive agent. Thus, it enlarges underground passages both by erosion and dissolution. Caves are formed and enlarged in drainage systems by these processes (Figure 19).

In limestone caves, deposition of carbonates in various forms by evaporation of solutions or loss of carbon dioxide in solutions gives rise to secondary deposits. These include familiar cave formations such as stalactites and stalagmites, collectively known as speleothems. Secondary formations are deposited in gypsum caves only by evaporation and are much less varied and less spectacular than in limestone caves. The small features deposited on the walls of caves and on the face of gypsum exposures in evaporite karst include crusty or filamentous forms such as "cave blisters" and "cave flowers" (Moore and Sullivan, 1978, pp 60-61). Weak stalactites are present as crusts. None of these features were observed during the present study.

Gypsum caves owe much of their existence to fracturing of bedrock and by their very nature are smaller and more ephemeral than caves in limestone. Fractures reduce the supporting walls and ceilings of gypsum caves to masses of unstable blocks that are more susceptible to rock falls than are the walls of limestone caves, which may even be reinforced by secondary cementation. Reams (1964) observed that ceilings and walls near entrances of gypsum caves in semiarid regions are enlarged by spalling.

Caves are numerous in southeastern New Mexico where gypsum is at or near the surface; but, except for some of the more accessible caves, relatively few have been explored. Smith (1969, 1971) reported more than 130 caves in 100 square miles on the Gypsum Plain. The caves in the vicinity of Chosa Draw are among the better known in the region. Eight caves totaling more than 5.8 km in length have been mapped in that area. These include a resurgence cave (Figure 23) and Parks Ranch Cave, which is 3730 m long (Sares, 1984, p 51). Border Cave and Wiggley Cave in Culberson County, TX, terminate in large rooms that are partially filled with water. The room in the latter cave is about 130 ft (40 m) below the surface (Smith, 1971).

Breccia Pipes

Breccia pipes are assumed to be the roots of deep-seated collapse sinks. Where highly eroded, the topographic expressions of such pipes are dome-like structures forming hills. These rise 50 to 100 ft (15 to 30 m) above the surrounding terrain. They are circular and ~1150 to 1200 ft (350 to 370 m) in diameter (Figure 24). They are characterized by brecciated cores composed of rock that is younger stratigraphically than the adjacent wall rock. The brecciated core is separated from the wall rock by a sharp faulted boundary. These unique domal features were first recognized in southeastern New Mexico by Vine (1960).
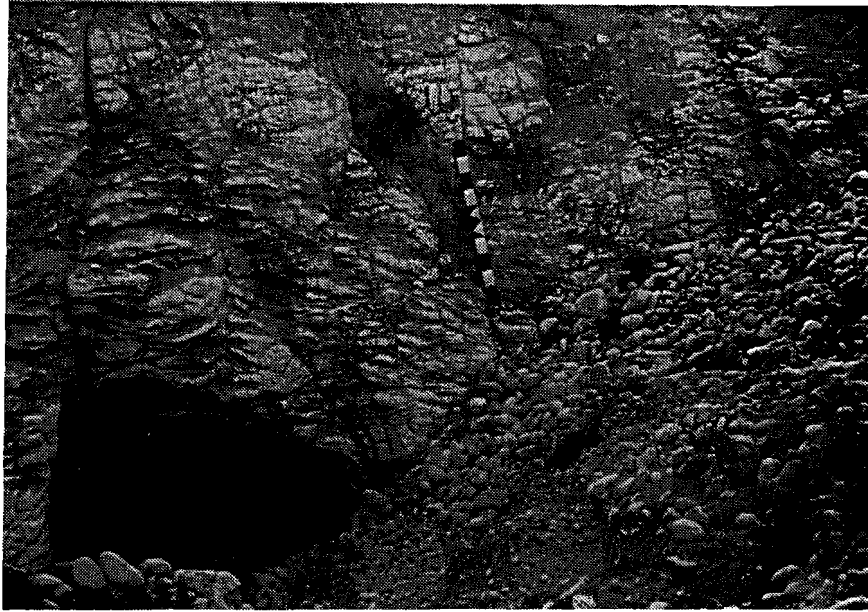
Figure 23. Entrance to resurgence cave, Chosa Draw, Eddy County, NM. (During periods of heavy runoff, water enters a passage upstream and flows underground to this cave where it emerges with enough velocity to tumble cobbles of gypsum as much as 4 to 5 m above the floor of the cave. (Sares, 1984, estimates discharges to 11.4 m^3/s on the basis of scallop measurements.) The cobbles are elongate and reach diameters of 30 cm. The scale is 1 m.)

Figure 24. Aerial view of Hill C, the topographic expression of a breccia pipe in eastern Eddy County, NM (W1/2 Sec. 5, T.21S., R.29E.). (The road in the lower right-hand corner of the photograph leads to the drill pad inside the breccia pipe where an exploratory hole (WIPP-16) was bored in breccia to a depth of 1800 ft (550 m). The arroyos that cut the surface in the background dissect alluvium to depths of 25 ft (7.7 m) in places. The thick alluvium in this area is believed to be evidence for subsidence of the surface surrounding Hill C.)

It is apparent that these breccia pipes are the result of dissolution and removal of underlying soluble rocks (Anderson, 1981; Bachman, 1980; Snyder and Gard, 1982). Although the various processes that form collapse sinks are understood, the processes and sequence of events that formed breccia pipes and their associated domal structures are less well known. The potential for the brecciated pipes to act as channelways for groundwater that would endanger the integrity of evaporites in the subsurface has prompted extensive studies. These studies include detailed surface mapping (Bachman, 1980), geophysical studies, and drilling (Snyder and Gard, 1982). This discussion includes a summary of these studies.

Hills A, B, and C—Three breccia pipes, designated Domes "A," "B," and "C" by Vine (1960), have been studied intensively. Vine's designation has been followed in later work, except that the domes are termed "Hills" A, B, and C.

Hill A is ~18.5 mi (30 km) east of Carlsbad, Eddy County, NM (SW 1/4 Sec. 35, T.20S., R.30E.). It is a low, circular breached hill ~1200 ft (370 m) in diameter and rises ~40 to 50 ft (12 to 15 m) above the surrounding terrain. The center of the hill has been eroded to a shallow basin by an arroyo system cutting headward from the west. The flanks of the hill are covered by the Mescalero caliche, which dips ~15° away from the rim (Figure 25).
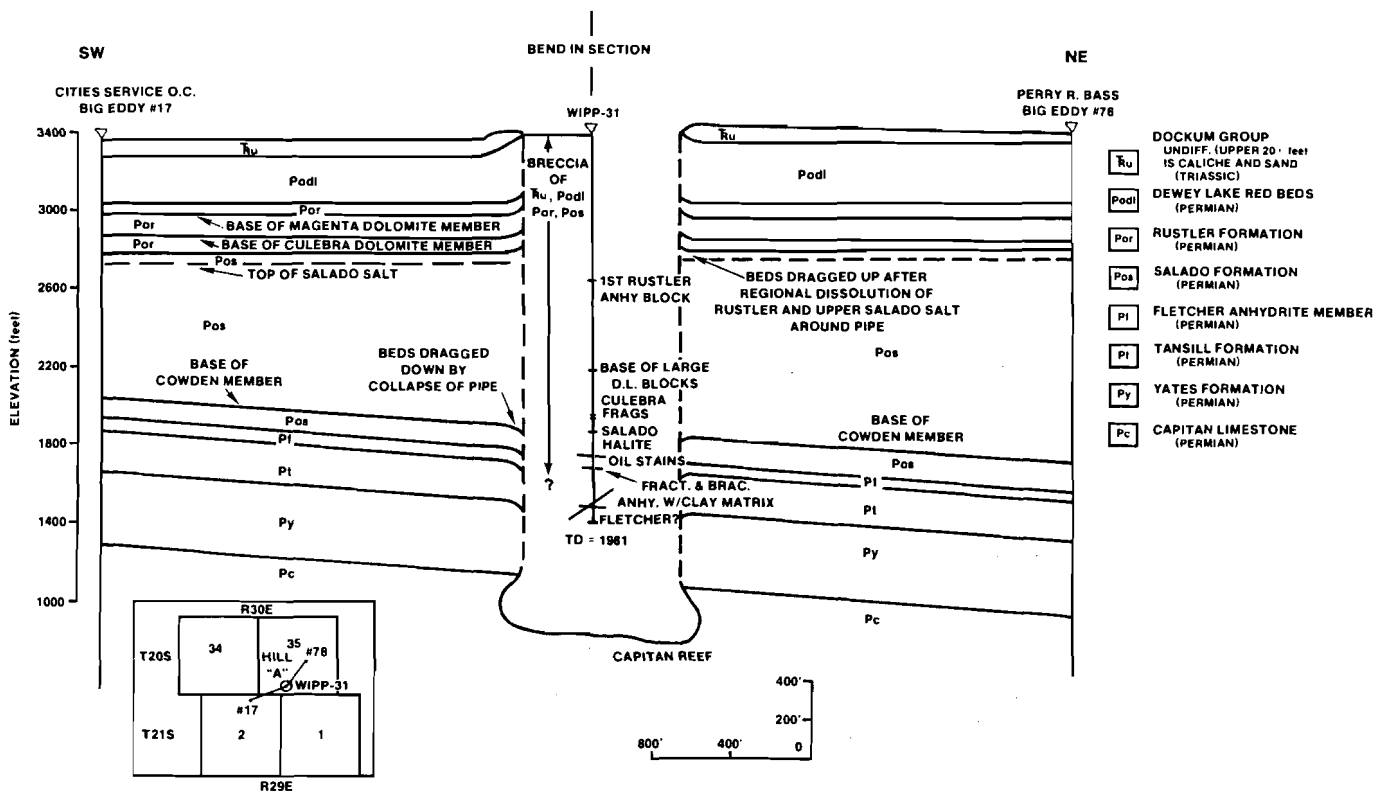
Figure 25. Southwest to northeast cross section through Hill A (after Snyder and Gard, 1982, Figure 10)

Dewey Lake Red Beds are exposed in the upturned walls of the dome on the east, north, and west sides. Average dips of these beds are ~15° away from the center of the dome, but locally dips steepen to 20° to 22°. Rocks of the Triassic Dockum Group rest on the Dewey Lake along the east and south sides of the basin.

