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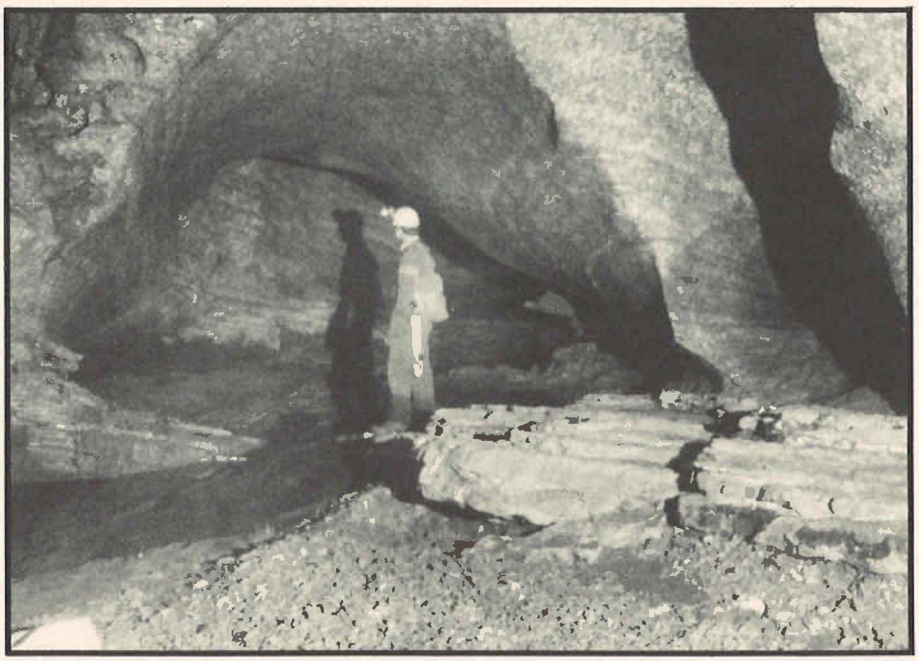
Hydrogeology and Karst of the Blaine Gypsum-Dolomite Aquifer, Southwestern Oklahoma

Kenneth S. Johnson



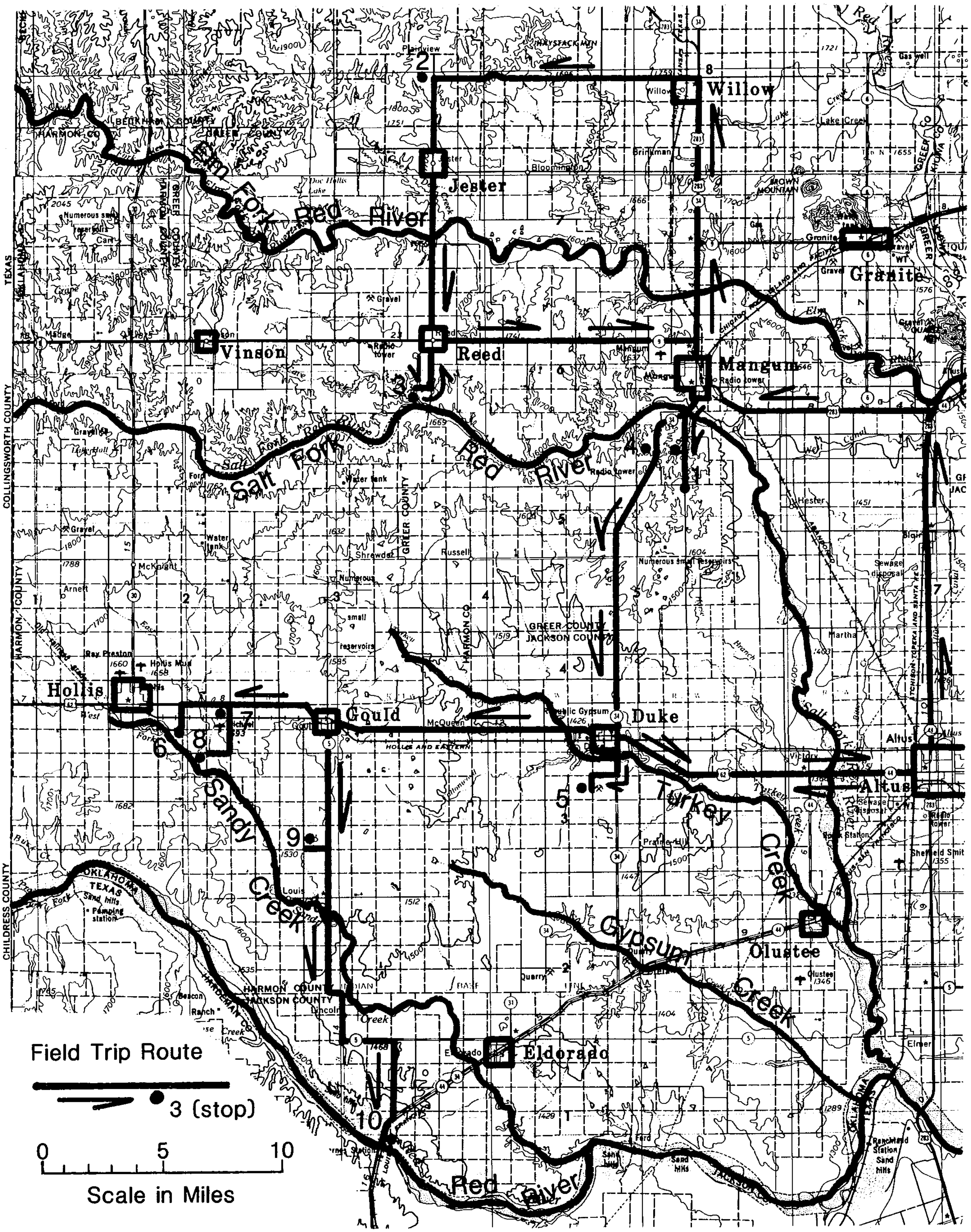
Field Trip #15
Guidebook

November 1–3, 1990

Geological Society of America
1990 Annual Meeting, Dallas, Texas

Sponsored by the Hydrogeology Division

GSA



Map showing route and stops for two-day field trip in southwestern Oklahoma. Stops 1–5 are on the first day, and stops 6–10 are on the second day.

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HYDROGEOLOGY AND KARST OF THE BLAINE GYPSUM-DOLOMITE AQUIFER, SOUTHWESTERN OKLAHOMA

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Guidebook for Field Trip No. 15, held November 1–3, 1990, following the annual national meeting of the Geological Society of America, October 29–November 1, 1990, Dallas, Texas. The trip was sponsored by the Hydrogeology Division of GSA.

Oklahoma Geological Survey
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The University of Oklahoma
Norman, Oklahoma

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SPECIAL PUBLICATION SERIES

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Front Cover

Cavern development in gypsum bed 1 of Van Vacter Member (Blaine Formation) in Jester Cave (Stop 2). John Bozeman is standing on Mangum Dolomite Bed. *Photo by Joe Looney.*

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PART 1

REGIONAL HYDROGEOLOGY

INTRODUCTION

Gypsum and dolomite beds of the Permian Blaine Formation make up a major karst aquifer that is being naturally and artificially recharged to provide irrigation water in 1,000 mi^2 making up the Hollis basin of southwestern Oklahoma. The Blaine aquifer is unique in being the only significant fresh-water aquifer developed in evaporite rocks in the United States. The Blaine typically is 180–220 ft thick, and it consists of a sequence of laterally persistent gypsum, dolomite, and shale interbeds. Gypsum and dolomite beds are partly dissolved by circulating ground waters, thus creating the karstic system comprising the aquifer. Karst features include major caves, sinkholes, disappearing streams, springs, and underground water courses. Irrigation wells in the district typically are 50–300 ft deep. They commonly yield 300–2,000 gpm of water containing about 1,500–5,000 mg/L dissolved solids; principal chemical constituents of the water are calcium, sulfate, and carbonate, and these have little or no adverse effect on crops being grown. In addition to the natural recharge that occurs through karstic outcrops in the district, landowners practice artificial recharge by diverting excess runoff and surface drainage to natural sinkholes or to recharge wells drilled 50–150 ft deep into cavernous gypsum-dolomite units.

The purpose of the two-day field trip is to examine outcrops, karst features, ground-water production, recharge, and the hydrologic regime of the Blaine aquifer. Day 1 consists of five stops focusing on evaporite geology, stratigraphy, caves, and karst development; day 2 consists of five stops dealing mainly with karst development, hydrogeology, water use, and aquifer recharge.

Previous studies of the geology, hydrology, and karst features of the Blaine Formation in the Hollis basin were conducted by Schoff (1948), Steele and Barclay (1965), Johnson (1967,1986,1990), Havens (1977), Bozeman and others (1987), Runkle and Johnson (1988), and Runkle and others (in review).

In preparing this guidebook, I received considerable assistance from my co-leaders, and special appreciation is expressed to them for their contributions. John R. Bozeman and Sue Bozeman provided information on cave studies and karst development at Jester Cave and Horseshoe Valley Cave. Donna L. Runkle contributed much of the hydrologic information incorporated into the guidebook, and Paul Horton provided data on ground-water production and recharge. Appreciation also is expressed to T. Wayne Furr and Charlotte Lloyd for drafting some of the illustrations, and to Christie L. Cooper and Jo Lynn Pierce for editing the manuscript.

STRUCTURAL SETTING

The Hollis basin is a large structural basin located in southwestern Oklahoma and adjacent parts of Texas. It is bounded on the north and northeast by the Wichita uplift, and on the south by the Red River uplift (Fig. 1). The Hollis basin developed in Early Pennsylvanian time with uplift and emergence of the Wichita and Red River blocks, and it contains a total of 3,000–12,000 ft of Late Cambrian through Permian strata that rest upon a basement complex of Cambrian and Precambrian igneous and metasedimentary rocks. The nearby Wichita Mountains consist of Middle Cambrian (about 525 mya) rhyolites, granites, and gabbros (Ham and others, 1964), and there is no evidence of post-Middle Cambrian igneous or hydrothermal activity in any part of the Hollis basin. Sediments in the Hollis basin have been deformed only by block faulting and folding, mainly during the Pennsylvanian Period, and the younger Permian strata have been gently deformed across these same structures (Fig. 2).

Outcropping Permian rocks in the Hollis basin are essentially flat lying. In most areas they dip at 5–45 ft/mi (0.05–0.5°), although in the northern part of the basin they dip about 90 ft/mi (1°). Along and adjacent to some of the local faults and flexures the strata dip 2–5°. All faults in the Hollis basin are believed to be normal, and vertical displacement across the faults ranges from 10 to 90 ft.

PERMIAN PALEOGEOGRAPHY

During the Permian Period, southwestern Oklahoma was on the eastern side of a broad, shallow, epicontinental sea that covered much of southwestern United States (Fig. 3). Because of slow, but continual, sinking of the earth's crust beneath this inland sea, a thick sequence of red beds and saline-sea evaporites (gypsum and salt) was deposited north of the major reefs and other marine carbonates of the Permian basin in west Texas. Normal marine water entered the Permian basin from the open ocean to the southwest, and, after passing over the carbonate reefs, it entered the shallow sea (or shelf areas) where evaporation took place.

Permian red-bed shales, siltstones, and sandstones deposited in southwestern Oklahoma were derived by the erosion of land areas in eastern Oklahoma, northeastern Texas, and other areas farther to the east. Streams and rivers draining these land areas carried mud and other fine sediments into the shallow sea where they were deposited in layers alternating with dolomite, gypsum, and salt. Thus, fresh and brackish water from the east mixed with the marine and saline

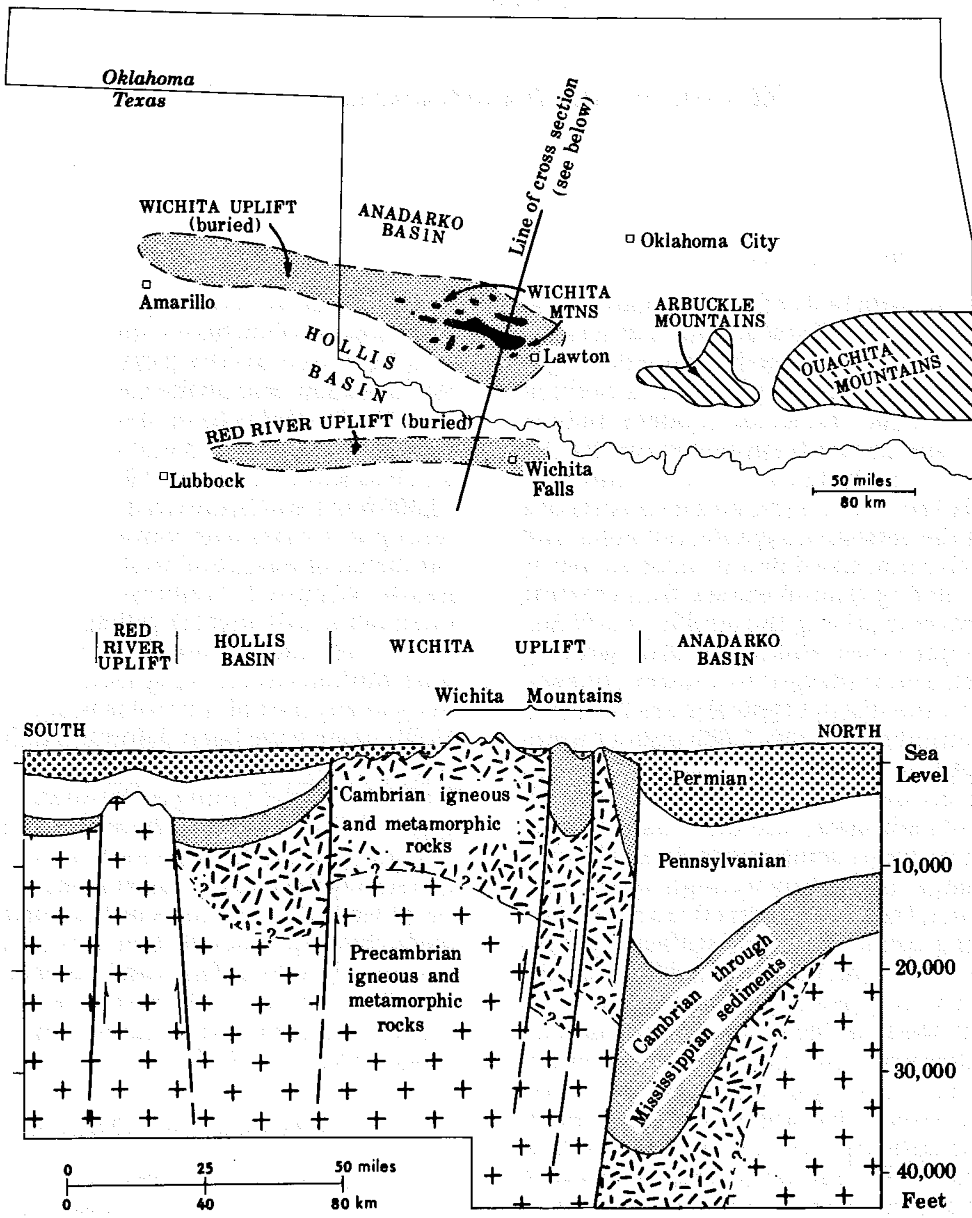


Figure 1. Map and cross section showing location of Hollis basin and other major provinces in the southwestern Oklahoma area (from Johnson, 1974).

waters from the southwest: shales, siltstones, and sandstones were deposited from the former, whereas evaporites were deposited from the latter.

Evaporite deposition in the Blaine Formation resulted from the evaporation of seawater. The concentration of dissolved solids in the seawater was raised by evaporation to the point where a series of "evaporite" rocks was precipitated on the seafloor, or in the muds just below. A typical, complete cycle of evaporite precipitation begins with the formation of a thin layer of dolomite (0.5–5.0 ft thick), followed by a massive layer of gypsum or anhydrite (5–30 ft thick), and finally, in areas more than 20 mi north and west of the Hollis basin, a unit of salt (halite, NaCl) (5–50 ft thick) at the top.