A "peripheral fault or ring-fault" (Vine, 1960, p 1905) cuts the wall rock. A breccia consisting of blocks of Triassic clay, sandstone, and conglomerate fills the core inside the peripheral fault. The brecciated debris ranges in size from clay to large angular blocks 4 to 5 m across.

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The brecciated core is partially covered by alluvium, and a minor channel on the north side of Hill A is filled with gravel of the Gatuña Formation. This fill is ~1.5 ft (0.5 m) thick and includes pebbles of pisolitic caliche similar to the Ogallala caprock that is exposed along Mescalero Ridge ~30 mi (48 km) to the east. A structureless caliche with abundant root casts overlies the Gatuña gravel. The caliche is laterally continuous with the Mescalero and is presumed to have been deposited contemporaneously with at least part of the Mescalero in a boggy depression.

The Gatuña gravel and the overlying structureless caliche, along with the Mescalero caliche, furnish clues for deciphering the history of Hill A. None of these rock types have been observed as components of the collapse breccia; therefore it is assumed that the collapse structure developed before the Gatuña was deposited in the shallow channel. The Gatuña and the structureless caliche were deposited in a moist depression that formed as a result of the collapse of the core of Hill A.

This interpretation of the central part of Hill A as laid down in a depositional basin requires topographic conditions much different from those at present. The Gatuña gravel is now at a higher elevation than the surrounding terrain, which requires differential movement since Gatuña time. The structure and morphology of the Mescalero caliche indicate that the flanks of the dome have tilted since deposition of the caliche. Columnar pedogenic structures, normally vertical, at the base of the Mescalero are tilted ~15° from the vertical on the east side of the dome. The tilting is away from the peripheral fault, which indicates either (1) the surrounding area has subsided since deposition of the Mescalero caliche or (2) the central core of Hill A has been uplifted. In addition, the normal pedogenic morphology of the Mescalero on the flanks of Hill A is disrupted by a network of pipes and fracture fillings. These suggest dissolution and readjustments of the caliche during, or after, its deposition.

A hole (WIPP 31) has been drilled in the center of Hill A to a depth of 1981 ft (604 m). Brecciated and steeply dipping rocks were encountered throughout the drilling. The breccia is a chaotic mixture of Triassic rocks, Dewey Lake Red Beds, Rustler Formation, and residues of the Salado Formation. A thick bed of anhydrite was encountered at a depth of 1903 ft (580 m). Laminae in the anhydrite dip as much as 50°. The anhydrite has been identified as the Fletcher Anhydrite (Snyder and Gard, 1982, p 21), the basal member of the Salado Formation.

Vine (1960) recognized that Hill A had been subjected to differential movement. He suggested that the movement could have been arching caused by hydration (and expansion) of anhydrite to form gypsum, intrusive flow of salt in the subsurface, or differential solution. Records of the drill hole (WIPP 31) indicate that there has been no hydration of anhydrite and that salt in the Salado and Rustler has been dissolved and removed from the core of Hill A. It is more probable that the area surrounding Hill A has subsided as a result of dissolution of evaporites in the Rustler Formation surrounding the hill. The brecciated core of Hill A is composed mainly of insoluble residues and is less susceptible to dissolution. The hill itself was left standing above the surrounding area.

Hill C is a structural dome ~15 mi (24 km) southeast of Hill A. It rises ~100 ft (30 m) above the surrounding landscape and is ~1150 ft (350 m) in diameter. It is breached in several places, but breaching is not as extensive as in Hill A. The Mescalero caliche covers the flanks of the hill and engulfs rocks of the Gatuña Formation, Dewey Lake Red Beds, and Triassic Dockum Group. Some of the Mescalero caliche that formed the caprock is preserved.

The Triassic rocks are brecciated and collapsed against wall rock of Dewey Lake Red Beds that dip away from the brecciated core. A segment of a possible ring fault is exposed in an arroyo on the western edge of the hill. Neither the Gatuña Formation nor the Mescalero caliche is incorporated in the brecciated core. Minor fractures offset the Mescalero caliche as much as 10 to 13 ft (3 to 4 m) in places, but these fractures are linear and are interpreted as resulting from the readjustment of the brecciated mass following deposition of the caliche. It is probable that similar fractures existed over the top of Hill A before the caprock was removed by erosion.

Hill C is situated above underground workings in the Mississippi Chemical Company potash mine. The brecciated core as well as the wall rock were encountered in mine workings ~1200 ft (366 m) below the surface. At the level of these workings, beds dip downward towards the brecciated core, thus reversing the structural relationships at the surface.

A hole was drilled in the center of Hill C (WIPP-16) to a depth of 1300 ft (400 m). Records of this drill hole and associated exploration in the mine workings have been discussed by Snyder and Gard (1982, p 28-55). They stated that the drill hole "penetrated brecciated rock of the Triassic Dockum Group, ... Dewey Lake Red Beds, and part of the Rustler Formation. Although the Rustler has been downdropped and shattered, the beds, unlike the overlying rocks, were in recognizable stratigraphic order. The contact of the Rustler and the overlying Dewey Lake has been downdropped ~189 m (620 ft). The Culebra was cored in WIPP-16, and this differs markedly from drill hole WIPP-31 at Hill A where no halite and no recognizable sequence of rock was found to represent the Rustler" (Snyder and Gard, 1982, p 31). They concluded that collapse in the pipes at Hill C and Hill A occurred at widely spaced times. On the basis of compaction data they calculated that collapse at Hill C was at a time when the overlying Dockum Group had not been deeply eroded and was almost twice as thick as now represented in adjacent areas.

A factor in the subsidence of the terrain surrounding the hill has been near-surface dissolution, suggested by strata dipping away from the core at the surface. Strata dip inward towards the brecciated core in the subsurface at Hill C, representing normal drag along the fracture. Dissolution has been restricted to the upper part of the stratigraphic section (evaporites in the Rustler Formation) since the collapse and is reflected in the reversal of dips.

Other assumed breccia pipes in the northern part of the Delaware basin include Hill B, which is south of, and adjacent to, Hill A. It is a structural dome rising 93 ft (28 m) above the surrounding landscape. It is capped by Mescalero caliche that has been only slightly eroded on the west and south sides, where some brecciated Triassic rocks are exposed. It is assumed that Hill B is a breccia pipe similar to Hill A.

At three other localities along the northern edge of the Delaware Basin there are probable breccia pipes. These include the "Wills-Weaver pipe" (Sec. 12, T.20S., R.29E.) where a drill hole penetrated 821 ft of brecciated rock (Snyder and Gard, 1982, p 55). Brecciated Triassic rocks associated with gravel of the Gatuña Formation crop out in a circular area at least 300 ft (92 m) in diameter near the workings of the Potash Company of America Mine (NW 1/4 Sec. 4, T.20S., R.30E.). This feature is probably a breccia pipe that has been deeply eroded. A similar occurrence of brecciated Triassic rocks associated with Gatuña gravels is present in a circular feature ~3000 ft (920 m) in diameter ~10 mi (16 km) northeast of Carlsbad (Sec. 25, T.20S., R.28E.). This feature resembles a breccia pipe at the surface, but holes bored for petroleum in the vicinity indicate that it is not deep-seated. It is presumed to be the remnants of an eroded collapse sink.

Origin of Breccia Pipes—It is evident that the breccia core at Hill A collapsed into a large underground cavity. However, the collapse occurred in past geologic time, and the processes that caused the collapse are less evident. Questions requiring explanation include the following:

- The process that formed the original underground cavity
- The reason for the distribution of apparent breccia pipes along the northern margin of the Delaware Basin
- The cause of the cylindrical form of known breccia pipes through more than 1000 ft (300 m) of stratigraphic section
- The time of collapse in the geologic past
- The occurrence of modern analogs.

The following discussion is an attempt to address these questions.

Breccia pipes are here considered to be a variety of ancient, large-scale collapse sinks. Collapse sinks result from simultaneous dissolution and subsidence, or from collapse of surface rocks and soil into subsurface voids. The voids into which collapse occurred to form breccia pipes are presumed to have formed from dissolution deep in the zone of saturation.

Drilling at Hills A and C, and underground workings in the Mississippi Chemical potash mine where mine workings encountered the brecciated core of Hill C, have proved that the cores of known breccia pipes are more deeply seated than is usual in collapse sinks. The brecciated core of Hill A is continuous to a depth of at least 1981 ft (609.5 m). The drilling record indicates that the Salado Formation was present at that depth. On the basis of projection and comparison with nearby drill holes, the base of the Salado should be encountered at a depth of ~1700 ft (523 m) at Hill A. This discrepancy represents a displacement of >280 ft (86 m) of rocks within the pipe.

In the vicinity of Hill A, the Salado Formation rests on, and laps across, the Capitan reef and interfingering back reef units of the Artesia Group in the subsurface. The Capitan and its adjacent back reef rock units are the geologic setting for the Carlsbad Caverns and for many other caves in the nearby Guadalupe Mountains.

Evidence in Carlsbad Caverns indicates that the dissolution that formed the cavern system occurred at, or just below, the water table (Gale, *in* Hayes, 1957). Dissolution in Carlsbad Caverns, and probably all the caves of the area, began along joint sets and took place before uplift of the Guadalupe Mountains and development of the present erosion surface (Bretz, 1949; Hayes, 1964). Most of the dissolution occurred during the Tertiary Period. "The late Pliocene or early Pleistocene uplift of the Guadalupe Mountains caused a lowering of the water table, and the vadose cycle of carbonate precipitation began." (Hayes, 1964, p 50.) The spectacular formations in Carlsbad Caverns began to be deposited, and are continuing, in this vadose cycle.

Carlsbad Caverns underlie an area nearly 1 mi long and 1/2 mi wide. They have a vertical range of 1025 ft. Other caves in the Guadalupe Mountain block range from <100 ft (31 m) to >1000 ft (310 m) long and have vertical ranges through at least 250 ft (77 m) (Bretz, 1949; Hayes, 1964, p 51). Caves occur in the San Andres Limestone, Capitan Limestone, and in the Seven Rivers, Yates, and Tansill Formations. Caves are found in the Guadalupe Mountains from the Capitan cliffs along the front of the range to the back reef area across a belt >18 mi (29 km) wide.

The Capitan Limestone, and its associated carbonate facies in the Artesia Group, is the major aquifer along the northern margin of the Delaware Basin (Mercer, 1983, pp 34-38). This indicates that these stratigraphic units are within the phreatic zone. This is a similar geologic setting to that in which Carlsbad Caverns and other caves in the Guadalupe Mountains were formed; therefore, dissolution along joint sets and the formation of caves of considerable dimensions in this zone are predictable. It is presumed that the breccia core at Hill A collapsed into one of these cavern systems.

Breccia pipes appear to maintain a cylindrical form to great depths. This form suggests that the channel into which the breccia collapsed was limited to a small radius, probably the intersection of the joint sets, and dissolved by a unique process. The process of brine density flow (Anderson and Kirkland, 1980) is one possible explanation for the dissolution of evaporites in a vertical column. According to this hypothesis, unsaturated water from the Capitan aquifer could rise through fractures from a hydrostatic head. Dissolution of salt increases the water density, which causes the downward flow of brine. This initiates a flow cycle that dissolves chambers in the salt. Collapse of these chambers results in the formation of breccia pipes.

This hypothesis outlines a possible mechanism for the formation of the breccia pipes at Hills A and C. During Pleistocene time when these pipes were formed, effective rainfall and runoff were much greater, and the regional drainage was at higher elevations than at present. Variations in the porosity and permeability of the limestone aquifer system underlying the area of the breccia pipes contribute to modification of the flow of groundwater (Motts, 1968; Hiss, 1975). These factors would have resulted in elevated potentiometric surfaces and increased hydraulic heads. About 2 mi (3.2 km) north of Carlsbad, collapsed blocks of stream deposits in the Gatuña Formation are at least 50 ft (15 m) above the present level of the Pecos River. In Pierce Canyon, channel gravels are >100 ft (30 m) above the channel of the Pecos. These relationships indicate that the minimum altitude of the drainage system was much increased during Pleistocene time. The resulting increase in hydraulic head may have been sufficient to initiate the stoping process required to dissolve chambers in salt as outlined above.

The formation of vertical shafts in limestone caves may be a more likely process that initiates the formation of the columns of breccia in breccia pipes. Vertical shafts cutting upward from horizontal cave passages form late in the history of a cave as the water table is lowered. The shafts form by dissolution of joint sets, range from 3 to 30 ft (1 to 10 m) in diameter, and may extend vertically >160 ft (50 m). They may underlie the heads of streams or collapse sinks. The major portion of these features forms in the vadose zone where water flows down the walls of the shaft, dissolving and enlarging the shaft with time to form dome pits (Pohl, 1955; Moore and Sullivan, 1978, pp 22-23). Such shafts may have formed in caves in the subsurface carbonate rocks and stoped their channels to the surface to form pathways for the collapse of breccia pipes.

The breccia pipes along the northern margin of the Delaware Basin are paleokarst features at least as old as mid-Pleistocene. Gatuña gravels on the surface of the breccia at several places, and Mescalero caliche that engulfs the breccia at Hill C, indicate a minimum age of 500,000 to 600,000 years for these features. The presence of Triassic debris within the breccia itself indicates that erosion has removed rocks of Triassic age from the vicinity of the breccia pipes for a distance of at least 1 mi (1.6 km) in the case of Hill A. If the brecciated feature 10 mi northeast of Carlsbad proves to be a breccia pipe, Triassic rocks have been eroded away from that area by a distance of 11 mi (17.6 km) since collapse occurred. Considerable geologic time is required to accomplish erosion of such magnitude.

The climatic regime during some intervals of Middle and Late Pleistocene time was much different from the present. Water surplus was greater, streams were more erosive and had greater carrying power, and stream systems followed different channels. Those past conditions dictate different groundwater conditions. Rates of dissolution were accelerated, and the landscape was modified more rapidly under those rigorous conditions.

Although the breccia pipes may have collapsed during unusual hydrologic conditions, modern analogs of breccia pipes may be present in other parts of southeastern New Mexico. San Simon Sink (Figure 16) is an active collapse sink that may have its roots in the phreatic zone of the underlying Capitan Limestone (Lambert, 1983, pp 42-43, 82; Bachman, 1984). It is probably an example of a "breccia pipe" in the early stages of formation. The feature will not appear as a "pipe" until the underground cavity is filled with breccia and the surface modified by some future cycle of erosion and dissolution.

Breccia pipes along the northern margin of the Delaware Basin may be little more than ancient, large-scale, deep-seated collapse sinks. Similar features in various stages of development have been reported in various parts of the world. Pipe-like sinkholes, termed "fossil penetration pipes," have been described in West Germany (Prinz, 1973). Some of these are as much as 325 ft (100 m) in diameter and penetrate to depths of 640 to 975 ft (200 to 300 m). They are presumed to maintain vertical walls even through bedded salt. Other breccia-filled pipes in nearby areas are presumed to penetrate strata to depths of 2925 ft (900 m) (Grimm and Lepper, 1973). Salt dissolution in these areas has been active from Late Tertiary to the present. The process of dissolution in these German examples has been described as "suberosion" which presumably includes the process of stoping—the upward migration of cavities. Miotke (1971) has described steep-walled collapse features that have their roots in cavities in bedded gypsum.

Shallow collapse features in Canada have some characteristics in common with breccia pipes. Beds of salt and associated evaporites rest on carbonate of Devonian age. Carbonate mounds as much as 325 ft (100 m) high and 0.6 to 5 mi (1 to 8 km) long rise above the main body of the deposit. During Devonian time, groundwater circulated through the reef-like mounds and dissolved portions of the overlying salt beds. Many of these features are in the subsurface and are known only through geophysical studies. Other collapse features that are the result of dissolution of salt are visible at the surface. Some of these episodes of collapse have been dated at 13,600 years ago (Christiansen, 1971).

Chimneys of cemented breccia have been exposed by differential erosion in northern Michigan. Some of these columns of breccia are continuous into the subsurface, with near-vertical walls. Some breccias may extend into the subsurface 1400 to 1500 ft (430 to 460 m) (Landes et al., 1945). Dissolution of salt and associated evaporites, with collapse into the subsequent cavities, is believed to be the origin of the breccia chimneys.

Shafts called jamas, which characterize deep limestone karst in the Dinaric Alps (Yugoslavia) are comparable to breccia pipes in dimensions and may be analogous to an early stage of breccia pipe formation. "The jama is the surface reflection of a deep ramified fissure system in the deep karst. Caves . . . are more typical of areas where the limestones are thinner, where shallow connecting channels have developed and where more abundant waste and alluvium have prevented the water from sinking into greater depths . . . Hence, jamas are associated with deep karst, while caves are typical of more shallow karst." (Sweeting, 1973, pp 240-241.)

Crveno Jezero is a jama on the east side of Imotski polje. It is "an immense collapsed abyss with its highest rim at over 520 m (1690 ft) above sea level; its diameter is about 400 m (1300 ft); at its deepest measured part the bottom is only 4.1 m (13.3 ft) above sea level, and so it has a relative depth of at least 500 m (1625 ft) from the highest point of its rim. . . . The bottom is uneven and it is probable that it is even deeper. The lower part of Crveno Jezero is filled with water, which periodically changes its level." (Sweeting, 1973, p 241.)

Karst Domes

Karst domes are low structural features that superficially resemble the domal expression of some breccia pipes. In the past, these two features have been confused with each other (Vine, 1960). However, the internal structure of karst domes and breccia pipes is entirely different (Figure 26). Karst domes are true structural domes characterized by strata that surround and dip away from a central core of relatively older rocks, whereas the central core of breccia pipes is brecciated fill composed of rocks younger than the surrounding strata.

BRECCIATED CORE ROCKS YOUNGER THAN WALL ROCK

WALL ROCK

BRECCIA PIPE

CORE ROCKS OLDER THAN WALL ROCK

WALL ROCK
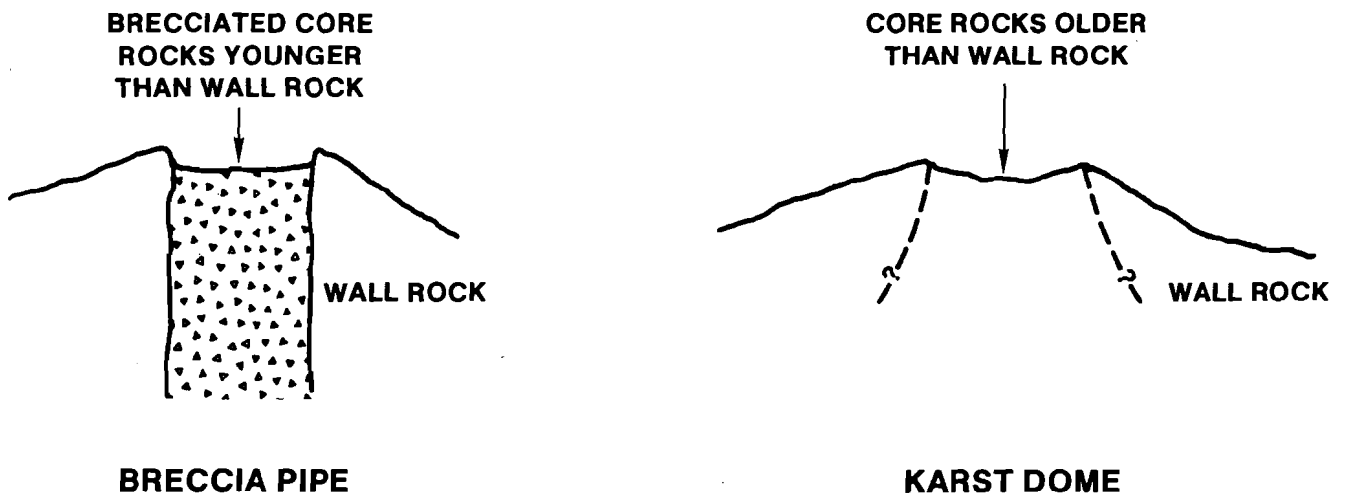
KARST DOME

Figure 26. Diagram showing difference between breccia pipe and karst dome

Karst domes are best developed in an area along the west side of the Pecos River near Malaga Bend. Although much of that area is underlain by chaotic breccia of the Rustler Formation, some domes stand out as conspicuous symmetrical features (Figure 27); at other places in the area the domal structure is disrupted and indistinct. The draping of relatively younger strata across older rocks on those features can be documented only by tracing and mapping individual rock units on the ground.