For a number of reasons, the complete evaporite sequence is not found everywhere within the region: evaporite precipitation may have been interrupted locally by an influx of less-concentrated water; certain chemicals may have been depleted from the saline waters before precipitation of a particular salt could start; or the more soluble units (salt, and sometimes gypsum) may have been deposited in some places, but were dissolved later.

Gypsum ($CaSO_4 \cdot 2H_2O$) is the common mineral making up the sulfate beds of the Blaine Formation in southwestern Oklahoma outcrops. Where the Blaine Formation is deeply buried, however, it normally contains some beds of anhydrite ($CaSO_4$) instead of gypsum.

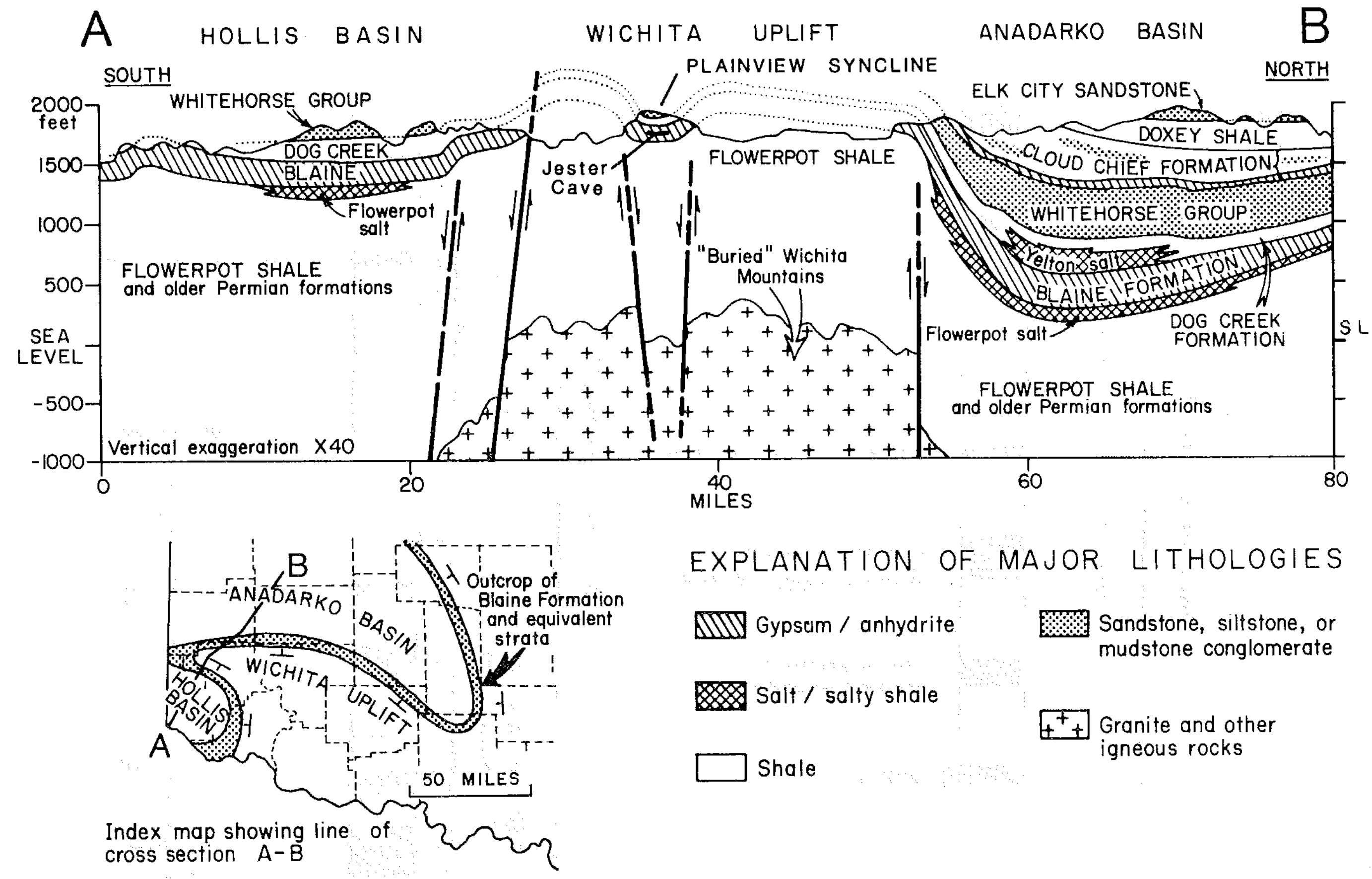


Figure 2. North–south cross section showing stratigraphy and structure of Permian rocks in southwestern Oklahoma (modified from Johnson and Denison, 1973). Principal emphasis for this report is in the Hollis basin and the Plainview syncline.

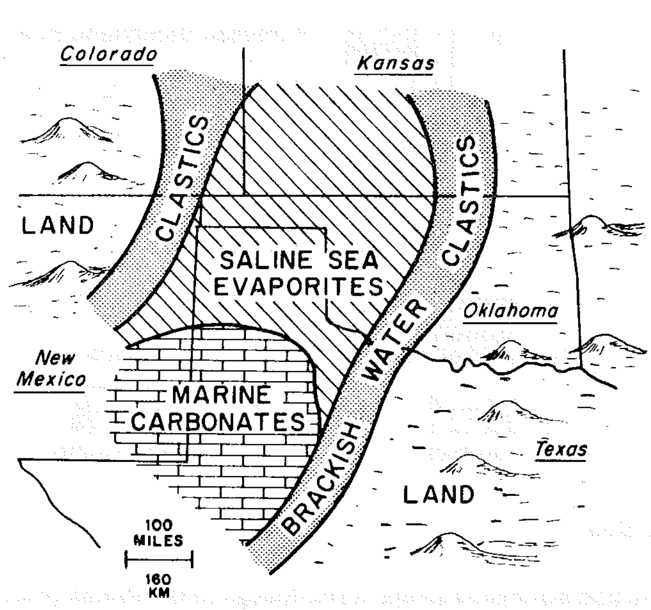


Figure 3. Paleogeography and principal facies in the Permian basin of southwestern United States during deposition of gypsum and other evaporites in the Blaine Formation (from Johnson and Denison, 1973).

Anhydrite is converted to gypsum by hydration (combining with water); this is a natural process in areas where ground water circulates through shallow sulfate deposits. Anhydrite has been observed in some Blaine outcrops of southwestern Oklahoma and in shallow cores drilled to depths of 25–100 ft. Anhydrite is more common at depths of 100–200 ft, and probably is the predominant sulfate mineral in the Blaine Formation and associated units where they are more than 200 ft below the land surface. In most of this report, we will refer to the sulfate rock as gypsum, whereas at depth much of it is, in fact, anhydrite.

STRATIGRAPHY

Outcropping sedimentary rocks of the Hollis basin area are Permian in age and consist mainly of flat-lying to gently dipping reddish-brown shales and sandstones interbedded with evaporites and thin dolomites. The stratigraphic sequence of Permian units in the study area is (in ascending order) the Flowerpot Shale, Blaine Formation, and Dog Creek Shale (Fig. 4): these units have been described by Johnson (1967, 1990). Of special interest for this report is the evaporite-bearing Blaine Formation of Guadalupian

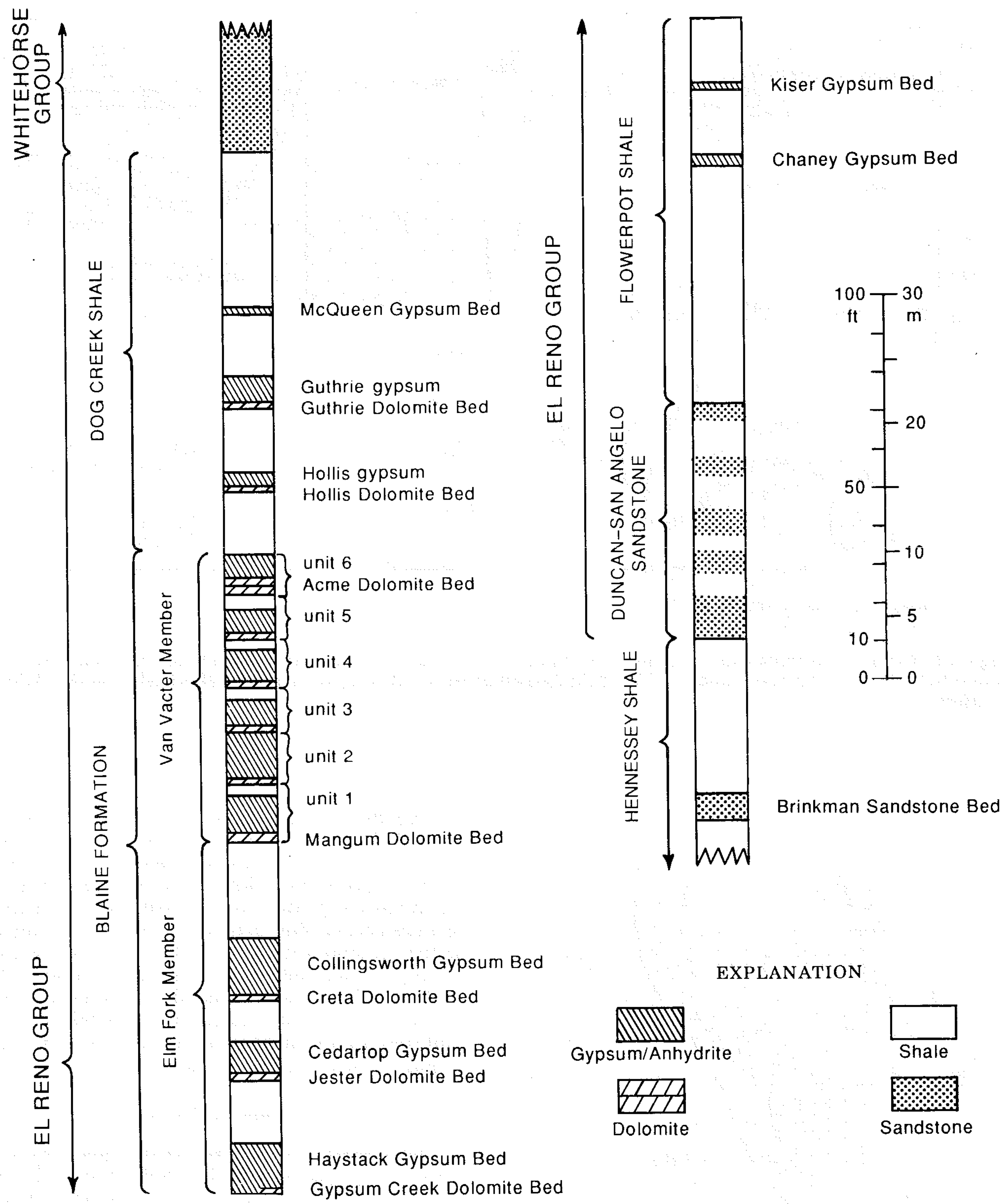


Figure 4. Standard outcrop section of Permian Blaine Formation and associated strata in southwestern Oklahoma (from Johnson, 1990). This figure is reproduced inside of back cover for ready reference.

(Hills and Kottlowski, 1983), or possibly late Leonardian, age. Most of the Blaine evaporite and dolomite units are excellent marker beds, traceable throughout the Hollis basin and over large parts of western Oklahoma and adjacent states. Locally the Permian strata are overlain by Quaternary alluvial and terrace deposits.

Flowerpot Shale

The Flowerpot Shale is a thick red-bed shale that underlies the Blaine Formation in all parts of the Hollis basin. The formation consists mainly of red-brown shale with several interbeds of gray shale and gypsum that are each 0.3–3.0 ft thick.

Total thickness of the Flowerpot is about 150–200 ft in the southeast and about 300 ft throughout the rest of the irrigation district (Johnson, 1967). Interbedded with red-brown shale in the northwestern part of the Hollis basin are layers of rock salt (halite), referred to as the Flowerpot salt. The Flowerpot salt is generally 50–200 ft thick, and its top is about 20–100 ft below the base of the Blaine Formation. The Flowerpot salt originally extended farther to the east and south of its present limits, but it has been largely dissolved by circulating ground waters, leaving patches of undissolved salt and/or high-salinity brine in the upper part of the Flowerpot Shale at various places in the Hollis basin.

Blaine Formation

The Blaine Formation consists of interbedded gypsum, shale, and dolomite in outcrops of southwestern Oklahoma. The formation commonly ranges from 180 to 220 ft thick and averages 200 ft thick (Fig. 4). Individual beds of gypsum and dolomite are laterally persistent throughout the region and can be traced into northwestern Oklahoma, Kansas, and north-central Texas. The Blaine Formation is a series of cyclic rock units, wherein each cycle consists of (in ascending order) dolomite, gypsum, red-brown shale, and green-gray shale. In southwestern Oklahoma, the Blaine typically has nine major cycles: light-gray dolomite beds at the base of each cycle commonly are 0.5–5.0 ft thick; white to light-gray gypsum beds are 5–30 ft thick; red-brown shales are 1–50 ft thick; and green-gray shales at the top of each cycle typically are 0.5–2.0 ft thick. Some of the dolomite beds locally consist largely of magnesite. Also, where deeply buried, the Blaine Formation normally contains beds of anhydrite instead of gypsum; and in some places anhydrite lenses exist within gypsum at shallow depths and in outcrops.

The Blaine Formation is subdivided into two members, the Elm Fork Member below and the Van Vacter Member above (Fig. 4) (Johnson, 1990). Although both members consist of interbedded gypsum, dolomite, and shale, there are distinct differences between them. The Elm Fork Member consists of three thick gypsum beds, each overlain by a thick shale (Figs. 4 and 5); these gypsum and underlying dolomite beds are readily identifiable in outcrops, where they form a conspicuous series of erosional benches. The Van Vacter Member, on the other hand, comprises six thick gypsum beds separated by thin shales (Figs. 4 and 6); the member typically crops out as low, rounded hills and is not readily subdivided, except in cores, boreholes, mines, and major bluff exposures. The two members are of nearly equal thickness (about 100 ft each) and are separated by the Mangum Dolomite Bed, an easily recognized, scarp-forming bed in most parts of the region.

Individual dolomite and gypsum beds in the Elm Fork Member are named (in ascending order): Gypsum Creek Dolomite Bed, Haystack Gypsum Bed, Jester Dolomite Bed, Cedartop Gypsum Bed, Creta Dolomite Bed, and Collingsworth Gypsum Bed (Figs. 4 and 5). Unnamed shales overlie each of the gypsum beds in the Elm Fork Member.

The six principal cyclic units of the Van Vacter Member are numbered consecutively, upward, as unit 1 through unit 6 (Figs. 4 and 6): they are given informal status inasmuch as the gypsum beds and most of the dolomite beds cannot be identified easily, except in boreholes, mines, or exceptional outcrops (Johnson, 1990). Each unit consists of, in ascending order, dolomite, gypsum, red-brown shale, and green-gray shale. Two of the Van Vacter dolomite beds are widespread and conspicuous, and they have been formally named: the Mangum Dolomite Bed, at the base of unit 1, and the Acme Dolomite Bed, at the base of unit 6.

Gypsum and dolomite beds of the Blaine are widespread and laterally persistent throughout the Hollis basin irrigation district. They crop out in much of the region (Fig. 7A) and the top of the formation is nowhere more than 400 ft below the land surface (Fig. 8). Owing to their shallow depth and importance as conduits for usable ground water, the Blaine strata have been penetrated in many water wells and other test holes. Thus there is a good understanding of the detailed stratigraphy and thickness variation of units within the Blaine throughout the Hollis basin (Johnson, 1967,1990).

The Blaine Formation, and individual gypsum beds within the formation, gradually thicken toward the west and southwest across the Hollis basin (Johnson, 1967). The thickness of the entire formation generally ranges from about 180–200 ft in the north and east to about 200–220 ft in the southwest. Individual gypsum beds of the Blaine typically range from 5 to 15 ft thick in the east to about 10–30 ft thick in the west; shale beds, on the other hand, typically thin westward across the basin. The thickest shale in the Blaine is near the middle of the formation, at the top of the Elm Fork Member; this shale is about 50 ft thick in the east and 30 ft thick in the west, and it is a good marker bed that is easily recognized after drilling through the massive gypsum beds in the overlying Van Vacter.