Figure 27. Aerial view of a karst dome near Malaga Bend (Sec. 19, T.24S., R.29E.), Eddy County, NM. (The Culebra Dolomite Member and associated beds of the Rustler Formation form the rim of the dome. Residuum of the Salado Formation is exposed in the core.)

One example of a symmetrical karst dome is ~0.8 mi (1.4 km) west of the Pecos River near Malaga Bend. (The southwest corner of Sec. 19, T.24S., R.29E. is near the center of this dome.) This dome is ~650 ft (200 m) in diameter and nearly circular. It rises ~35 ft (10.8 m) above the surrounding terrain. Pink to dark-red gypsum and associated insoluble residues of the Salado Formation are exposed to the core. Chaotic breccia of the Rustler Formation rests on the Salado core, and the Culebra Dolomite Member of the Rustler Formation is draped around the rim of the dome. Variations of these relationships are displayed on several dozen features westward from Malaga Bend for a distance of ~12 mi (19 km).

Karst domes have been mapped in detail (Reddy, 1961; Bachman, 1980), but individual domes have not been penetrated by boreholes. Consequently, their internal structure and roots are unknown and their origin can only be speculated.

Previous workers in the region have suggested that the karst domes are piercement features resulting from upward migration of salt bodies through overlying sediments (Vine, 1960; Reddy, 1961; Kelley, 1971). Kelley (1971, p 54) stated that "they are surficial structures related to the salt in the Salado at rather shallow depth, and sink hole collapse in near surface evaporitic beds." He stated further that he mapped these features "to include sink holes as well as piercements and there appears to be every gradation in kind and stage of development."

The distinctive structure of karst domes with a core of rocks older than the annular outcrops surrounding them separates these domes from collapse sinks. Although it is apparent that the domes are related to the Salado Formation, it is improbable that they are piercement bodies derived from the underlying Salado Formation. Salt plugs and domes occur as piercement bodies along the Gulf Coast of the United States, where they have been forced to the surface from mother salt beds at profound depths (5 to 10 km beneath the surface). They are the result of the pressures of loading from overlying sediments. Those pressures cause the salt to flow and become detached mobile bodies that, being of less specific

gravity than the enclosing rocks, force their way to the surface. The stratigraphic relationships in the vicinity of Malaga Bend are not conducive to the development of salt flowage. The Salado Formation is at or near the surface, and that region has never been buried to depths comparable to the salt beds on the Gulf Coast.

Examination of records of wells drilled for oil and gas in the vicinity of Malaga Bend shows that the base of the Salado Formation in that area ranges from 1050 to 1575 ft (323 to 485 m) below the surface. In some wells a few distinctive Salado marker beds are recognizable, but most marker beds are disrupted and masked by dissolution and collapse. Except for lag deposits, much of the Rustler has been eroded from the Malaga Bend area; where present in drill holes it is a mass of indistinctive collapse breccia. Wells drilled no more than 10 mi (16 km) east of Malaga Bend have penetrated complete stratigraphic sections of both the Rustler and Salado Formations (Figure 28). There the regional dip has plunged the base of the Salado nearly 4000 ft (1230 m) beneath the surface. Karst domes have not been found in that area.
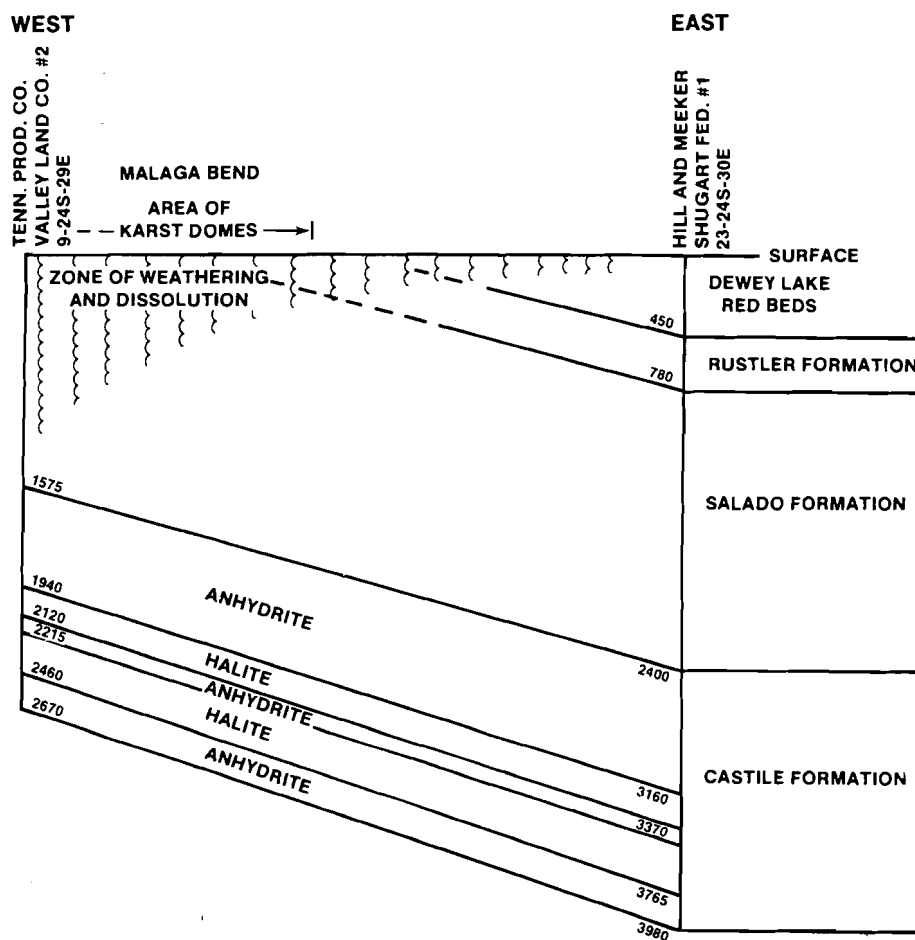
Figure 28. Diagram showing stratigraphic relationships in southeastern Eddy County, NM, where karst domes are located. (Regional dip is to the east. The eroded edges of the Salado and Rustler Formations are exposed at the surface in the vicinity of Malaga Bend, where they consist of discontinuous beds in breccia and residuum. Numbers beside drill holes indicate depth, in feet, from the surface.)

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It is here suggested that the karst domes are indeed "related to salt in the Salado Formation at rather shallow depth," as stated by Kelley; but that the domes are the surficial remnants of pervasive near-surface dissolution of salt instead of piercement. The ancestral Pecos River flowed across this area at least as early as Pleistocene time and was responsible for the dissolution of evaporites to depths >1000 ft (308 m) where it flowed across southeastern Eddy County (Figure 29). That dissolution caused general subsidence of the region around Malaga Bend and left remnants of insoluble residue projecting above the surrounding landscape. Karst domes are part of those remnants. It is noteworthy that dissolution has not affected the salt beds in the Castile Formation underlying the Salado along the Pecos River in the New Mexico part of the Delaware Basin (Bachman, 1984).
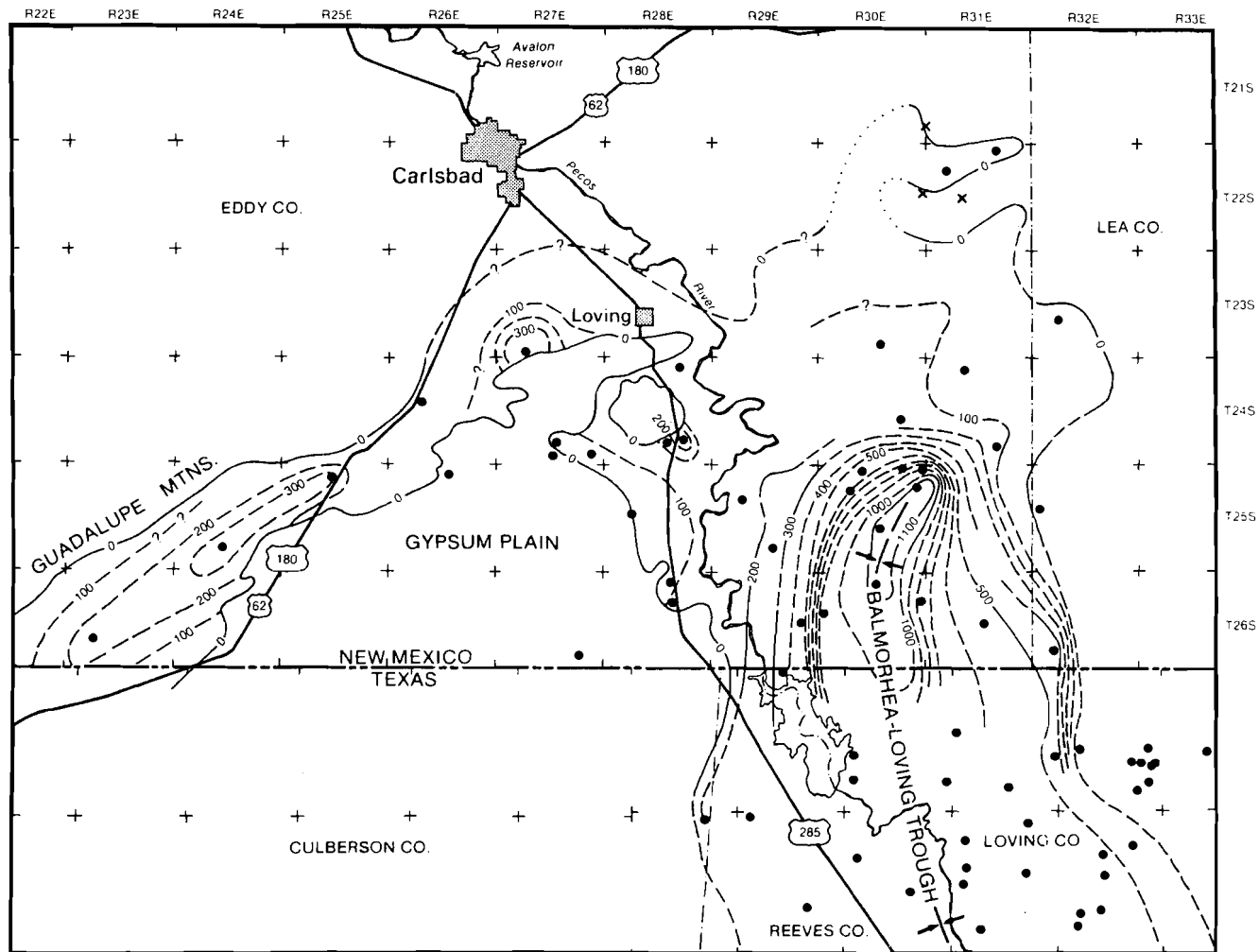
Figure 29. Isopach map of Cenozoic fill in Balmorhea-Loving trough and adjacent minor deposition basins. (Contour interval is 100 ft. There is abrupt thickening of fill where contours are crowded which may be explained by local collapse sinks later filled with sediments. Contours, except for zero contour, are less reliable in Texas owing to unavailability of samples for comparison with wire line logs. Solid zero lines are based on bedrock exposures.)