Dog Creek Shale and Younger Strata

Overlying the Blaine Formation in the western part of the area is the Dog Creek Shale (Figs. 7A and 8), which consists of 100–180 ft of red-brown shale with several gypsum-dolomite units in the lower 50 ft of the formation. The dolomite beds commonly are 0.2–2.0 ft

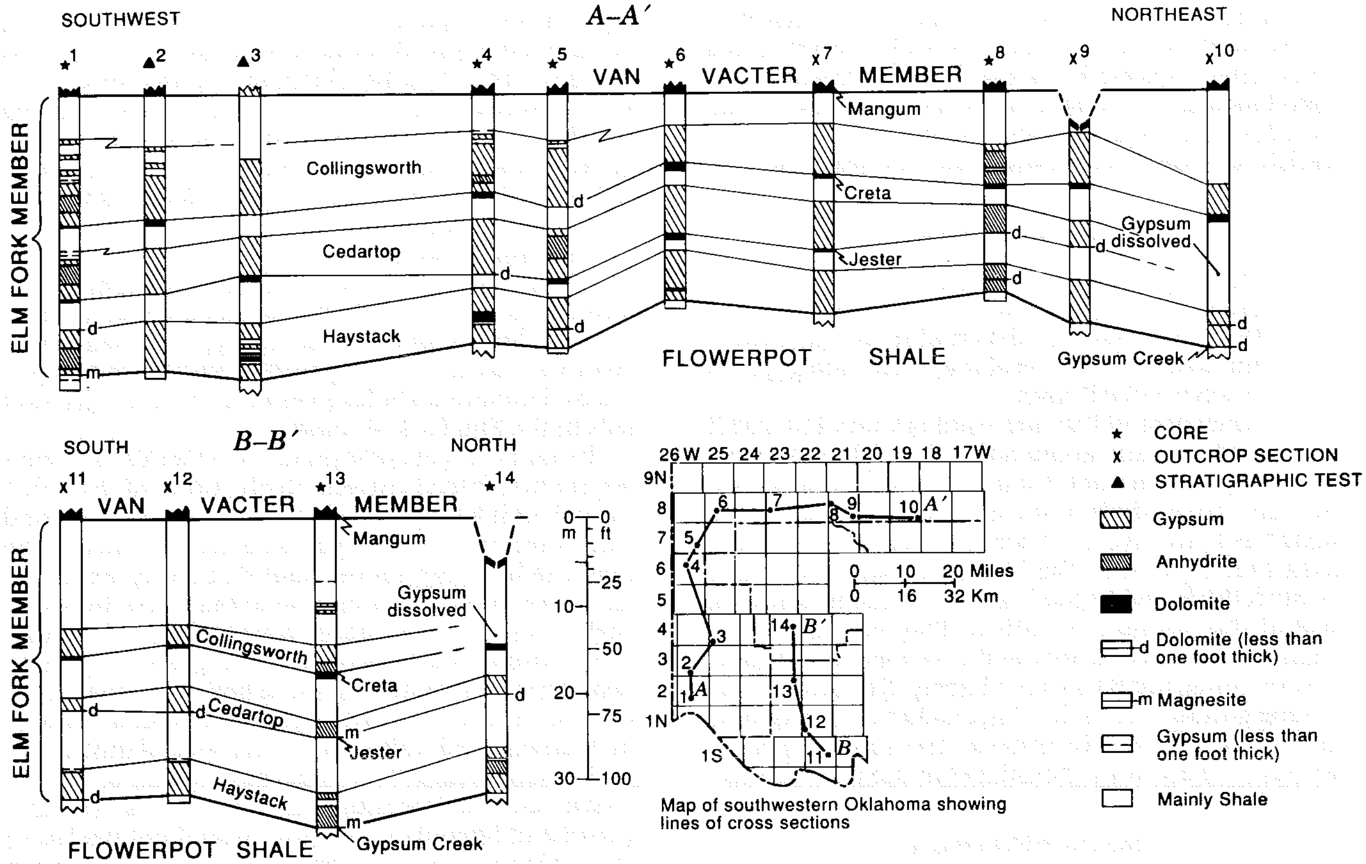


Figure 5. Stratigraphic cross sections of Elm Fork Member of Blaine Formation in southwestern Oklahoma (from Johnson, 1990).

thick and the gypsum beds commonly are 2–10 ft thick. The Dog Creek Shale is, in turn, overlain by 50–200 ft of sandstone of the Permian Whitehorse Group in the northwestern part of the Hollis basin. The Whitehorse Group consists of orange-brown, quartzose sandstone and siltstone, poorly cemented by gypsum or calcite, and it contains several thin shale and gypsum beds.

Overlying the Permian strata in parts of the Hollis basin are Quaternary alluvial and terrace deposits derived from rivers and streams flowing to the east and southeast. Terrace deposits typically are 15–80 ft thick and consist of tan, fine- to coarse-grained, quartzose sand with some gravel. Most alluvium is 10–50 ft thick and consists of mixtures of sand, salt, clay, and gravel.

KARST FEATURES

Gypsum and dolomite beds of the Blaine Formation contain extensive caves and make up a major karst aquifer that is providing irrigation water for much of the Hollis basin area (see cover photo). The following discussion of karst features is taken, with minor modifications, from Johnson (1986).

Gypsum and dolomite beds of the Blaine have been partly dissolved by circulating ground water, thus creating the cavernous and karstic system. Karst features include major caves, sinkholes, disappearing streams, springs, and underground water courses. These features are most common in areas where the Van Vacter Member is at or near the surface (Fig. 7A), because the Van Vacter consists predominantly of thick gypsum beds, and the low-permeability shale interbeds that it contains are quite thin. Areas where the karstic Van Vacter Member crops out, or is in shallow subsurface beneath the Dog Creek Shale, are the areas where the Blaine aquifer consistently yields large quantities of water to irrigation wells.

Development of porosity, and eventually of the open conduits through which water flows, occurs most commonly in the dolomite layers and the lower part of each of the overlying gypsum beds. Dolomite porosity occurs along bedding planes and cross-cutting fractures, and locally is due to early dissolution of (1) fossils, mainly pelecypod shells), (2) cement around oolites and pellets, and/or (3) small nodules of gypsum. In other places, the porosity is intercrystalline. In many places the development of porosity is so advanced that the dolomite beds have a honeycombed appearance, with only a skeletal framework of rock supporting a system of voids.

Early flow of water through the dolomite beds causes gypsum dissolution and cavern development to occur most commonly at, and just above, the gypsum-

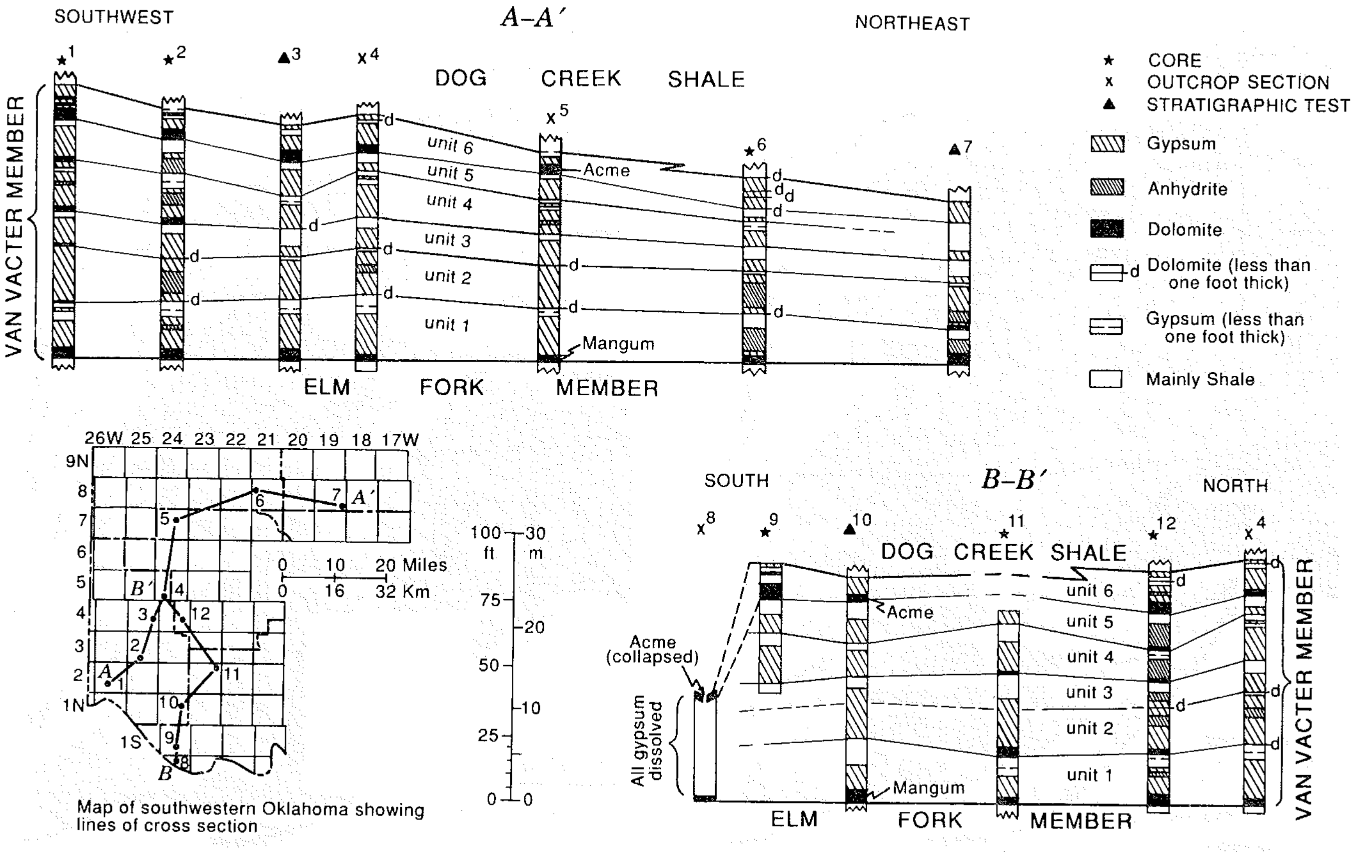


Figure 6. Stratigraphic cross sections of Van Vacter Member of Blaine Formation in southwestern Oklahoma (from Johnson, 1990).

dolomite contact (Fig. 9). Locally, the caverns have been developed along joints and bedding planes in the gypsum, but no preferential orientation of the dissolution features is known yet.

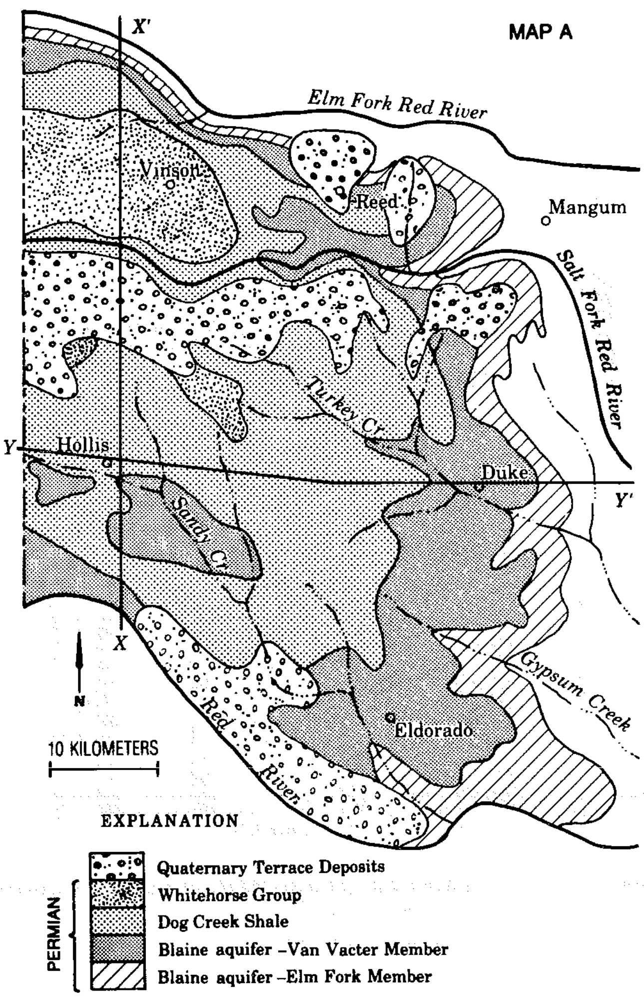
Caverns generally have a height and width that ranges from several inches to about 10 ft, although some caves reach 50 ft wide. Enlargement of individual cavities and caverns is due to dissolution of the soluble rock, as well as abrasion of the rock by gravel, sand, and silt carried by through-flowing waters. The sediment carried by ground water is deposited locally in the underground caverns, and it partly or totally fills some of the openings.

The D. C. Jester Cave system, the longest known cave in Oklahoma and the longest known gypsum cave in the western world (Bozeman and others, 1987), is an excellent and well-studied cave just north of the Hollis basin (Stop 2, this report). The main passage is 7,918 ft long, but along with the side passages the total surveyed length is 33,023 ft. Jester Cave and Horseshoe Valley Cave (Stop 3) follow angulate and sinuous courses, have passageways generally 4–50 ft wide and 3–20 ft high, and many parts of the caves are air-filled and dry (except during and after periods of moderate to heavy rainfall). In addition to the main caves, there are many tributary branches, sinkholes, and other entrances to the cave systems.

In many parts of the Hollis basin the underground caverns have become so wide that their roofs have collapsed to partly close the caverns. The collapse structures and fractured strata thus enable vertical movement of water in many parts of the aquifer. Such enhancement of vertical flow of water through fractures and collapse structures accelerates dissolution of all affected strata, thus increasing the amount of karst features in those areas. Dissolution and resultant collapse also create many problems in the local correlation of strata making up the aquifer: in some boreholes, one or more of the individual gypsum beds have been completely removed by dissolution, and overlying strata have collapsed to a lower stratigraphic position.

Karst features are generally sparse to nonexistent in those areas where the Blaine is buried at depths in excess of 200 ft below the surface. In these areas there has been little or no hydration of massive beds of anhydrite to gypsum, and the pathways for limited ground-water movement appear to be mainly in the dolomite beds.

Northwestern Oklahoma is the site of another major gypsum cave in the Blaine Formation (Myers and others, 1969). Alabaster Caverns, in Major County, has a maximum width of 60 ft and a maximum height of 50 ft. The public-access portion of the cave is 2,300 ft



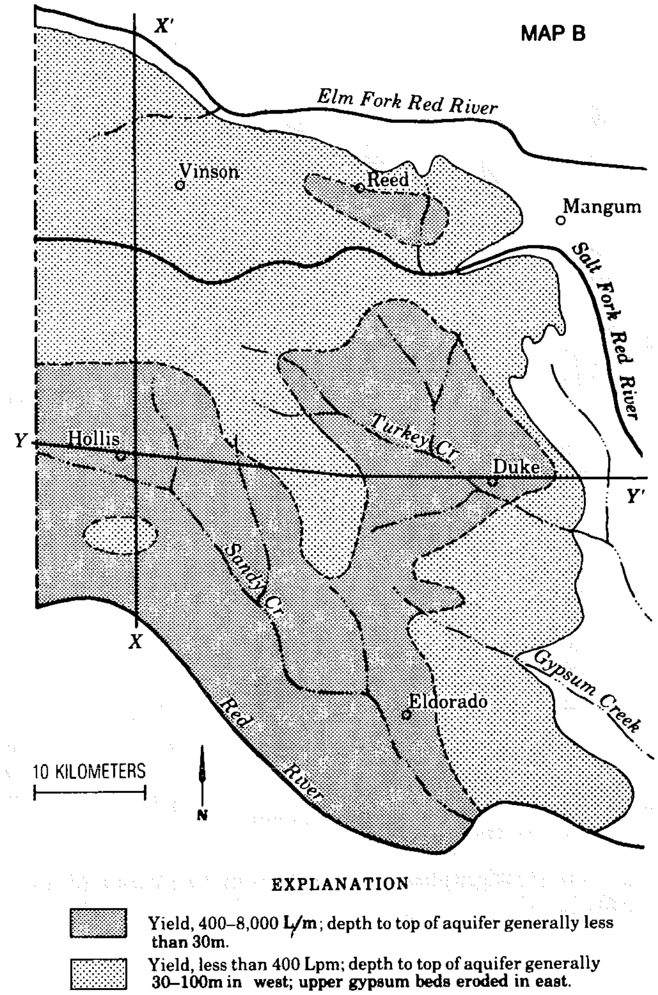


Figure 7. Maps of Hollis basin in southwestern Oklahoma showing generalized surface geology (Map A) and ground-water yields of Blaine aquifer (Map B) (from Johnson, 1986). Cross sections *X–X′* and *Y–Y′* are shown in Figure 8.

long, and it has been developed as a part of Alabaster Caverns State Park.