The Balmorhea-Loving Trough

Maley and Huffington (1953) described three local areas in western Texas and southeastern New Mexico where major accumulations of sedimentary fill coincide with places where extreme amounts of salt have been dissolved from underlying Permian beds. These areas of accumulation include a narrow belt on the eastern side of the Delaware Basin "essentially over and just west of the buried Capitan reef" (Maley and Huffington, 1953, p 541); a basin centered ~6 mi (9.6 km) north and slightly west of Pecos, Reeves County, TX; and a body of fill that trends northwest and straddles the New Mexico-Texas state line east of the Pecos River. Hiss (1975) combined the latter two bodies and named them the Balmorhea-Pecos-Loving trough. He named the body of fill on the eastern side of the Delaware Basin the Belding-San Simon trough. The northern portion of the western body is here called simply the Balmorhea-Loving trough after the usage of Lambert (1983, p 84).

Since the work of Maley and Huffington, many wells have been drilled in the vicinity of the Balmorhea-Loving trough in the search for oil and gas. Records and cuttings of some of these drill holes have been examined during the present study in an effort to further define the nature of the trough.

As noted by Maley and Huffington, the thickness of sedimentary fill in the Balmorhea-Loving trough cannot be perceived by casual examination of the surface. Surficial deposits, some engulfed by caliche, cover the region in the southeastern Eddy County east of the Pecos River. Low, rolling hills characterize the surface, which has been dissected by shallow erosion at only a few places. Even depressions such as Big Sinks do not give an impression of the considerable thickness of the underlying fill.

Well-sorted gravels, interpreted as stream deposits, are present on the surface at many localities in southeastern Eddy County and southward into Loving County, TX. These gravels consist mostly of chert and limestone pebbles, but they include occasional pebbles of igneous rock that may have been transported from the Capitan Mountains ~130 mi (210 km) to the northwest. (Igneous dikes are present in poor exposures in south-central Eddy County 35 mi (56 km) to the west, but they are deeply weathered and disintegrate before being carried away by surface runoff.)

Cuttings from boreholes were examined in conjunction with wire line logs (chiefly sonic or acoustic and gamma ray logs) from the same drill holes across southeastern Eddy County. Rock types are distinctive in the sequence. Where cuttings are available, it is usually possible to differentiate between Cenozoic sediments and bedrock that had been brecciated during the processes of dissolution and collapse, although this differentiation is difficult at times even with cuttings. At boreholes where only wire line logs are available for study, it is difficult and often impossible to separate sedimentary fill from chaotic dissolution breccia. Consequently, this discussion and the accompanying illustrations are efforts to accommodate the available data.

Sediments that are assigned definitely to "Cenozoic fill" include well-sorted medium-to-coarse, white to light-gray and reddish-brown sand, granular-to-pebbly sand, and some gray to reddish-brown silty clay. The conspicuous foreign sediment in this suite is the white to light-gray, well-sorted sand. Some of these sands may be of eolian origin. Except for the silty clay, none of these sediments resemble the clastics in the underlying Permian. The silty clays were included with the Cenozoic suite only where they appeared to be interbedded with other sediments more readily assigned to "Cenozoic fill." All these sediments are interpreted to have been deposited by streams on flood plains and in ponds. At times the sediments were probably reworked by the wind.

Maley and Huffington (1953, Plate 1) indicate that Cenozoic fill is more than 1400 ft thick in southeastern Eddy County, NM. This amount of fill was not substantiated during the present study, but some discrepancies could be explained by different interpretations of "fill" (i.e., sediments derived from allochthonous sources) and brecciated collapse debris (autochthonous sediments). The thickest deposit of fill observed during this study was ~1160 ft (357 m) (Figure 29). Records of nearby drill holes suggest fill of ~820 ft (254 m) overlying collapse debris ~480 ft (148 m) thick. The total of fill and collapsed rock is ~1300 ft (400 m).

In many drill holes along the Pecos River in Eddy County and Loving County, TX, where samples are not available for examination, it is apparent from the wire line logs that considerable quantities of salt have been dissolved from the Salado Formation. Even at places in the subsurface where the Rustler Formation is recognizable and relatively intact, the underlying Salado may be less than half its normal thickness.

Maley and Huffington attributed the localization of these major fill deposits to the dissolution of Permian evaporites, particularly salt, in the subsurface, accompanied by collapse and subsidence of the surface. The areas of subsidence became depositional troughs. They explained the dissolution of evaporites in the western part of the Delaware Basin by tilting of the basin to the east during late Tertiary time, which caused the western part of the evaporite section to be elevated and exposed at the surface. Downward-percolating waters from the surface dissolved the salt and some anhydrite.

On the eastern side of the Delaware Basin, Maley and Huffington observed that the Capitan reef is buried in the subsurface and that dissolution of salt has been localized in a belt parallel to the reef front. They believe that slight warping of beds overlying the reef concentrated surface runoff and percolating groundwater to dissolve salt in the subsurface.

The mechanism of evaporite dissolution and removal of evaporites from the subsidence basins has been considered by several workers. Maley and Huffington (1953, p 53) stated that "it is . . . clear that the fill does not represent the location of channels or valleys cut by one or more prehistoric rivers that are no longer present; according to this concept, the thickness of fill should represent the depth of the former stream valley." However, this is not a valid conclusion. Dissolution of soluble rocks in the subsurface occurs below groundwater level—at times far beneath stream surfaces (Moneymaker, 1941; Morgan, 1941). Such dissolution requires considerable time, and the resulting subsidence of the surface is a gradual process. Subsidence may thus carry the surface below the level of effective stream erosion, and the basin will be filled if a constant influx of sediments is carried into the subsiding basin. It is also predictable that the sediment loads of through-flowing streams will be dropped where the stream gradient decreases in such basins.

Anderson (1981) suggested that salt was removed from the Salado Formation by a process of brine density flow, which had been proposed earlier for the origin of breccia pipes (Anderson and Kirkland, 1980). This mechanism assumes that the Bell Canyon Formation underlying the Castile Formation is an aquifer from which fresh water migrates upward through fractures in the Castile to dissolve salt in the Salado Formation. However, permeability of the Bell Canyon Formation indicates that it is not an effective aquifer (Mercer, 1983, pp 26-32). In addition, beds of halite in the Castile Formation are intact and have not been subject to the dissolution that removed salt from the overlying Salado (Bachman, 1984).

Bachman (1984) presented evidence that during some portions of its history the ancestral Pecos River carried sediment loads far exceeding the capabilities of any recent drainage system in southeastern New Mexico. From this evidence it is assumed that rainfall, runoff, infiltration, and dissolution were more powerful agents and processes than at any time since. Salt beds in the Salado Formation were selectively dissolved by groundwater in a hydrologic system initiated by those climatic and drainage conditions.

The method of removal of the great quantities of brine that resulted from localized dissolution of salt has been considered to be a problem. Maley and Huffington (1953, p 544) stated that "the exact methods by which this is accomplished are not entirely clear, some of the liquids are apparently removed by subsurface circulation eventually bringing them near enough to the surface so that they drain into the Pecos River. . . ." Anderson (1981) would have the brine, by its greater density, sink into the Bell Canyon aquifer and be carried down-dip across the Delaware Basin. However, the Bell Canyon does not appear to be an adequate aquifer, and there is no evidence that fractures cut across the Castile Formation in the subsurface to connect the Bell Canyon with the overlying Salado Formation.

Chaturvedi and Rehfeldt (1984) said that Bachman (1984) did not address the question of the disposal of solution brine. However, Bachman (1984) assumed that the ancestral Pecos River system was capable both of dissolving and of removing salt from the Balmorhea-Loving trough. A drainage system capable of carrying sediment debris of the size and quantity demonstrated by deposits in ancestral Pecos channels (Figure 30) must have established far more forceful hydraulic heads than any represented by present groundwater systems.

Figure 30. Aerial view of a lenticular gravel in the Gatuña Formation on the south side of Pierce Canyon, Eddy County, NM (SW 1/4 Sec. 26, T.24S., R.29E.). (This channel deposit is ~78 ft (24 m) thick and 780 ft (240 m) wide. It contains clasts to 2.8 in. (7 cm) in diameter. Most of the clasts are Permian limestone, but ~8% are Tertiary porphyry similar to rock types in the Capitan Mountains ~120 mi (190 km) to the northwest. The Mescalero caliche forms the caprock of the mesa and engulfs the upper beds of the gravel. The gravel collapsed into a local sink and is not visible in the next canyon to the south.)

On the basis of elevations plotted by Hiss (1976a, 1976b) and data collected during the present study, a probable groundwater flow system can be inferred for the time of development of the Balmorhea-Loving trough. Channel fill deposited by the ancestral Pecos River is at an elevation of ~3050 ft (930 m) above sea level in Pierce Canyon, which is the approximate northern edge of the major axis of the Balmorhea-Loving trough. The base of the sedimentary fill in southeastern Eddy County to the south of Pierce Canyon is ~1250 ft (385 m) above sea level. This difference in elevation could create a potential hydraulic head of ~1500 ft (460 m). Even under present climatic conditions, fresh water storage in that area is much greater than in any adjacent area (Cooper, 1962).

On the northwest side of the Delaware Basin where the modern Pecos River flows across the position of the buried Capitan reef, the base of the sedimentary fill ranges from ~300 to 1000 ft (92 to 308 m) above sea level, while the top of the reef itself in that area is ~250 ft (77 m) below sea level. The hydraulic gradient of Permian strata and Cenozoic fill related to salt dissolution in Permian rocks is sufficient to warrant the assumption that dissolution brine could have been flushed out of the region by a through-flowing groundwater system. This paleo system has not been modeled.

The time of dissolution of the Balmorhea-Loving trough and associated features and the age of the sedimentary fill in the trough can be deduced only by indirect evidence. Maley and Huffington (1953, p 541) stated that "although most of the fill is probably Quaternary in age, it is likely that some of the older deposits are Tertiary, since processes responsible for the accumulation of Quaternary fill must also have been active during much of Tertiary time following the cessation of Cretaceous deposition." During the present study, the Mescalero caliche was observed to engulf some portions of the upper part of the ancestral Pecos River gravels. This is evidence only that sediments were deposited in the basin before the Mescalero caliche began to be deposited ~500,000 years ago. The age of the basal beds in the sedimentary fill is presumed to be much older—possibly as old as Late Tertiary.

Karst and the Pecos River

The Pecos River is an unusual drainage system; its history has been influenced almost as much by corrosion as by corrasion. The course of the Pecos River in much of New Mexico has been determined by karst, at least since mid-Pleistocene time and probably as early as late Tertiary time. The Pecos flows through an area of collapse sinks and solution-subsidence troughs around Santa Rosa, NM, ~40 mi (64 km) northwest of Fort Sumner (Sweeting, 1972). At places, waters from the Pecos feed small perennial lakes in collapse sinks in that area. The town of Santa Rosa itself is situated within a broad coalesced collapse sink ~6 mi (9.6 km) in diameter and as much as 400 ft (123 m) deep (Kelley, 1972).

The karst in the Santa Rosa area results from dissolution of San Andres Limestone and gypsum in the underlying Permian sequence. The modern collapse features began to form during Pleistocene or late Tertiary time (Kelley, 1972), but local intraformational disruptions in Triassic rocks in that area indicate that some karst processes were active during Triassic time (Bachman, 1976). Triassic karst activity in that area is one reason for suspecting that similar processes may have been active during Triassic time farther to the south around Carlsbad.

The most spectacular event in the history of the modern Pecos drainage system was the capture of the upper Portales Valley by the Pecos River. Baker (1915, pp 52-54) described the evidence that defines the event. This evidence was reexamined in the field during the present study. Early in Pleistocene time the Pecos River flowed through two separate valleys. A northern Pecos flowed southerly from the Sangre de Cristo Mountains in north-central New Mexico to the vicinity of Fort Sumner, where it entered the Portales Valley and flowed southeasterly. Near Fort Sumner, the Portales Valley is ~12 mi (19 km) wide and ~250 ft (77 m) deep. The valley is cut between the eroded edges of the High Plains Ogallala Formation and drains into the upper reaches of the modern Brazos River in western Texas (Figure 31).

At the same time, to the south the shorter ancestral Pecos drainage system flowed easterly from the back slope of the Sacramento and Capitan Mountains to the approximate position of the present Pecos Valley near Roswell and Carlsbad. It then flowed southerly near the position of its present flood plain. During this part of its history, tributaries flowed westerly from the High Plains into the main stem of the river. These ancient tributaries are now represented by deposits of gravel. Only a very few of these tributaries now exist as remnants of that former system.

During mid-Pleistocene time, the northern (Portales) and southern Pecos drainage systems became integrated. The southern Pecos cut northward from near Roswell to the vicinity of Fort Sumner. The river that formerly flowed through the Portales Valley was captured and began to flow along its present course between Roswell and Fort Sumner. The Portales Valley was left with an underfit drainage system and is presently covered in places with windblown sand and dotted with small playas.

Leonard and Frye (1975, p 14) stated that although subsidence, or collapse as a result of of salt solution at depth, is evident in the southern part of the region today, it "played a minor role in the drainage development prior to the Wisconsinan [Late Pleistocene]. Ogallala deposits, or Pleistocene deposits older than Wisconsinan as fillings of subsidence or collapse areas are not evident." This apparent absence of definite Cenozoic fill in subsidence areas older than Late Pleistocene presents problems in interpreting the time of subsidence. Still, there is evidence that the capture of the Portales Valley drainage by the Pecos River resulted from dissolution of evaporites in the subsurface accompanied by subsidence of the surface.

Along Five Mile Creek west of the Pecos River and ~25 mi (40 km) north of Roswell, R. L. Borton (Office of the New Mexico State Engineer) recently discovered remnants of broken and steeply dipping Triassic rocks that have collapsed into Permian strata of the Artesia Group (Figure 2). These rocks are ~15 mi (24 km) west of the nearest continuous exposures of Triassic strata and have collapsed vertically

at least 200 ft (62 m). If the easterly regional dip is considered, they may have collapsed vertically as much as 400 ft (124 m). These displaced Triassic rocks are here considered to be evidence for subsidence of the surface across the region that was formerly part of the southern Pecos drainage system. This subsidence is interpreted to have increased the gradient of the Pecos to the extent that capture of the Portales Valley could occur.
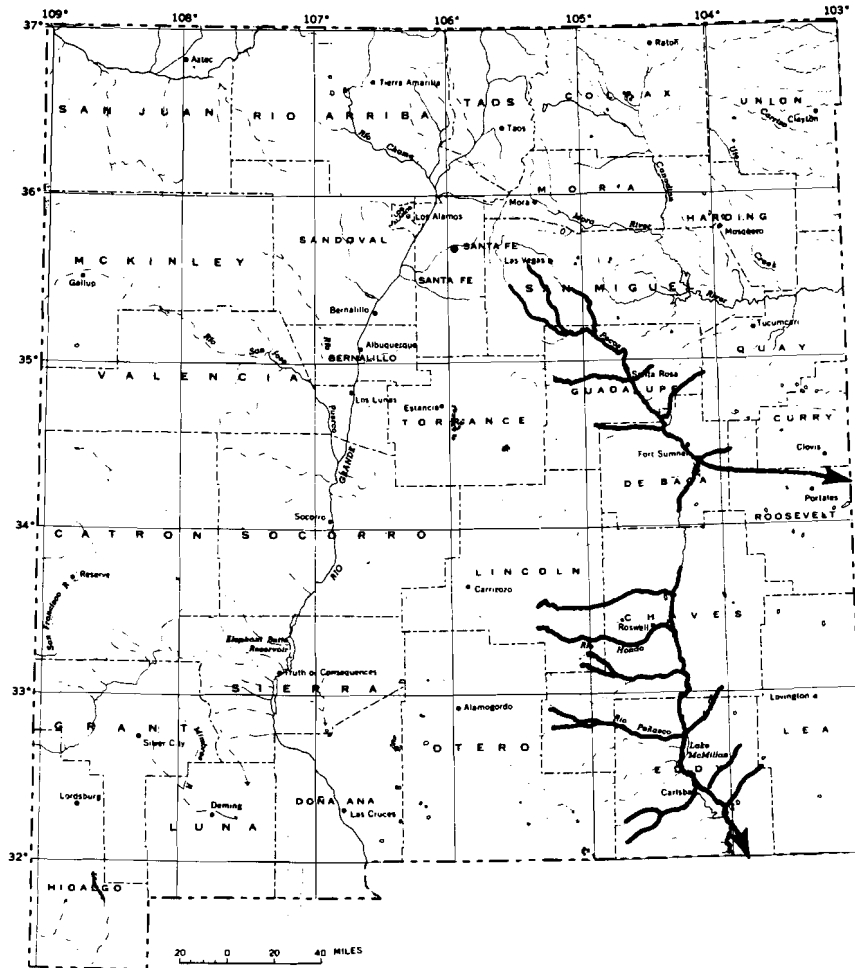
Figure 31. Map showing drainage patterns of Pecos River in eastern New Mexico during early to mid-Pleistocene time, when the Pecos was divided into separate northern and southern systems.

In the vicinity of Roswell, subsurface dissolution has removed 400 to 600 ft (123 to 184 m) of evaporites from the San Andres Limestone (Welder, 1983, p 8). Much of this dissolution probably occurred during Late Permian (Artesia) time (Borton, 1972, p 9; Welder, 1983, p 8), but the stratigraphic thinning that resulted from that dissolution appears to influence the position of the present channel of the Pecos River. Closed, gravel-filled depressions ~200 ft (62 m) deep are present at the base of the valley fill along the Pecos River near Roswell (Welder, 1983, Figure 5). These appear to be filled collapse sinks.

From the latitude of Carlsbad southward into Texas, the course of the Pecos River has undergone drastic changes since mid-Pleistocene time. During mid-Pleistocene time, and possibly earlier, the course of the ancestral Pecos River was as much as 12 mi east of its present channel in southeastern Eddy County, NM. During part of that early history the Pecos had tremendous carrying power. It transported cobbles and pebbles from as far away as the Capitan Mountains (Figure 30). Then, by as early as mid-Pleistocene time, dissolution of evaporites resulted in an extensive karst plain southeasterly from the latitude of Malaga. Where the Gatuña is exposed in that area it commonly fills ancient collapse sinks. The Balmorhea-Loving trough was created along the ancestral Pecos drainage by selective dissolution of subsurface halite beds in the Salado Formation and subsidence of the surface (Bachman, 1984, pp 17-20).

Today the Pecos River in southeastern New Mexico is generally a sluggish stream that flows partially underground. It is fixed in a channel that has been incised by both erosion and dissolution. Presumably it will remain in its present course unless the climatic regime undergoes a drastic change towards greater precipitation.