HYDROGEOLOGY

The Blaine aquifer has been yielding significant amounts of irrigation water for more than 40 years. High-yield wells produce water that is not potable, but it is suitable for producing crops. Irrigation wells completed in the Blaine aquifer typically are 50–300 ft deep and commonly yield 300–2,000 gpm. Depth to water in the Blaine commonly ranges from 5 to 80 ft below land surface.

The Elm Fork and Van Vacter Members of the Blaine Formation have quite different hydrogeologic characteristics. The Van Vacter commonly is 100 ft thick; it has six evaporite-clastic sequences, which consist of thick gypsum-dolomite units and thin (commonly 0.5–3.0 ft thick) shales. Hence, in many areas, water moving through the Van Vacter Member has dissolved and removed the gypsum and has eroded the thin shales, thus developing a good hydraulic connection among the various gypsum and dolomite beds. The Elm Fork Member also commonly is 100 ft thick; however, it has only three evaporite-clastic sequences, consisting of thick gypsum-dolomite units and equally thick shales. These shale beds, generally 10–30 ft thick, impede the vertical movement of water, and thus the hydraulic connection among the various gypsum and dolomite beds is not as well developed as in the Van Vacter Member.

Water from the Blaine aquifer in the Hollis basin is commonly of fair to poor quality, with the dissolved-solids content generally ranging from 1,500 to 5,000 mg/L (Table 1). There are large concentrations of calcium and sulfate in the water, which reflects the chemical constituents dissolved from the host gypsum beds of the aquifer. The water is highly mineralized and generally is not suitable for human consumption, but it is suitable for irrigation use in the region. There is no evidence of buildup of harmful amounts of calcium sulfate in lands irrigated for many years with water from the Blaine aquifer. Most of the irrigated lands have loamy soils with good drainage properties (Ford and others, 1984), and the crops include cotton, wheat, alfalfa, sorghum, and grasses.

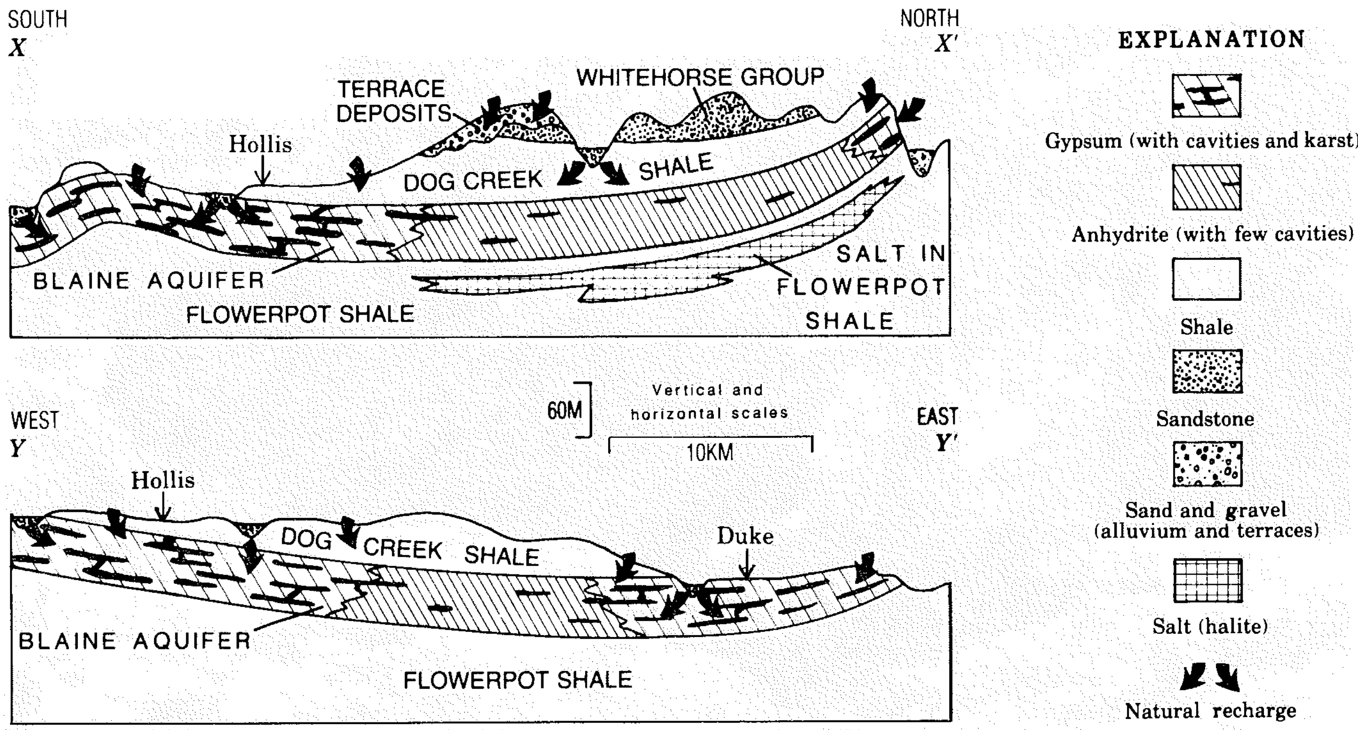


Figure 8. Generalized hydrogeologic cross sections showing Blaine aquifer in the Hollis basin (from Johnson, 1986). Locations of cross sections are shown in Figure 7.

The regional flow of water in the Blaine aquifer is from northwest to southeast. Areas that yield large quantities of water to wells (Fig. 7B) are along and parallel to Sandy Creek and Turkey Creek near the towns of Hollis and Duke. Sandy and Turkey Creeks recharge the Blaine aquifer during storm-runoff periods when the stream stage is higher than normal, and receive discharge from the Blaine aquifer during most of the rest of the year. This ground-water and surface-water interaction near principal streams enhances the conversion of anhydrite to gypsum and the dissolution of gypsum, thereby increasing the permeability of the Blaine aquifer. In many areas less than 3 mi from the streams, gypsum and dolomite beds have been partially or totally dissolved, creating conduits that allow large quantities of water to move rapidly though the aquifer.

Transmissivity of the Blaine aquifer in three single-well pump tests, conducted during the 1986–88 aquifer study, ranged from 5,450 to 43,353 ft^2/d, and averaged 18,504 ft^2/d (Table 2). The estimated storage coefficient ranges from about 0.0004 to 0.03, and averages about 0.016 (Steele and Barclay, 1965). The following average annual values for the Hollis basin area are from Pettyjohn and others (1983): precipitation, 25 in.; runoff, 0.5–1.0 in.; and evapotranspiration, 25 in.

In much of the western part of the study area, the Blaine aquifer is overlain by the Dog Creek Shale. Where more than 100 ft thick, the Dog Creek shale is considered a confining unit. However, where it is less than 100 ft thick, karst features in its three gypsum-dolomite units make the Dog Creek Shale a leaky confining unit.

Between the towns of Duke and Hollis and north of Hollis, where depths to the top of the Blaine aquifer are greater than 100 ft, anhydrite or a combination of anhydrite and gypsum are present (Fig. 8). The presence of anhydrite indicates that only small, slow-moving quantities of water have infiltrated the aquifer. Because of the relatively limited exposure to fresh, fast-moving water, anhydrite has not been converted to gypsum and the gypsum has not been extensively dissolved. In these areas the transmissivity of the Blaine aquifer is small (Fig. 7B).

Overlying the Dog Creek Shale, sandstones and siltstones of the Whitehorse Group are only a minor aquifer in the Hollis basin, and are not in hydraulic connection with the Blaine aquifer.

Quaternary terrace deposits, on the other hand, are an important source of good-quality ground water in several areas: (1) north of Red River; (2) south of Salt Fork Red River; and (3) north of Salt Fork Red River, near the town of Reed (Fig. 7A). Terrace deposits provide drinking water for the small towns in the study area and several rural water-supply associations (Table 1).

Alluvial aquifers of limited importance occur within the valleys of the Red River and the Salt Fork Red River. Water from the Red River alluvial aquifer has a dissolved-solids concentration of about 20,000 mg/L,

Figure 9. System of small caves developed in lower part of a gypsum bed in the Blaine Formation. Horizontal dolomite bed, about 1.5 ft thick, is at base of cavernous gypsum.

TABLE 1. — CHEMICAL ANALYSES OF GROUND WATER FROM BLAINE AQUIFER AND TERRACE DEPOSITS IN HOLLIS BASIN AREA OF SOUTHWESTERN OKLAHOMA

(from Steele and Barclay, 1965) Data (except pH) are in mg/L

	Blaine Aquifer[a]		Terrace Deposits[b]	
	Range	Mean	Range	Mean
SiO_2	13–18	15	9–19	13
Ca	240–761	603	53–87	72
Mg	75–219	151	12–21	17
Na+K	42–480	260	28–51	40
HCO_3	55–292	228	106–357	278
SO_4	949–2,250	1,810	19–188	60
Cl	38–565	280	8–27	16
TDS	1,840–4,690	3,110	315–458	390
pH	6.9–7.4	7.2	7.2–7.9	7.5

[a]Data from 33 water samples (5 samples from unused wells with chloride concentrations of 1,400–85,000 mg/L are not included).

[b]Data from 4 water samples.

TABLE 2. — TRANSMISSIVITY OF THE BLAINE AQUIFER FROM THREE SINGLE-WELL PUMP TESTS IN SOUTHERN HARMON COUNTY

(data from D. L. Runkle, U.S. Geological Survey)

Location	Pump Rate (gpm)	Drawdown (ft)	Transmissivity
Sec. 1, T. 1 N., R. 25 W. SE¼NW¼NE¼ (or, ABD)	808	4.25	50,191 gpd/ft (6,710 ft^2/d)
Sec. 34, T. 1 N., R. 24 W. SW¼SW¼SE¼ (or, DCC)	1,050	6.8	40,764 gpd/ft (5,450 ft^2/d)
Sec. 25, T. 3 N., R. 26 W. NW¼NW¼NE¼ (or, ABB)	737	0.6	324,280 gpd/ft (43,353 ft^2/d)
		Average	138,412 gpd/ft 18,504 ft^2/d)

with about 5,800 mg/L of sodium and about 11,200 mg/L of chloride, making the water unsuitable for human or stock consumption, or for irrigation. Water from the Salt Fork Red River alluvium is suitable for stock water and irrigation, but is not suitable for human consumption because of large concentrations of dissolved solids.

Both the terrace deposits and the alluvial aquifers (including those along Sandy, Turkey, and Gypsum Creeks) locally overlie, and are in hydraulic connection with, the Blaine aquifer.

Underlying the Blaine aquifer is the Flowerpot Shale, an aquitard that locally yields only small quantities of water that commonly is brackish or saline. One water sample collected in 1987 from the Flowerpot Shale near Eldorado had a concentration of 82,900 mg/L of sodium and 194,100 mg/L of chloride (Runkle and Johnson, 1988).

AQUIFER RECHARGE

Natural recharge of the Blaine aquifer is dependent upon the amount of precipitation falling on outcrops of the Blaine Formation and lower Dog Creek Shale in the Hollis basin, and also upon percolation from streams that flow across the karst areas. After a rainfall, water can be seen flowing into many of the depressions and sinkholes. Many of the sinkholes receive surface runoff from drainage areas of about 10–100 acres. Many natural depressions, which may be sinks plugged with sediment and debris, contain shallow lakes and ponds after heavy rains: some of these lake and pond waters seep down to the zone of saturation.

To supplement this natural recharge, landowners have worked on a cooperative program of artificial recharge over the past 25 years. Landowners formed the Southwest Water and Soil Conservation District (SWSCD) to collect (from natural-gas suppliers) an override of 3% of the value of natural gas used to power pumps on irrigation wells. Most of the artificial-recharge projects funded by SWSCD have consisted of diverting surface drainage to sinkholes, caves, or other natural openings into the ground, thus allowing surface water to flow into the subsurface. However, in areas of few sinkholes, SWSCD drilled recharge wells 50–150 ft deep into underlying cavernous gypsum-dolomite units, or converted abandoned irrigation wells into recharge wells. These wells generally are constructed in or near local depressions or drainage ways, or the surface flow is diverted from drainage ways directly to the recharge wells. More than 75 separate projects have been completed by SWSCD to assist in artificial recharge of the aquifer.

In addition, the Oklahoma Water Resources Board plans to construct five new recharge wells to depths of 100–300 ft into cavernous zones of the Blaine aquifer, and will install several monitoring wells around each site to determine water-level fluctuations and water quality resulting from the recharge. The project, supported by the U.S. Bureau of Reclamation, will begin construction early in 1991, with all wells to be within 5 mi of Hollis.

PART 2

DESCRIPTIONS OF FIELD-TRIP STOPS

This section outlines a 1½-day field trip that covers principal geologic, stratigraphic, karstic, and hydrologic features of the Permian Blaine Formation, the major aquifer in the Hollis basin. Participants will begin in Dallas, Texas, at the end of the annual meeting of the Geological Society of America, on Thursday, November 1, and drive to Altus, Oklahoma. The field trip will be conducted on Friday and Saturday morning, with return to Dallas by about 6:00 p.m. Saturday.

The customary road-log method will not be used on this trip, but a brief summary of major features between stops is provided. The accompanying route map (Fig. 10 and inside of front cover) shows the route we will follow. The stratigraphic column (Fig. 3) has been reproduced on the inside of back cover for the convenience of the reader. Particularly useful companions to this field-trip guide are the geologic maps of the region (Carr and Bergman, 1976; Havens, 1977).

DAY 1

GEOLOGY BETWEEN ALTUS AND STOP 1

We depart Altus going north on U.S. 283 to the edge of the Wichita Mountains, and then turn west to Mangum. Our route most of the way to Mangum is upon the Permian Hennessey Shale, a thick red-bed shale that is about 300 ft below the Blaine Formation. The Hennessey contains thin, light-gray siltstone interbeds, and it dips gently to the west–southwest into the Hollis basin. Over much of the distance the Hennessey is mantled by 5–10 ft of soil. Much of the land in this area is irrigated by canals carrying water southward from Altus Lake, an impoundment just north of the Wichita Mountains.

To the north, the Hennessey Shale flanks Wichita Mountains peaks that rise 300–1,000 ft above the plains. These peaks consist of Cambrian (525 mya) granites that have been uplifted some 20,000–40,000 ft (Fig. 1) in Pennsylvanian and Early Permian times. The eroded mountains were then covered by Permian sediments (Hennessey Shale and younger strata), and the peaks are now exhumed by erosion of that Permian cover.

As we approach Mangum, we can see (to the southwest) low, east-facing escarpments capped by resistant beds of dolomite and gypsum in the Blaine Formation. In the lower part of the escarpment is the Flowerpot Shale which underlies the Blaine Formation.

We turn south from Mangum on Oklahoma Highway 34 to cross Salt Fork Red River. The alluvium here is about 0.5 mi wide and about 25–50 ft thick. In spite of the name "Salt Fork Red River," the water (when the river flows) is not salty: apparently it was named in error on original survey maps, and the name was probably intended for the salty Elm Fork Red River, just north of Mangum.

STOP 1

ELM FORK MEMBER AND MANGUM DOLOMITE BED, SOUTH OF MANGUM

Secs. 9 and 16, T. 4 N., R. 22 W., Greer County. The several sites at this stop are public-access areas. Do not cross fence lines onto private land.

This stop consists of several sites where the lower part of the Blaine Formation is well exposed and where there are examples of gypsum karst and porosity in dolomite beds (Fig. 11). The Elm Fork Member is about 80 ft thick in the area, and it is overlain by 2–3.5 ft of the Mangum Dolomite Bed at the base of the Van Vacter Member (Fig. 12). Strata here dip about 20 ft/mi to the west–southwest, toward the Hollis basin.

Cyclic deposition of the Blaine Formation is readily seen at this locality. Each cycle begins with a basal dolomite, which is overlain, successively, by beds of gypsum, red-brown shale, and green gray shale. Dolomite beds here are tan or light gray, fine crystalline, and 0.3–3.5 ft thick. Gypsum beds are white to light gray, fine crystalline, and 5.5–10.4 ft thick; gypsum beds are relatively thin in these outcrops on the east

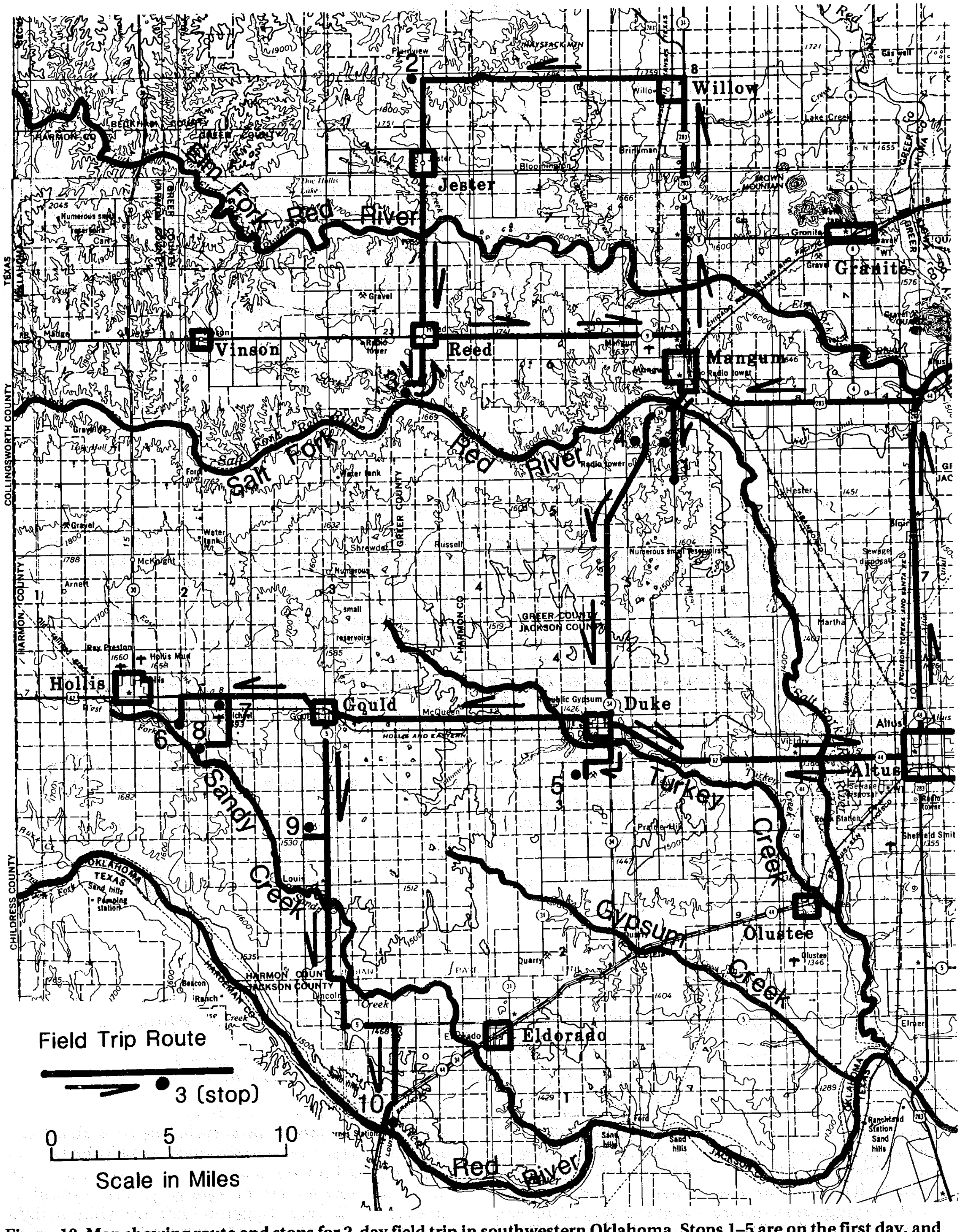


Figure 10. Map showing route and stops for 2-day field trip in southwestern Oklahoma. Stops 1–5 are on the first day, and stops 6–10 are on the second day. This figure is reproduced inside of front cover for ready reference.

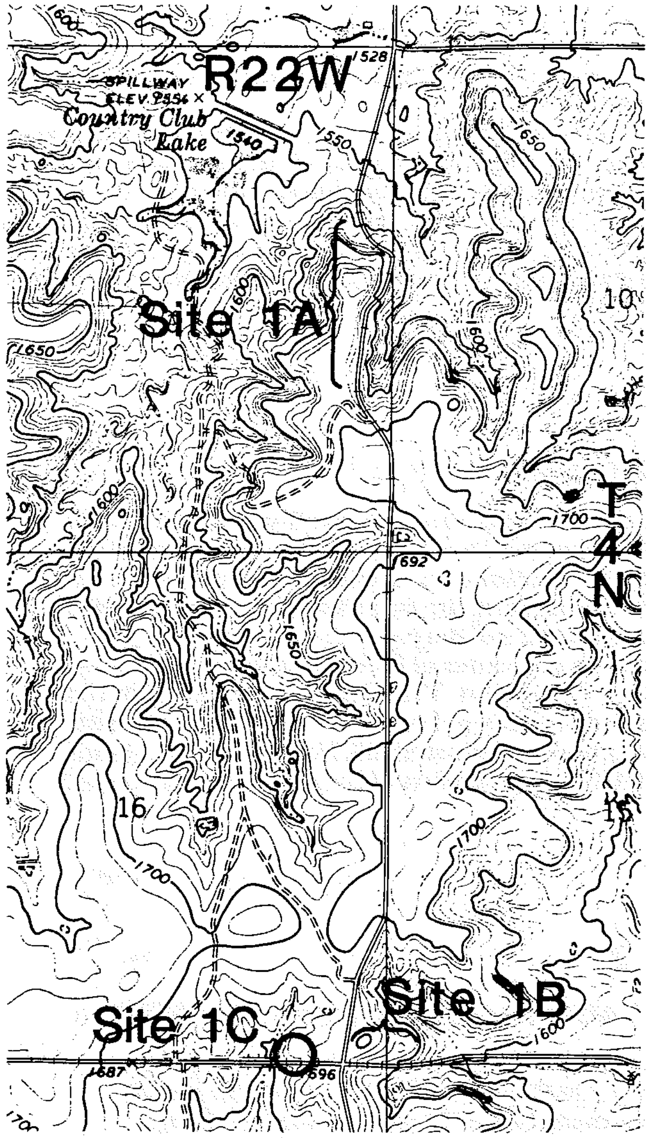


Figure 11. Location map for Stop 1, south of Mangum. Elm Fork Member and Mangum Dolomite Bed are well exposed in road cuts.

side of the Hollis basin, but they are 20–30 ft thick farther west. Red-brown shales, typically with a few thin interbeds of green-gray shale, are 10–22 ft thick. Green-gray shales at the top of each cycle are 1–2 ft thick.

Site 1A provides good exposures of the Elm Fork Member, and a good view of lateral persistence of the gypsum and dolomite beds that form benches just to the east across the stream valley. The upper part of the Flowerpot Shale is also well exposed. The Flowerpot is 200–300 ft of red-brown shale with thin interbeds of green-gray shale, siltstone, gypsum, and dolomite. The unit is an aquitard beneath the Blaine aquifer, and it underlies all parts of the Hollis basin.

The Haystack Gypsum Bed, about 10 ft thick, is the most persistent thick gypsum in the area. It has some karst development and locally has red-brown clay

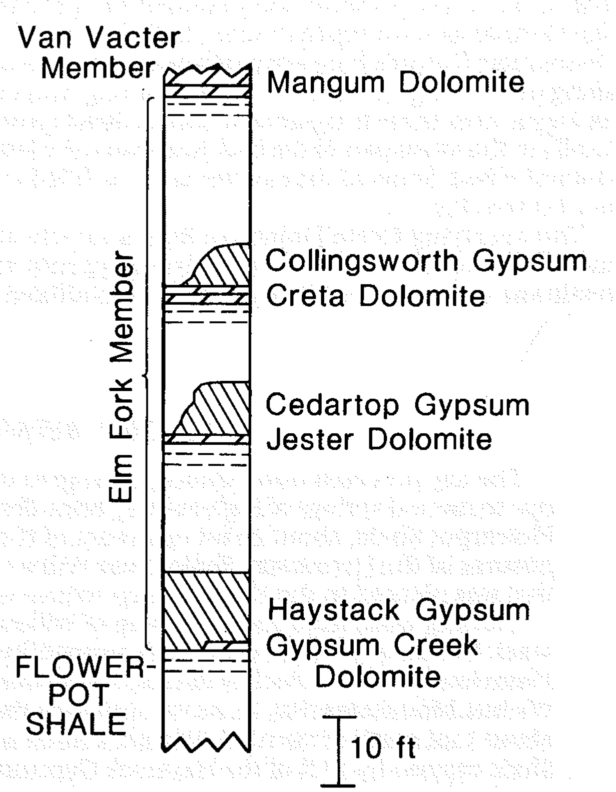


Figure 12. Stratigraphic column of the Elm Fork Member (Blaine Formation) at Stop 1 (data from Johnson, 1990, measured section H).

filling the dissolution cavities. The next higher gypsum, the Cedartop Gypsum Bed is about 5 ft thick and it has extensive karst features. Locally all the gypsum is dissolved, and cavities 1–5 ft wide are now filled with clay. In some areas the Cedartop Bed has a sculptured upper surface, showing that part of its dissolution began at the top of the bed.

Above the Cedartop Gypsum at Site 1A, the Creta Dolomite Bed has been partly recrystallized to limestone and is moderately to highly porous and permeable. The overlying Collingsworth Gypsum Bed is totally removed here, although residual masses of this unit are present throughout most parts of sec. 15, just to the southeast. The Mangum Dolomite Bed is highly disrupted and chaotic here due to dissolution of the underlying Collingsworth Gypsum. The Mangum is porous and permeable due to extensive recrystallization to limestone.

Absence of the Van Vacter gypsum beds (above the Mangum) and the Collingsworth Gypsum Bed illustrates the result of extensive dissolution of the gypsum beds by ground water. Dissolution is now working down into the next lower gypsum (the Cedartop Bed), and eventually will also remove much or all of the Haystack Bed.

Site 1B provides another good exposure of the Elm Fork Member, especially the Cedartop Gypsum Bed.

The Cedartop Bed here is about 8 ft thick, is well laminated, and contains some wavy lamination (probably due to expansion during hydration) in the bottom half. Dissolution features here are partly vertical, probably along preexisting fractures, and partly along irregular passages. Also, there is significant cavern development locally in the lower part of the bed, just above the Jester Dolomite Bed. Some of the cavities are now filled with red-brown clay.

The overlying Creta Dolomite Bed is locally disturbed due to dissolution of underlying gypsum and resultant collapse, and it is partly recrystallized to limestone. About 5 ft of Collingsworth Gypsum is preserved above the Creta, but its local dissolution has caused collapse of the overlying Mangum Dolomite Bed.

Site 1C has a good exposure of the Mangum Dolomite Bed. The unit is up to 3 ft thick, largely recrystallized, and it is slumped due to dissolution of underlying gypsum beds. The extensive recrystallization probably results from this unit having been the principal conduit for local ground-water movement in the past.

GEOLOGY BETWEEN STOP 1 AND STOP 2

The trip proceeds north through Mangum onto U.S. 283. We cross Elm Fork Red River, which *is* salty due to natural springs of high-salinity brine derived by dissolution of bedded salt in the upper part of the Flowerpot Shale, about 25 mi upstream of the crossing. We continue on to Willow, crossing more exposures of the Hennessey Shale. Near Willow is a road sign marking the first shelterbelt (row of trees) that was planted in the 1930s to help reduce wind erosion of the soil during dust-bowl days.

Turning west, from the north side of Willow, we travel on exposures of Flowerpot Shale, cross Haystack Creek, and approach bluffs capped by the Blaine Formation. The Blaine is preserved in the Plainview syncline, which is the surface expression of a deep-seated graben on the crest of the "buried" Wichita Mountains (Fig. 2). As we approach the bluffs, a prominent isolated hill, Haystack Butte, stands about 1 mi north of the road. Haystack Butte is about 190 ft high; it consists of about 173 ft of Flowerpot Shale capped by 17 ft of the Haystack Gypsum Bed.

STOP 2

JESTER CAVE AND EXPOSURES OF VAN VACTER MEMBER, NORTH OF JESTER

Secs. 23 and 26, T. 7 N., R. 24 W., Greer County. You must obtain permission for access to property: for sec. 23, contact Mr. Foy Rogers, Rt. 1, Willow, OK 73673 (phone: 405/683-4384); for sec. 26, contact Dr. Loyd Harris, 2514 S. Pickard Ave., Norman, OK 73072 (phone: 405/321-2990).

Jester Cave (Fig. 13) provides a unique opportunity to enter and examine a gypsum-karst system similar to that which is yielding ground water farther south in the Hollis basin. The cave was surveyed and studied by members of the Central Oklahoma Grotto between 1983 and 1987 (Bozeman and others, 1987). The surveyed length of the main passage is 7,918 ft, but along with the side passages the total length is 33,023 ft, making Jester Cave the longest known cave in Oklahoma and the longest gypsum cave in the western world. A proposal currently under review would designate Jester Cave as a "National Natural Landmark."

Jester Cave is located in the Plainview syncline, above the crest of the "buried" Wichita Mountains (Fig. 2). In this area the entire Blaine Formation is about 150 ft thick, and the Van Vacter Member (in which the cave system is developed) is about 80 ft thick (Fig. 14). The full thickness of the Van Vacter Member is not exposed at any one site in the immediate vicinity of the cave, but overlapping segments of the stratigraphic sequence are exposed in various cave openings, sinkholes, and bluffs along the course of the cave system. Dolomite beds at the base of units 1, 2, 3, and 6 are about 0.3–3.5 ft thick, gypsum beds are about 3.5–19.5 ft thick, and the shale interbeds are about 0.1–3.5 ft thick.

Almost all parts of the main cave system are developed in unit 1 gypsum, just above the Mangum Dolomite Bed (Fig. 15 and cover photo). The Mangum in this area is tan to light-gray dolomite 1.5–2.0 ft thick. In places it is pelletoidal and/or oolitic. In most parts of the cave, the Mangum is either the floor of the cave or is just a few feet below the floor (Fig. 15). The Mangum is well exposed at and near the Main Entrance, and it is the bedrock ledge that forms Pouroff falls, just west of the Main Entrance, where it is at an elevation of 1,794 ft.