Groundwater in Evaporite Karst

Introduction

Water is a scarce commodity in southeastern New Mexico, and numerous studies have treated the hydrology of the Carlsbad and Roswell areas (Fiedler and Nye, 1933; Robinson and Lang, 1938; Hale, 1945; Motts, 1968; Cooper and Glanzman, 1971; Hiss, 1975). More recent studies address specific problems in these areas (Mercer, 1983; Welder, 1983). It is not the purpose of the present report to reexamine the details of hydrology in this region. Only the effect of evaporite karst on the groundwater system will be discussed.

Terminology used here follows the usage outlined by Mercer (1983, pp 24-26). Permeability is used as a qualitative term to refer to the ability of a rock to transmit fluid. The term "aquifer" is not used here in its formal sense. "Formal usage defines an aquifer as a geologic formation or group of formations or a part of a formation that is capable of yielding economic quantities of water to a pumped well or to springs. The use of the term aquifer in reference to most water-bearing zones . . . would be misleading. More appropriately these zones will be referred to as 'hydrologic units' or 'water-bearing zones' in this report." (Mercer, 1983, p 25.) Transmissivity is the rate at which water moves through a hydrologic unit of a given gradient.

The term "water table" refers to the "upper surface of a zone of saturation except where that surface is formed by an impermeable body. No water table exists where the upper surface of a zone of saturation is formed by an impermeable body." (Meinzer, 1923, p 22). The water table is a particular potentiometric surface; the potentiometric surface is the elevation to which water will rise in tightly cased wells that penetrate confined hydrologic units (Mercer, 1983, p 25; Lohman, 1972, p 8). Those distinctions between "water table" and "potentiometric surface" are basic to the consideration of the occurrence of groundwater in karst.

Discussion

Although evaporites are highly soluble, they are relatively impermeable. Most groundwater in these rocks is confined to secondary interstices such as joints, fractures, and solution cavities. Large amounts of water may be introduced to an evaporite sequence only through open fractures or bedding planes in the evaporites or permeable interbeds of dolomite, limestone, or clastic rocks (Lambert, 1983, pp 83-86). Where these zones of permeabiltiy serve as conduits for a constant supply of fresh water, they may become part of an open circulating system that dissolves the adjacent evaporites and carries away the solutes. Dissolution of the evaporite salts ceases when the water becomes saturated. If there is no outlet for the system, it will stagnate into a body of saturated brine.

Much groundwater in karst regions is stored in confined hydrologic units whose upper surfaces are impermeable bodies. In evaporites the hydrologic units may be so isolated locally that they do not contribute to dissolution. Similarly, evaporite beds may be isolated by impermeable bodies to the extent that they are not affected by nearby aquifers or hydrologic units. In southern Eddy County, beds of anhydrite and halite in the Castile Formation are preserved intact beneath Cenozoic clay, silt, sand, and gravel as much as 800 ft (246 m) thick. At least four saturated sand beds, some as much as 50 ft (15 m) thick, are present within the Cenozoic deposits (Cooper, 1962) of the Balmorhea-Loving trough; yet the underlying evaporites in the Castile Formation have not been dissolved. Thick anhydrite beds in the Castile itself appear to isolate the interbeds of halite and to protect them from dissolution (Bachman, 1984).

It is now generally assumed that major excavation of limestone caves occurs in the shallow phreatic zone just below the water table. The most continuous groundwater circulation is in that zone beneath the vadose and above the deep phreatic zone of static groundwater. Various levels within cavern systems have been localized by changing levels in the zone of saturation (Thrailkill, 1968). Higher cavities are presumed to indicate falling water-bearing zones.

Most of the limestone caves in the Guadalupe Mountains, including Carlsbad Caverns, are believed to have been excavated in a shallow phreatic zone during Tertiary time before uplift of the Guadalupe Mountain block (Bretz, 1949; Gale *in* Hayes, 1957). These caves appear to be related to an ancient drainage and hydrologic system (Horberg, 1949). The secondary formations in these caves, such as stalactites and stalagmites, are being deposited in the vadose zone since uplift of the mountains.

Application of the concepts of cave excavation in limestone leads to problems in the explanation of caves in gypsum. For example, most caves in gypsum appear to be related in some way to present surface runoff. Occasional pools of water resulting from surface runoff are found in some of these caves, but apparently the greatest excavation of gypsum caves has been in the vadose zone.

This relationship to surface drainage suggests that most of the excavation of these caves must be the result of surface runoff during flooding, or they are remnants of a former hydrologic cycle. Lee (1925) described the process of "erosion by solution and fill" that is responsible for enlarging some cave systems in gypsum, but this process may not be responsible for many caves such as those on the Gypsum Plain.

Enormous volumes of dissolved solids are presently being carried down the Pecos River. Morgan (1941) observed that the quantity of dissolved solids increases progressively downstream from the vicinity of Santa Rosa to Red Bluff Reservoir (Figure 2). He calculated that 77% of the dissolved solids entered the river above Artesia and was contributed largely from the San Andres Limestone. However, 23% of the dissolved solids enter the Pecos River between Malaga Bend and Red Bluff Reservoir.

Hale et al. (1954) believed that large quantities of salt were being discharged into the Pecos along its east side near Malaga Bend. That discharge is from a "brine aquifer" situated at the Salado-Rustler contact. Unusual quantities of dissolved solids have not been reported in Black River, which drains the northern part of the Gypsum Plain and empties into the Pecos from the West.* This suggests that, in spite of the many karst features on the Gypsum Plain, dissolution may not be as significant in that area at present as it has been in the past.

Among the evaporite karst features along the Pecos drainage system that can be dated, it is evident that many are relict from Pleistocene time. The breccia pipes, karst domes, buried collapse sinks in the vicinity of Pierce Canyon and Malaga, and other major features such as Nash Draw, were formed during Middle Pleistocene time when water surplus was greater and both erosion and dissolution were active processes on a large scale. The magnitude of that erosion and dissolution is illustrated by the headward cutting and collapse of the surface that rerouted the Pecos River southward from the vicinity of Fort Sumner into the Roswell Artesian Basin.

During Late Pleistocene time, climate was much different from the present in southeastern New Mexico. Temperatures were lower and effective precipitation higher than at present. Perennial lakes were common in many parts of New Mexico, including the Pecos Valley. Some of those lakes have been dated by radiocarbon methods to have been extant no more than 18,000 to 13,000 years before the present (Leopold, 1951; Antevs, 1954; Leonard and Frye, 1975). Lake deposits in Clayton Basin may have formed at that time. Summer rains were probably greater in the nearby Guadalupe Mountains than at present (Harris, 1970; Van Devender et al., 1979). Near Carlsbad a major climatic change occurred ~8000 years ago that resulted in a reduction in winter rainfall followed by drought (Van Devender, 1980). It is presumed that these climatic fluctuations resulted in changes in the groundwater regime.

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*In 1966, for which records are available, the specific conductance of waters from Black River ranged from 1940 to 2110 microsiemens/cm. During the same year waters from near Malaga Bend ranged from 6620 to 17,200 μS/cm. (US Geological Survey, Water Resources Data for New Mexico, Part 2, 1967, pp 136-138.)

Karst processes may reorganize drainage systems, or they may contribute to the development of new systems. Dissolution of evaporites and collapse of the surface contributed to the rerouting of the Pecos River from the vicinity of Fort Sumner to the Roswell artesian basin. Karst processes disrupted Middle Pleistocene Gatuña drainage to form Nash Draw. Sares (1984) has studied the drainage systems of Chosa Draw and Black River, which are tributaries of the Pecos River. He concluded that Chosa Draw—once a continuous drainage into Black River—has an intricate history of erosion and subsurface diversion. These systems are now characterized by perennial surface drainage that is in the process of being reorganized by subsurface piracy. Sares stated (1984, p 87) that lowering of groundwater levels at the end of Pleistocene time was responsible for the collapse of some dolines in the Chosa Draw drainage system.

As dissolution enlarges caves and subsurface channels, surface drainage is captured and flows underground. This process reduces surface flow and results in underfit streams—streams that appear to be too small to have eroded the valley in which they flow. Underfit streams are not obvious in arid and semiarid regions, where much surface drainage is characteristically ephemeral. However, even ephemeral surface drainage may be greatly reduced during heavy rainstorms where runoff pours into the underground channels. Some reaches of arroyos are abandoned completely by this process.

In addition, in southeastern New Mexico the zone of saturation has been further lowered since Late Pleistocene by changes in climate toward water deficiency. The result of the massive lowering of the zone of saturation is an underfit vadose groundwater system.

The Cycle of Erosion in Evaporite Karst

One approach to understanding the development of land forms is the theory of cycles of erosion, which considers that landscapes may evolve through stages of youth, maturity, and old age. The general theory was elaborated by its chief exponent, William Morris Davis, during the first half of this century. Cycles of erosion in karst were also discussed early in American literature by Beede (1911) and later adopted by Europeans (Sanders, 1921). The concept has since undergone many modifications (Sweeting, 1950), and today is much less popular than during the early part of this century.

Although there are objections to the concept of cycles of karst erosion, direct observations in the field indicate that evaporite karst features undergo stages of development in geologic time. These stages range from dissolution-widened joints through the development of collapse sinks to a final stage of chaotic dissolution breccia, which masks all previous stages of development. For purposes of this discussion these are divided into three stages:

1. *Early*. Dissolution along joint planes at the surface, with minor collapse sinks. Drainage into individual sinks, if present, is weak and may be represented by gullies no more than a few meters long (Figure 9).

2. *Middle*. Collapse sinks are well-developed. Surface drainage is in strong arroyos with blind valleys or solution-subsidence troughs. Surface runoff drains into caves of various dimensions (Figure 13).

3. *Late*. Dissolution has reduced entire evaporite beds to a breccia of insoluble residue. Individual karst features are obliterated (Figure 12).

The relative age of karst activity in southeastern New Mexico is based on regional stratigraphic relationships. The probability that evaporites were dissolved and removed from southeastern New Mexico during Triassic and Jurassic time has been suggested earlier in this report. The geologic setting, especially during Triassic time, was very favorable for dissolution of evaporites. Streams flowed across lowlands overlying Permian rocks, and the hydrologic system must have penetrated some evaporites along the extreme western and southwestern edge of the Delaware Basin. Many of the dissolution breccias of the Castile and Salado Formations in that area may have been formed during that remote time in the past. However, the extent of that dissolution cannot be proved with available field evidence.