Directly above the Mangum is gypsum bed 1, consisting of about 16 ft of white, fine- to coarse-crystalline gypsum. The lower part of the bed typically has "nodular" or "chicken-wire" texture, where gray films or seams of dolomite isolate nodules of gypsum 0.5–2.0 in. across. Chicken-wire texture resulted when early growth of gypsum nodules in the upper part of

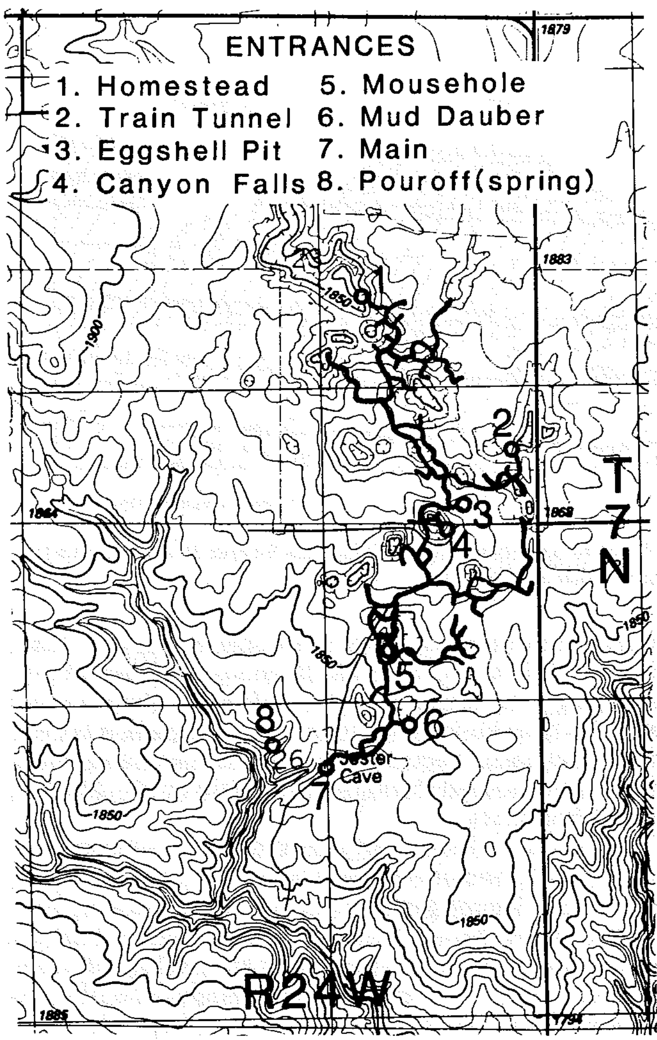


Figure 13. Location map for Jester Cave at Stop 2, north of Jester. Heavy lines show approximately the passages that have been surveyed (Bozeman and others, 1987).

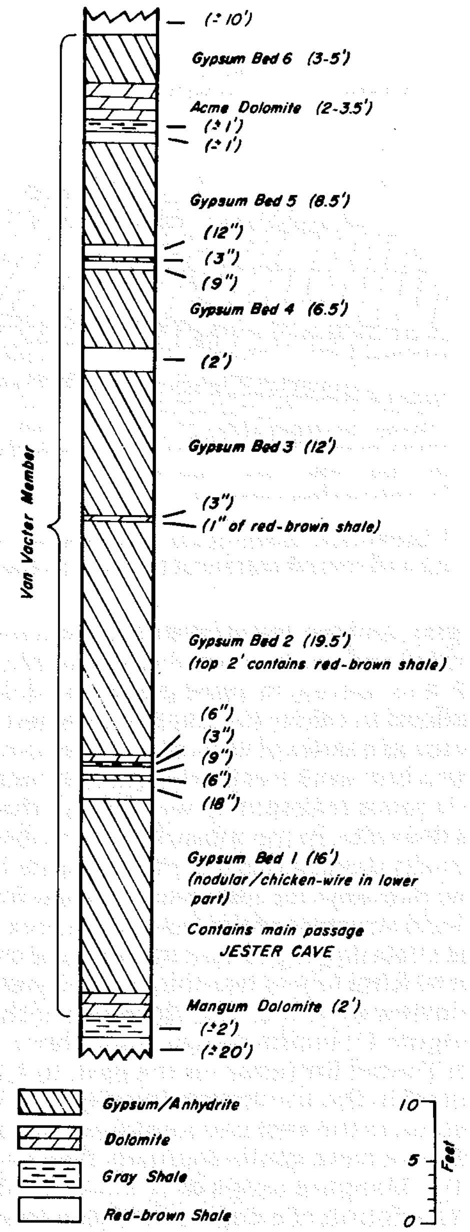


Figure 14. Stratigraphic column of the Van Vacter Member (Blaine Formation) in the Jester Cave area, Stop 2 (Bozeman and others, 1987).

the dolomite, just beneath the seafloor surface, caused thin seams of dolomite to be pushed aside during nodule growth.

Most of the sinkhole entrances to the cave are developed in strata above gypsum bed 1, but these entryways invariably lead down to the main cave in unit 1. Among the more interesting sinks and other sites in the Jester Cave area are (from north to south) Homestead, Train Tunnel, Eggshell Pit, Canyon Falls, Mousehole, Mud Dauber, and Main Entrances, along with the perennial spring at the Pouroff (Fig. 13).

Much of the Jester Cave system consists of angulate passages, characterized by sharp bends with intermediate straight sections; portions of the cave system are sinuous, with broad sweeping curves and few straight stretches. Although the cross section of passages is quite variable, most of Jester Cave has elliptical-type passages (see front cover). The main passage typically is 20–50 ft wide and 5–20 ft high; the side passages typically are 5–10 ft wide and 2–10 ft high (Bozeman and others, 1987).

Jester Cave is air-filled along part of its course, although there are some segments where pools of standing water partly or totally fill the passages. In times of moderate to heavy rainfall, most of the passages can be filled with flowing water; at such dangerous times no part of the cave system should be entered.

Early-stage dissolution of dolomite and development of secondary porosity and permeability can be seen in the Acme Dolomite Bed (unit 6) immediately above, and to the south of, Eggshell Pit Entrance (Fig. 13). The Acme Bed consists of 2.0–3.5 ft of light-gray,

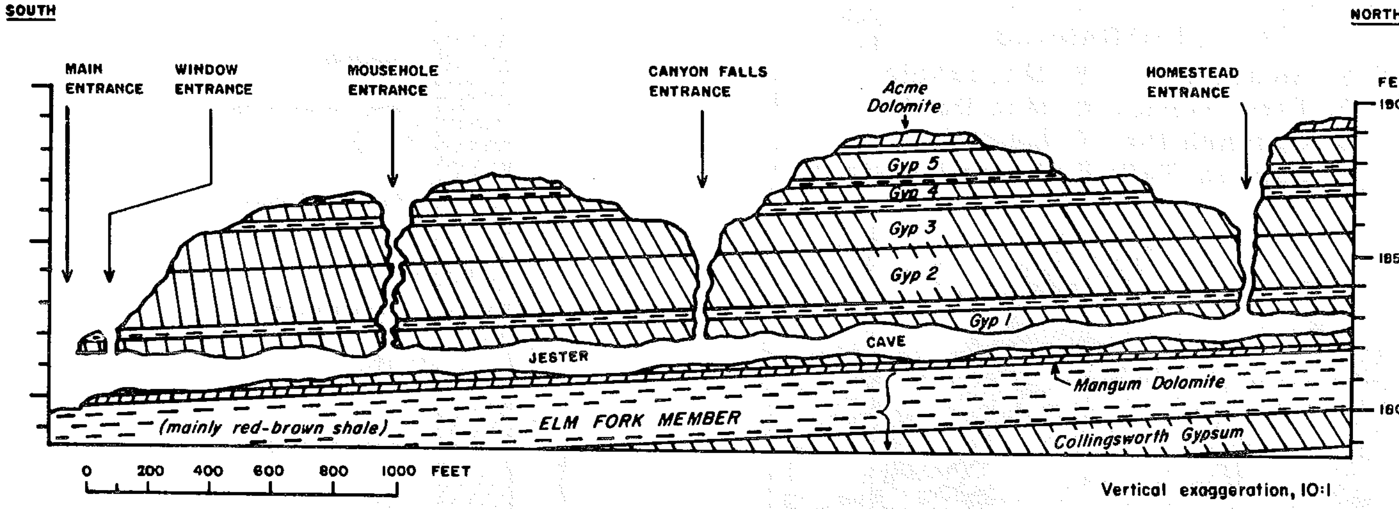


Figure 15. Schematic north–south cross section showing Jester Cave in gypsum bed 1 of the Van Vacter Member (Blaine Formation) and several entrances to the cave system (Bozeman and others, 1987).

yellow-gray, and tan, microcrystalline dolomite that is thin bedded and tends to weather out as thin slabs or plates 2–6 in. across. In some places the dolomite is recrystallized to calcite (or limestone): where the calcite occurs as a series of vertical and horizontal veins, it creates a box-work weathering pattern because the calcite is more resistant to weathering than is the original dolomite. In the subsurface, such dissolution and porosity development make dolomite beds the preferred pathways for early ground-water flow.

The local structure of the Jester Cave area consists of Blaine strata dipping toward the west and southwest at about 60 ft/mi (about two-thirds of a degree) within the Plainview syncline. The elevation of the top of the Mangum Dolomite ranges from about 1,840 ft, at Train Tunnel Entrance on the east, to 1,794 ft at the Pouroff in the southwest. In spite of this dip of Blaine strata to the west and southwest, the cave system follows a more gentle southerly flow path along which the Mangum drops only about 20–25 ft/mi (about one-fourth of a degree) between Homestead Entrance on the north and the Main Entrance on the south (Fig. 15).

A cursory examination of the alignment of various segments of the Jester Cave system suggests that the course of the cave's flow path may be controlled by several sets of joints or fractures in the rock (Bozeman and others, 1987). The preferred orientation of the principal segments along the main course of the cave is between N10°W and N30°W, and it appears that the main cave follows a set of joints with this range of orientations. Another prominent orientation of cave segments is N60°E, and several lesser orientations are N30°E and N60°W.

The age of development of Jester Cave is not known, but it may have begun at any time within the span of several thousand years ago to perhaps 600,000 years ago. Caves are now forming at moderate depths below the water table elsewhere in the region, inasmuch as ground water is flowing through open cavities in the Blaine aquifer at depths of 100–200 ft below the existing land surface and water table near Hollis and Duke, in the Hollis basin, some 25–30 mi south of Jester Cave. Although the size and extent of these deep-seated cavities near Hollis and Duke are not known, it is clear that they could become the caves of tomorrow when they are exhumed through the erosion of 100–200 ft of overlying strata.

In a like manner, early development of openings that finally became Jester Cave may have occurred when Van Vacter gypsum bed 1 was as much as 100–200 ft below the land surface. Gustavson and others (1980) estimate that it has taken about 600,000 years for the Canadian River to erode through 200 ft of strata in the nearby Texas Panhandle. If we assume the same rate of downcutting in the Jester area, then gypsum bed 1 would probably have been 100–200 ft deep some 300,000–600,000 years ago. Thus it is possible that the earliest development of cavities in the Jester Cave system may have occurred as much as 600,000 years ago. It is possible, however, that the major development may have occurred in the past 50,000 years or so, when the nearby streams probably cut down below the topographic level of the cave and permitted diversion of surface waters through sinkholes into the subterranean cavern above the water table.

A final note about Jester Cave is the current proposal before the National Parks Service, U.S. Department of the Interior, to designate Jester Cave as a "National Natural Landmark." This is a well-deserved status, inasmuch as the area clearly meets the following National Parks Service definition:

> "National Natural Landmark" means an area of national significance, located within the boundaries of the United States or on the Outer Continental Shelf, designated by the Secretary of the Interior because it contains an outstanding representative example(s) of the Nation's natural heritage, including terrestrial communities, aquatic communities, landforms, geological features, habitats of native plant and animal species, or fossil evidence of the development of life on earth.

GEOLOGY BETWEEN STOPS 2 AND 3

We turn south, traveling through Jester and on to Reed. Between Jester and Reed we traverse terrace deposits that mantle the Flowerpot and Hennessey Shales, and also cross (again) the salty Elm Fork Red River. About 1,000 ft south of Elm Fork Red River is Jay Buckle Spring, on the east side of the road. This is a well-known fresh-water spring that issues from terrace deposits overlying the Hennessey Shale.

We continue 2 mi south and 0.5 mi west of Reed. From 1.0 to 1.5 mi south of Reed the road crosses the lower part of the Dog Creek Shale. The Dog Creek is mostly red-brown shale, but it contains several thin green-gray shales, tan dolomites, and white gypsums. The dip here is about 60 ft/mi to the south.

STOP 3

HORSESHOE VALLEY CAVE AND EXPOSURES OF VAN VACTER MEMBER, SOUTH OF REED

N½ sec. 35, T. 5 N., R. 24 W., Greer County. You must obtain permission for access to property from Mr. Rance Ellis, 801 W. Electra, Mangum, OK 73554 (phone: 405/782-2853).

Horseshoe Valley (Fig. 16), a small, flat-floored valley just north of Cave Creek, contains a number of karst features similar to those at Jester Cave (Stop 2), but the explored portion of the cave system is much smaller than at Jester Cave. A survey of Horseshoe Valley Cave was started by members of the Central Oklahoma Grotto in 1989, and thus the data available are only preliminary. At present, about 4,500 ft of passages have been surveyed and mapped, with the major entrances to the south in Horseshoe Valley and where the spring flows into Cave Creek. Additional entrances have been discovered in some of the sinkholes farther north and northeast in secs. 35 and 26, but their connection to the main cave has not yet been surveyed.

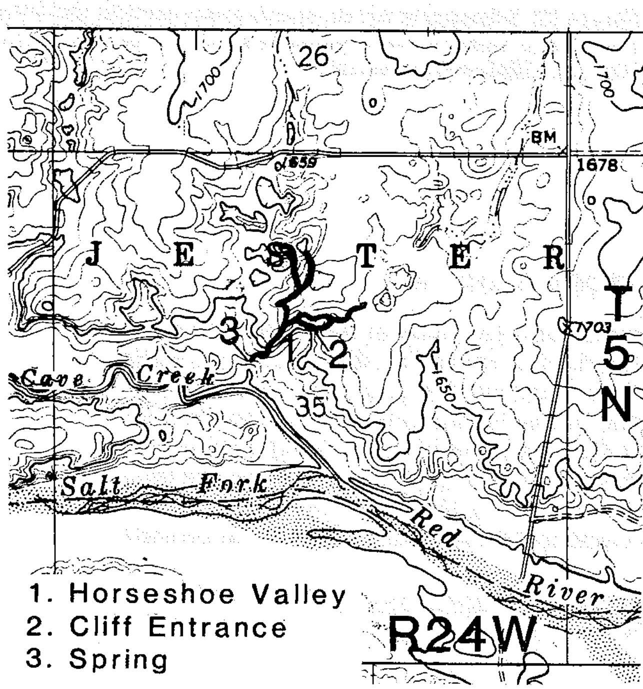


Figure 16. Location map for Horseshoe Valley Cave at Stop 3, south of Reed. Heavy lines show approximately the passages that have been surveyed.

Horseshoe Valley Cave is on the northern flank of the Hollis basin (Fig. 2). The entire Blaine Formation is about 190 ft thick in the area, and the Van Vacter Member is about 92 ft thick just 3 mi west of the cave (Johnson, 1990, measured section G). Most Van Vacter gypsum beds in the cave area are 8–18 ft thick, dolomite beds are commonly 0.5–2.5 ft thick, and the shale interbeds range from 1 to 5 ft thick (the shale and medial gypsum at the top of unit 1 is thickest). The detailed stratigraphy is quite similar to that shown at Jester Cave (Fig. 14).

All surveyed passages of the cave are in gypsum bed 1, above the Mangum Dolomite, except just inside Cliff Entrance (Fig. 17) where roof collapse has exposed unit 1 shale and the dolomite and gypsum at the bottom of unit 2 (Fig. 18). Domes and crevices extend upward into units 2 and 3 elsewhere in the cave, but there are no known horizontal passages in these beds.

Parts of Horseshoe Valley Cave are air-filled, although pools of standing water partly fill some of the passages throughout the year. Most of the cave fills with flowing water during and after moderate to heavy rainfall, and the cave should not be entered at these times!

The explored portion of Horseshoe Valley Cave consists mainly of sinuous passages 10–40 ft wide and 5–20 ft high. A large part of the north arm of the cave is used as a summer maternity site by at least 50,000 *Myotis velifer* (commonly known as "cave bat").

Strata dip to the south and southeast at 20–40 ft/mi. Although there is no evidence of joints or fractures in the rock, there is a slight preference for cave segments to be oriented between N30°E and N40°E.

The only dye test conducted so far was at the east end of the eastern arm, where 1 pound of fluorescein

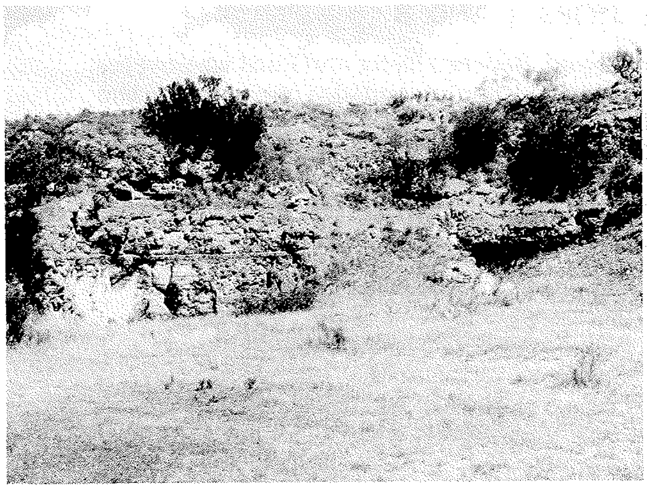


Figure 17. View of northeast end of Horseshoe Valley showing Van Vacter units 2 through 5 and location of Cliff Entrance (base of cliff on right).

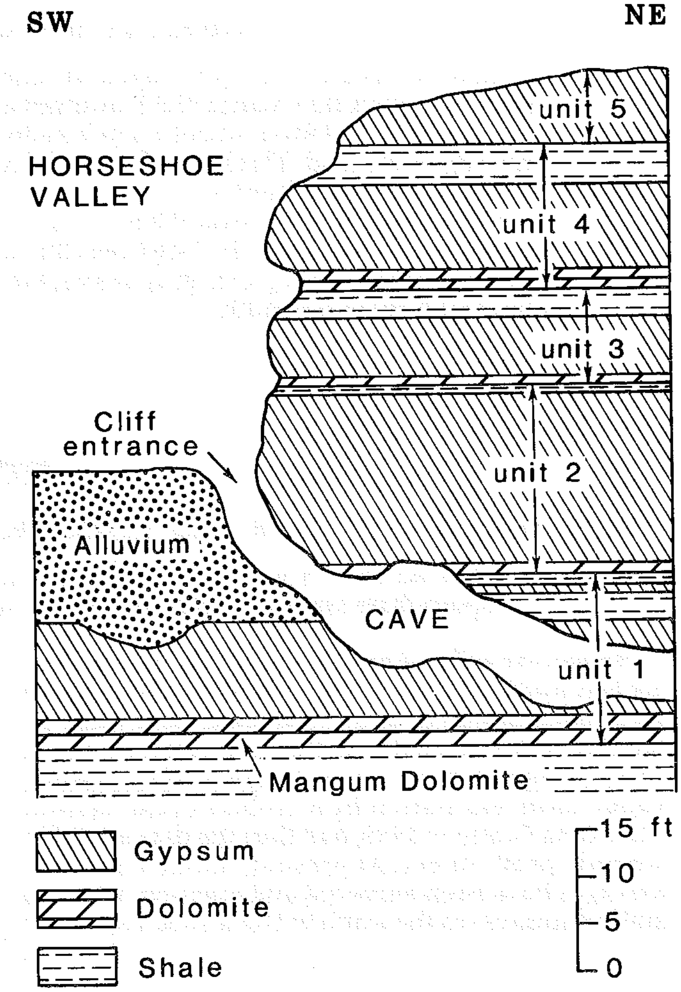


Figure 18. Schematic north–south cross section showing Horseshoe Valley Cave in gypsum bed 1 of the Van Vacter Member (Blaine Formation).

dye was added to water disappearing below the main passage in gypsum bed 1. Monitoring was conducted for 72 hours at the spring in the southwest and along the banks of Cave Creek, but no trace of the dye was detected. It seems likely that the dye was carried through unknown passages to the south; it probably was emitted into the alluvium of Cave Creek or Salt Fork Red River, and did not surface until more than 72 hours after release.

Sites of importance in the area are the Cliff Entrance in Horseshoe Valley, the spring in the southwest, and the several sinkholes and depressions to the north and northeast of Horseshoe Valley (Fig. 16).

GEOLOGY BETWEEN STOPS 3 AND 4

Retrace our route to Reed, and then turn east on Oklahoma Highway 9 toward Mangum. South of the highway, for the first 5 mi east of Reed, is an irrigation district where wells produce water from karst zones in the Blaine and from the overlying terrace. East of the irrigation district are poorly exposed remnants of the Blaine Formation; most of the gypsum beds have been dissolved, leaving mainly red-bed shales and highly weathered dolomites.

Just north of Mangum are four wind generators set up to supply electricity to the city. Mangum purchases most of its electricity from the Oklahoma Municipal Power Authority, but the city also has an old diesel-powered power plant and these wind generators to provide some electricity.

We then turn south on Oklahoma Highway 34 through Mangum, cross Salt Fork Red River, and stop at the road cuts about 2 mi southwest of the river.

STOP 4

BLAINE FORMATION WITH EXTENSIVE DISSOLUTION OF GYPSUM BEDS, SOUTHWEST OF MANGUM

Oklahoma Highway 34 road cuts in E½ sec. 8, T. 4 N., R. 22 W., Greer County. The road cuts are public-access areas, as long as you don't cross fence lines onto private land.

Regional studies show that the Cedartop, Collingsworth, and Van Vacter gypsums all were undoubtedly deposited at Stop 4 (Fig. 19), but now all these gypsum units are absent from the road cuts due to dissolution. The Elm Fork Member here can be compared to the better-preserved gypsum exposures at Stop 1, about 1 mi to the east, and the Van Vacter Member can be compared to gypsum exposures at Stop 3 (10 mi to the west) and at Stop 5 (14 mi to the south).

The Haystack Gypsum Bed, at the base of the Blaine, is well preserved at Stop 4, and the focus here is on the overlying strata. The Jester Dolomite is extensively recrystallized in the west road cut, and the overlying Cedartop Gypsum Bed is absent. Between the Jester and Creta Dolomite Beds is about 12 ft of apparently undisturbed red-brown and green-gray shale. Such preservation of this shale indicates that Cedartop Gypsum was removed here by sheet-like dissolution that allowed a gradual lowering of the overlying shale into the space originally occupied by gypsum. Remnants of Cedartop Gypsum up to 5 ft thick are still present locally in the bench below the Creta Dolomite about 0.5 mi to the north–northeast. The Collingsworth Gypsum Bed also is missing from exposures on the west side of the highway, although residual masses of the gypsum remain locally to the east of the highway. The Mangum Dolomite is highly weathered, recrystallized, and contains some pink, secondary calcite.

Above the Mangum Dolomite Bed, all gypsum beds of the Van Vacter Member are missing. Remnants of the Acme Dolomite Bed can be seen at the top of small bluffs in an amphitheater on the east side of the highway and just north of a house on the west side of the highway. The Acme Bed here consists of chaotic blocks, 1–2 ft thick, that are poorly exposed and are highly recrystallized and honeycombed; scattered blocks of the Acme Bed are about 30 ft above the Mangum Bed. Material between the Mangum and Acme Beds is a chaotic mix of Van Vacter shale interbeds along with Quaternary cave-filling sediment.

Just above the Acme Bed, and for the next 4 mi to the south along the highway, is a terrace deposit left behind by Salt Fork Red River. This attests that the river previously had flowed directly above Stop 4 (and also above Stop 1), and the abundant ground water associated with the river was the most likely cause of extensive gypsum dissolution at these sites.

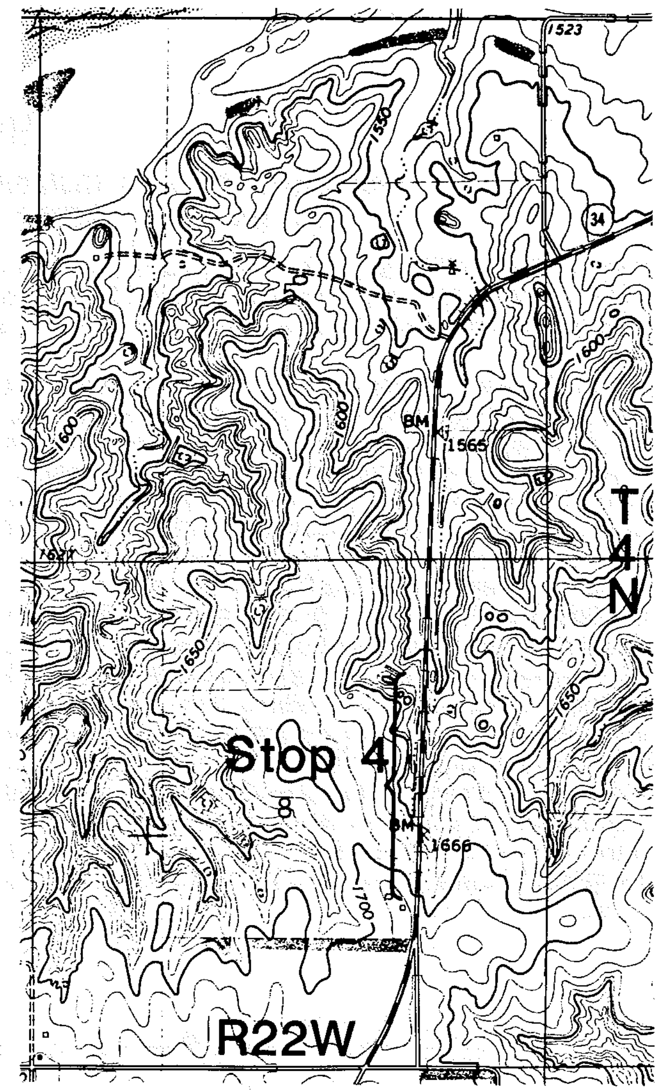


Figure 19. Location map for Stop 4, southwest of Mangum. Blaine Formation is well exposed in road cuts.

GEOLOGY BETWEEN STOP 4 AND STOP 5

The trip proceeds south on Oklahoma Highway 34 to the Republic Gypsum Co. quarry, 2 mi south of Duke. The first 4 mi of the trip is upon sand and gravel terrace deposits of Salt Fork Red River, and the next 4 mi is along the dip slope of the Creta Dolomite, which dips 20–100 ft/mi to the southwest into the Hollis basin. The last 5 mi traverses farmland where Van Vacter gypsums are partly dissolved and typically are covered by 5–10 ft of soil; gypsum is exposed only in a few stream cuts and on several of the hills.

STOP 5

REPUBLIC GYPSUM COMPANY QUARRY AND EXPOSURES OF THE VAN VACTER MEMBER, SOUTH OF DUKE

W½ sec. 24 and E½ sec. 23, T. 2 N., R. 23 W., Jackson County. Permission for access to the gypsum quarry must be obtained from Mr. Odell Miranda, Quarry Superintendent, Republic Gypsum Co., Drawer "C," Duke, OK 73532 (phone: 405/679-3626, -3553, or -3391). The company's wallboard plant and main offices are 0.5 mi west of Duke.

The Republic Gypsum Co. quarry is located 2 mi south of Duke, just west of Oklahoma Highway 34 (Fig. 20). Gypsum beds 1, 2, and 4 (bed 3 is absent) of the Van Vacter Member are being mined for the manufacture of wallboard at Duke. The many fine exposures in the quarry contain dissolution cavities, most of which are filled with Quaternary clays and silts.

Gypsum bed 1, which is 9 ft thick, is overlain successively by 8 ft of shale, 1 ft of dolomite, 18 ft of gypsum bed 2 (the main bed being mined), 9 ft of shale, 0.5 ft of dolomite, and 9 ft of gypsum bed 4 (Fig. 21). All 3 gypsum beds consist of white, fine- to medium-crystalline gypsum; they generally are 96–98% pure (Johnson and Denison, 1973), but the plant feed is diluted to 93–94% by accidental inclusion of clay and soil that fill solution cavities. Republic Gypsum Co. produces about 360,000 tons of gypsum per year, and is the second-largest gypsum-producing company in the State.

The quarry was first opened in 1964 in gypsum bed 2, and subsequently gypsum bed 4 was recovered. Until 1987, mining also extended down to include gypsum bed 1, just above the Mangum Dolomite Bed, because the local water table was below the base of that gypsum. Increased rainfall and reduced groundwater production over the past 3 years has raised the water table at least 10 ft, and now gypsum bed 1 is totally submerged in the pit; thus mining is now confined to beds 2 and 4.

Each of the gypsum beds locally contains irregular, clay-filled dissolution cavities and sinkholes, and the base of each bed is somewhat uneven, due to dissolution. This is particularly true for bed 2. The discovery of partial remains of a Columbian elephant (with a possible age range of 10,000–50,000 years ago) in a debris-filled sinkhole that cuts bed 2 indicates that at least some of the cavern development and filling occurred in the Late Pleistocene.

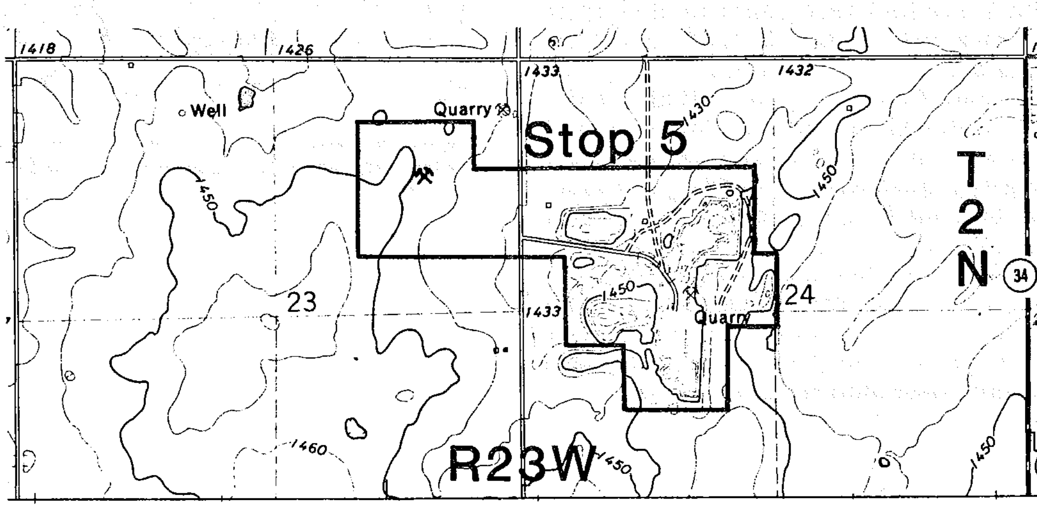


Figure 20. Location of Republic Gypsum Co. quarry, south of Duke.

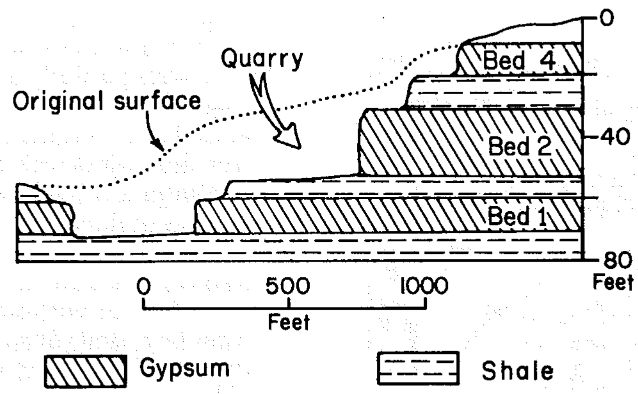


Figure 21. Schematic cross section showing Van Vacter gypsum beds and quarrying activities at Stop 5.

GEOLOGY BETWEEN STOP 5 AND ALTUS

We return to Duke, and then turn east on U.S. 62 to Altus. For the first 3 mi east of Duke we traverse the soil-covered Van Vacter Member, and for the next 3 mi we cross the Elm Fork Member: gypsum in the Blaine Formation has been dissolved or is mantled by soil along this route. After leaving the Blaine Formation, we descend through the Flowerpot and Hennessey Shales. These Shales are mantled by a 4-mi-wide strip of alluvium and terrace deposits of Salt Fork Red River.