Among the stratigraphic units most useful for estimating the time of karst activity are the Gatuña Formation and the Mescalero caliche. The minimum age of the Gatuña is established as Middle Pleistocene (>600,000 yr before the present), and the basal portions of the formation may be as old as Late Tertiary. The Mescalero caliche began to form ~500,000 years ago. The Gatuña-Mescalero sequence is readily identified in the field and is a basis for establishing the age of some karst activity that has affected it.

In Pierce Canyon and at places along Salt Draw ~5 mi (8 km) south of Malaga, beds in the Gatuña dip steeply and appear to be limbs and wall rock of buried collapse sinks. At these localities, beds in the Gatuña Formation were at least partially lithified before the time of collapse; but the Mescalero caliche is deposited across the collapsed beds without itself being affected. This indicates karst activity in that area near the close of Gatuña time, but before deposition of Mescalero soil was well under way. Pre-Mescalero karst represents middle to late stages of activity, which were followed by a long period of karst inactivity and soil deposition. The Mescalero caliche could not have been deposited as a soil without surface stability.

The sediment-filled collapse basins in the Balmorhea-Loving trough in southern Eddy County are overlain by Mescalero caliche and represent dissolution that occurred long before Mescalero deposition, probably during Late Tertiary time. Pre-Mescalero karst activity was intense in the vicinity of Crow Flat ~15 mi (24 km) east of Artesia. There dissolution removed most of the evaporites in the Rustler Formation, and Triassic rocks foundered into sinks in the Rustler.

Most of the visible and well-defined karst features on the Gypsum Plain, Burton Flat, Nash Draw, Chalk Bluffs, and along the east side of the Pecos River near Roswell are the result of dissolution after Mescalero deposition. Many are probably relict features that began to form during Late Pleistocene time. These are divisible into early and middle stages of development. Absolute ages of these features have not been determined; however, some collapsed sinks have formed during the past few years. One sink in Lake Arthur collapsed in June 1973 (Bachman, 1974). San Simon Sink has collapsed in a series of events. The last recorded activity occurred about 1930 (Nicholson and Klebsch, 1961).

These stages indicate that karst processes follow a sequence of development from the initiation of dissolution on a soluble rock to the complete dissolution of the rock unit. However, an absolute time span cannot be predicted for each stage. The continuation of karst processes depends on the continuing availability of circulating, unsaturated water through geologic time accompanied by well-developed drainage. Little is known about the subsurface pathways of groundwater in karst in southeastern New Mexico, and data on cycles of climate through geologic time are based on indirect evidence. For these reasons it is difficult to determine the relative stage of development of many karst features. Attempts to assign absolute rates of dissolution to particular bodies of rock or to rates of formation of topographic features in a karst landscape (Bachman, 1974; 1980) may be based on false premises and at best are no more than conjectures.

Conclusions

Karstification of evaporites is a continual process in southeastern New Mexico. Dissolution to some degree began during Permian time when salt was dissolved from the San Andres Formation and has continued intermittently ever since.

Many shallow karst features that are observable today in the landscape may have begun to form at least as early as the last pluvial, 10,000 to 25,000 years ago, when precipitation was greater and more effective than at present. Most deep-seated features such as breccia pipes, karst domes, and the Balmorhea-Loving dissolution trough were excavated even earlier. These features are at least as old as mid-Pleistocene and may be much older.

Although some dissolution may be active at present at the contact between the Rustler and Salado Formations (the so-called "brine aquifer"), the most obvious deep-seated dissolution accompanied by collapse of the surface is occurring along the east side of the Delaware Basin at San Simon Sink and Wink Sink. Both these collapse sinks appear to be modern analogs of breccia pipes.

Many minor karst features in gypsum in southeastern New Mexico remain to be catalogued and described in detail. In particular, shallow caves in gypsum have not been examined for evidence of excavation under phreatic conditions. Much evidence will have been destroyed by corrasion in those caves subject to flooding, but evidence probably survives at some places that will indicate a history of groundwater conditions. Candidate caves for exploration of this type are the caves in the Castile Formation on the Gypsum Plain, and caves in the Rustler Formation and Artesia Group from Burton Flat northward through the Roswell basin. The caves in the Rustler Formation in Nash Draw are subject to intense flooding and probably will not produce evidence of their origin.

All karst processes begin with the infiltration of surface water into soluble rocks. The original water in rocks is usually so charged with solutes that it cannot dissolve additional mineral matter. Even the continual process of dissolution must be accompanied by mixing of waters from the surface, or the solutions will become saturated and stagnant. The nature of the pathways of water in the subsurface in karst regions present the most problems to understanding karst hydrology and the nature of rock excavation by dissolution.

LeGrand (1976, p 880) noted the lack of knowledge of hydrologic systems between recharge and discharge points and emphasized the need for a "total karst system approach" to the study of karst hydrogeology. He acknowledged that tracer studies help define the major directions of flow, but these studies do not help to understand the presence of water-level mounds over relatively impermeable strata and water-level valleys in areas of greater permeability.

The abrupt raising and lowering of groundwater levels is a characteristic of karst regions. These fluctuating water levels have not been observed in the region east of Livingston Ridge to The Divide, which suggests that karst is not present in that area. Yet, each region must be examined individually to determine the pathways of groundwater flow. In spite of the intensive groundwater studies undertaken in southeastern New Mexico, questions of areas of recharge, position of impermeable barriers, and direction of flow between neighboring features such as Clayton Basin and Nash Draw are not understood. Answers to these questions could be obtained by monitoring on a regional basis as opposed to local studies.

The history of the deep dissolution basins such as the Balmorhea-Loving trough may never be known in detail, but much more could be learned if more subsurface information was available. Available information in these basins consists only of data obtained by chance as a by-product of exploration for petroleum products. That exploration is aimed towards deep stratigraphic horizons; the shallower bodies of sedimentary fill are ignored or are logged only in a desultory manner. Drilling in southeastern Eddy County on the axis of the Balmorhea-Loving trough with the purpose of examining the stratigraphic section for records of environment of deposition, stages of dissolution of salt, and episodes of collapse of the surface would add significant information to the geologic history of that region.

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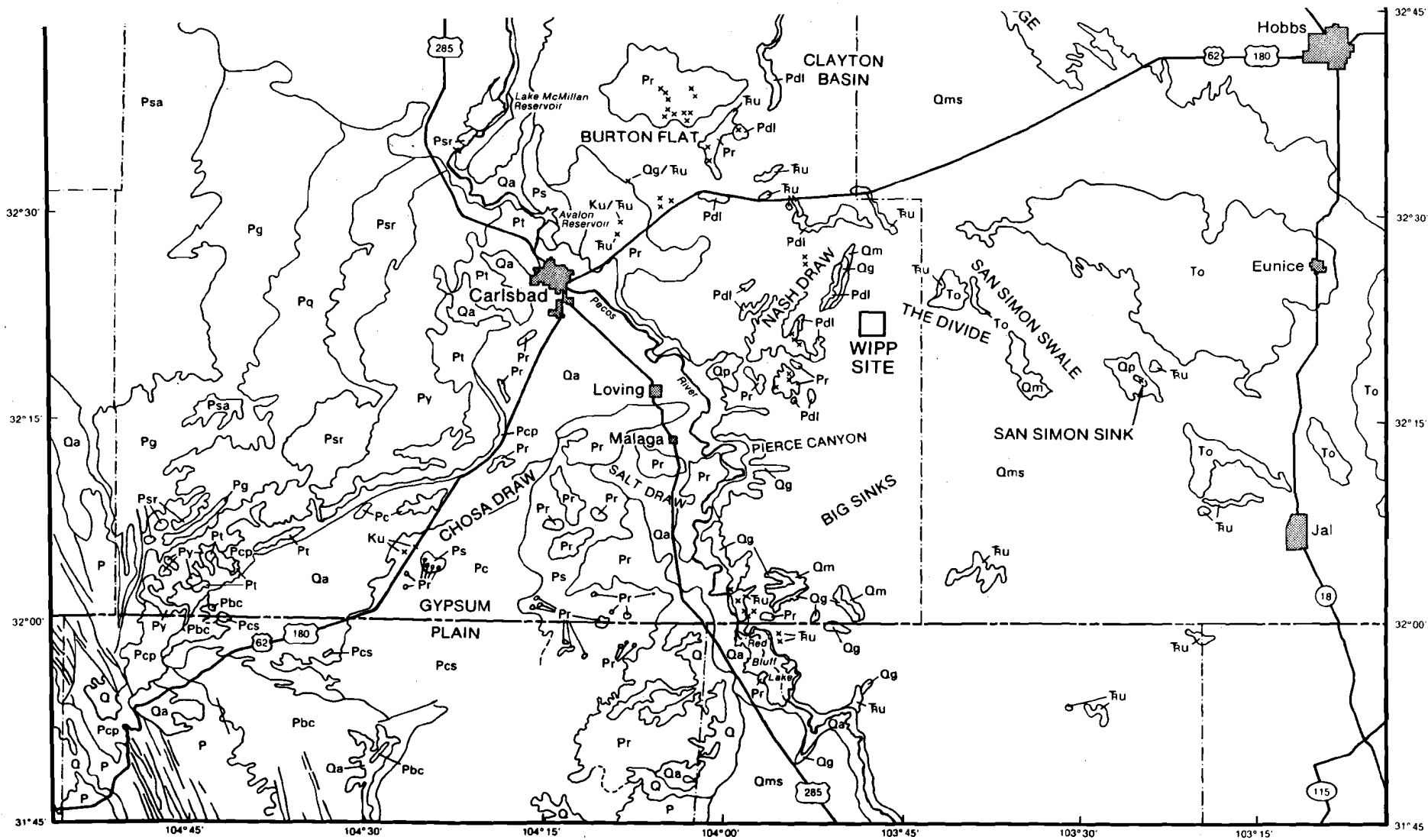
Figure 2. Geologic map of southeastern New Mexico

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