DAY 2

GEOLOGY BETWEEN ALTUS AND STOP 6

This morning we drive west on U.S. 62 across the Hollis basin (Fig. 8, cross section *Y–Y′*). We retrace our route to Duke, and can see gypsum exposed in the banks of Turkey Creek just west of the Republic Gypsum Co. wallboard plant. Just 0.5 mi west of Turkey Creek we pass onto outcrops of the Dog Creek Shale and remain on that formation until we reach Hollis.

The Dog Creek Shale is a red-bed shale containing some thin beds of green-gray shale, dolomite, and gypsum. The several dolomite and gypsum beds are in the lower 50 ft of the formation, and the gypsum-karst features make this a leaky confining unit above the Blaine aquifer. In the vicinity of Gould a maximum of 180 ft of Dog Creek Shale overlies the Blaine Formation. At this great depth, the Blaine sulfate beds are mainly anhydrite (not gypsum), and in this area there is only minor ground water in the Blaine (Fig. 8, cross section *Y–Y′*).

STOP 6

IRRIGATION WELL, SOUTHEAST OF HOLLIS

NE¼ sec. 12, T. 2 N., R. 26 W., Harmon County. This site is owned by Paul Horton, Box 745, Hollis, OK 73550 (phone: 405/688-9223 or -2392). He allows public access, but care must be taken to not damage the fence or equipment.

This well, located 2 mi east and 1 mi south of Hollis in the northeast corner of the section (Fig. 22), is representative of the irrigation systems in the Hollis basin area. Ground level here is 10–20 ft above the top of the Blaine–Dog Creek contact, although the Dog Creek Shale is mantled by soil. The well is 200 ft deep and the bottom of the borehole is probably just above the Haystack Gypsum Bed. Thus almost the entire Blaine aquifer is penetrated, and water probably is entering from several of the dolomite and gypsum units. The Mangum Dolomite Bed is about 110 ft deep in this well.

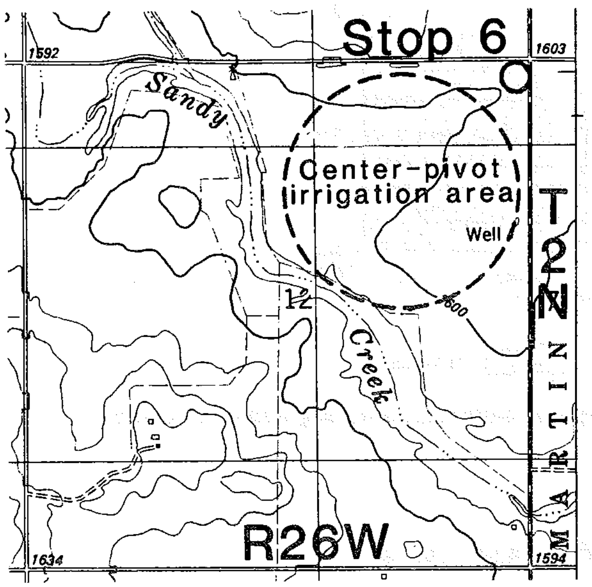


Figure 22. Location of irrigation well at Stop 6, just southeast of Hollis.

Water-level fluctuations and water-quality data have been collected on this well as part of a monitoring program of the U.S. Geological Survey (Runkle and others, in review). The water level in the well has fluctuated from depths of 17–42 ft below ground level (between June 1986 and March 1988), and the water contains 3,300 mg/L calcium sulfate, 481 mg/L sodium chloride, and 3,566 mg/L residual solids, evaporated at 105°C.

Stop 6 is about 0.5 mi from West Fork Sandy Creek, and some of the Blaine gypsum and dolomite beds have been partially or totally dissolved due to the extensive interaction of ground water and surface water. Wells drilled within 3 mi of the main creeks typically have dolomites with high permeability and gypsums with large conduits that allow large quantities of water to move rapidly through the aquifer.

An 8-in. turbine pump, powered by natural gas, is used on this well, and it is capable of pumping 1,100 gpm of water without lowering the water level. The water level does drop as much as 15 ft during July and August, when many wells in the area are being used and the regional water level declines, but the aquifer is so permeable at this site that the pumping level and the static level of water in the well are the same.

Water is pumped to a center-pivot irrigation system through an 8-in. diameter pipe, buried at a depth of 48 in. This center-pivot system is 0.25 mi long; it irrigates 120 acres in a circular plot on 160 acres (one-quarter of a section). Water is applied 3 or 4 times between July 10 and September 1, at a rate of 3–4 in. per application. Row irrigation is used on about 85% of the irrigated lands in the Hollis area, center-pivot irrigation is used on about 10%, and side-roll irrigation is used on about 5% (Paul Horton, personal communication).

Using 60 pounds of nitrate and 60 pounds of phosphate per acre, irrigated land at this stop has yielded 2–2.5 bales of cotton per acre over the past 3 years; in earlier years it yielded about 40 bushels of wheat per acre. This well also provides water for another center-pivot irrigation system in the SE¼ of the section, where about 4,100 pounds per acre of cleaned and dried peanuts were grown in 1989.

GEOLOGY BETWEEN STOP 6 AND STOP 7

Return to U.S. 62, go east for 2 mi, and then south 0.25 mi. Along this short route we are traveling on lower Dog Creek Shale, mantled by soil.

STOP 7

RECHARGE WELL, EAST OF HOLLIS

Center E½NE¼ sec. 5, T. 2 N., R. 25 W., Harmon County. This site is available to public access, but care must be taken to not damage or obstruct surface facilities.

A number of recharge wells in the Hollis area are currently funneling excess precipitation down into the Blaine aquifer. Some wells were drilled specifically for this purpose, and others are old irrigation wells that have been converted to recharge wells. Stop 7 is at a recharge well drilled in the early 1970s by the Southwest Water and Soil Conservation District (SWSCD). It is located 4 mi east of Hollis (Fig. 23), although any other recharge well could be visited on future field trips. Also, by late 1991 there should be five new recharge wells in the Hollis area that will be part of an artificial-recharge-demonstration project being carried out by the Oklahoma Water Resources Board and the U.S. Bureau of Reclamation. A schematic diagram for existing recharge wells in the area is given in Figure 24.

The well at Stop 7 is 95 ft deep, and the land-surface elevation is about 1,602 ft. Outcropping strata (beneath the soil cover) at the site are in the lower part of the Dog Creek Shale, about 40 ft above the top of the Blaine Formation. The bottom of the hole is about 37 ft above the Mangum Dolomite Bed (which here has an elevation of about 1,470 ft), and thus the well is completed near the middle of unit 3 gypsum of the Van Vacter Member. Blaine and Dog Creek strata here dip gently to the northeast at 20 ft/mi, toward the central part of the Hollis basin.

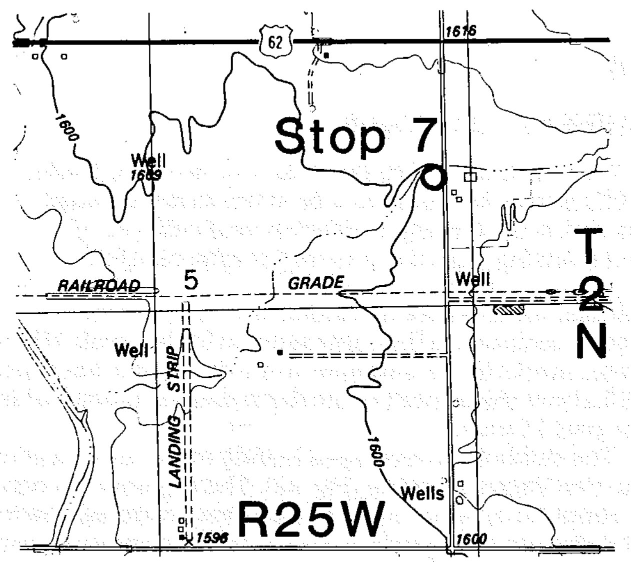


Figure 23. Location of recharge well at Stop 7, just east of Hollis.

In all likelihood, cavities in gypsum bed 3 at this well are interconnected with similar gypsum cavities and permeable dolomites in other units of the Van Vacter, and recharge of gypsum bed 3 is also locally recharging the entire Van Vacter portion of the Blaine aquifer. The interconnection within the Van Vacter results from an intricate network of caves, cavities, sinks, and dissolution-collapse structures that are parallel to, and locally cut across, bedding planes in the subsurface.

A shallow excavation has been made here along a small stream bed, and an earthen embankment has been constructed around the intake area so as to impound runoff. A screen has been placed over the 14-in. intake pipe to prevent debris from entering the well. The drainage area supplying runoff to this well is about 1 mi^2, although several surface impoundments capture some of the runoff in the upper half of the drainage area.

The water level in this well has been monitored from May 1986 through March 1988, and its depth has ranged from 31 to 76 ft (Fig. 25). The two lowest levels occurred in August of 1986 and 1987, during the dry season when nearby irrigation wells were being used (the nearest wells are about 1,200 ft to the south and northwest). The water level rebounded sharply each year with cessation of irrigation, and then more gradually during the winter and spring as precipitation recharged the aquifer. Water flows into the well under the influence of gravity, and the flow rate is probably as much as 8,000–12,000 gpm.

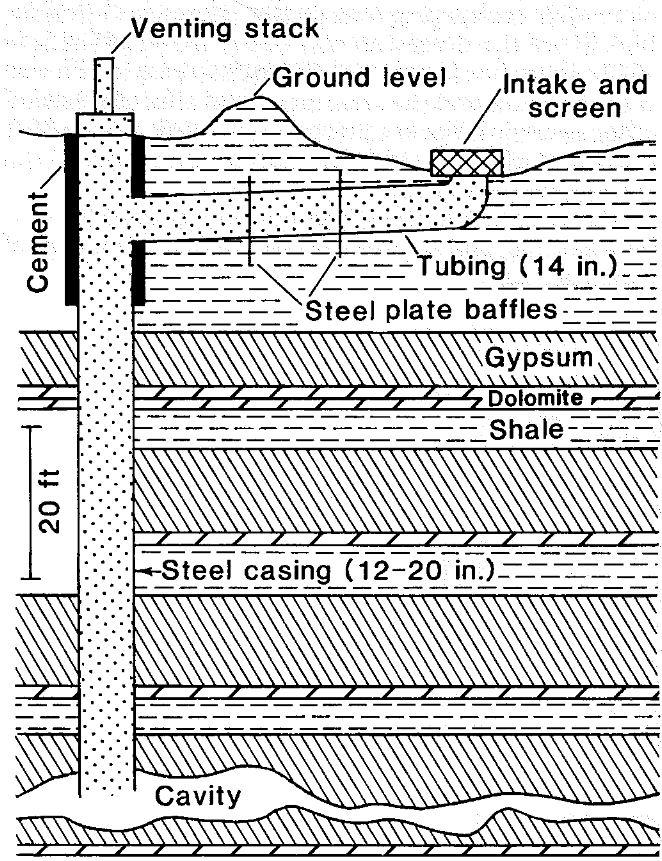


Figure 24. Schematic diagram of a typical recharge well.

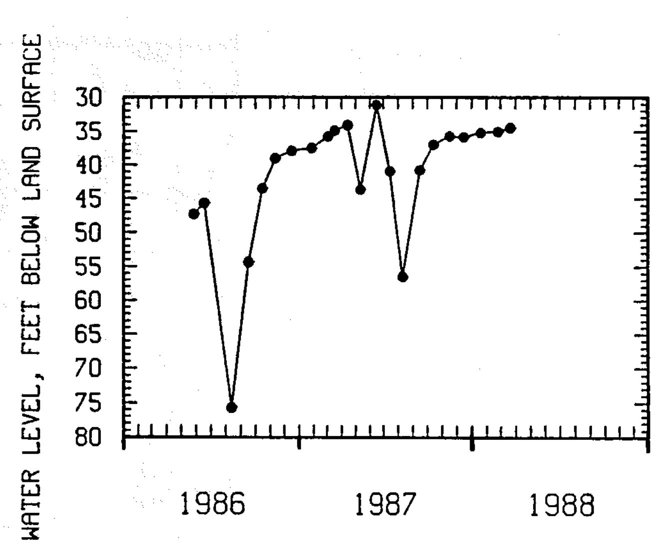


Figure 25. Hydrograph of recharge well at Stop 7 (Runkle and others, in review).

GEOLOGY BETWEEN STOP 7 AND STOP 8

Return to U.S. 62, go west 1 mi, turn south for 2 mi, and then west about 0.4 mi. Most of the route traverses soil-covered Dog Creek Shale, although the last 1 mi is upon soil-covered Van Vacter Member.

STOP 8

RECHARGE SINKHOLE, SOUTHEAST OF HOLLIS

Near center N½ sec. 18, T. 2 N., R. 25 W., Harmon County. This site is owned by Paul Horton, Box 745, Hollis, OK 73550 (phone: 405/688-9223 or -2392). He allows public access, but care must be taken to not damage the fence or equipment. Also, this site is hazardous: the sink is continuing to develop, and collapse of the ground can occur in the vicinity of the sink without warning, especially during or after rainfall.

Many natural sinkholes in the Hollis basin recharge the Blaine aquifer by funneling runoff from several tens of acres into the subsurface. This sinkhole has been modified by the SWSCD to receive some of the excess precipitation from about 100 mi^2 in the upstream drainage area of West Fork Sandy Creek.

The sinkhole is about 100 ft north of the creek (Fig. 26), and originally it was separated from the creek by a 15-ft-high natural embankment of dirt and blocks of gypsum. In the late 1970s the SWSCD used a bulldozer to cut a diversion ditch through the embankment to allow creek water to flow into the sinkhole. The original cut provided for a vertical drop of only 1 ft between the creek and the sinkhole, but the large flow of water has eroded the cut and the drop now is 5 ft (Fig. 27).

West Fork Sandy Creek was a perennial stream in the Hollis area prior to 1950; however, drought conditions from 1950–57 and development of irrigation wells (mainly in 1952–58) caused the water table to drop below stream level. Therefore, the creek is now an intermittent stream that has surface flow only during and after rain events. With low rainfall, virtually all creek water is diverted into the sinkhole; with high rainfall, the sinkhole is flooded and much of the excess water continues to flow downstream in the creek. High-water marks in the sinkhole show that water has risen 15 ft above the bottom of the depression at times within the past 15 years.

The sinkhole is developed mainly in the lower half of the Van Vacter Member (Fig. 27). The regional dip here is about 30 ft/mi to the northeast, but scattered blocks of dolomite and gypsum attest local dissolution and collapse. The swallow hole is in gypsum bed 2, and the water table probably is about 5 ft below the swallow hole much of the time. Most likely there is karst development down through gypsum bed 1, and the water is ultimately recharging mainly the Mangum Dolomite Bed. Since the diversion was constructed in the late 1970s, there has been a significant increase in the size of the sinkhole and the amount of land affected. Several other swallow holes are present in the sink, within 50 ft northeast of the main hole, and the dirt walls of the depression have caved in many times.

The land just north of the sinkhole is used for dryland farming, and it generally yields about 25 bushels of wheat per acre.

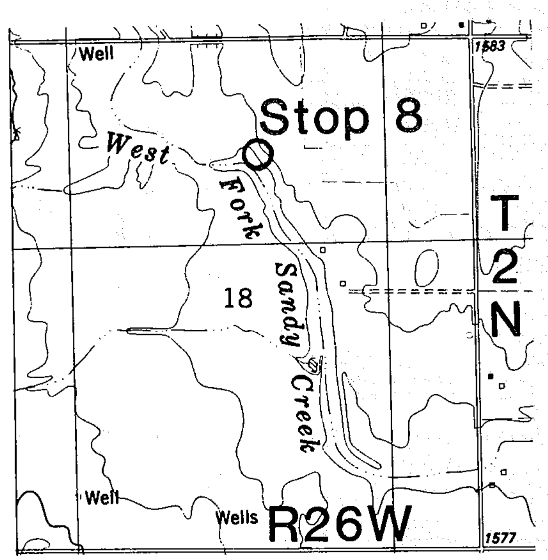


Figure 26. Location of recharge sinkhole at Stop 8, southeast of Hollis.

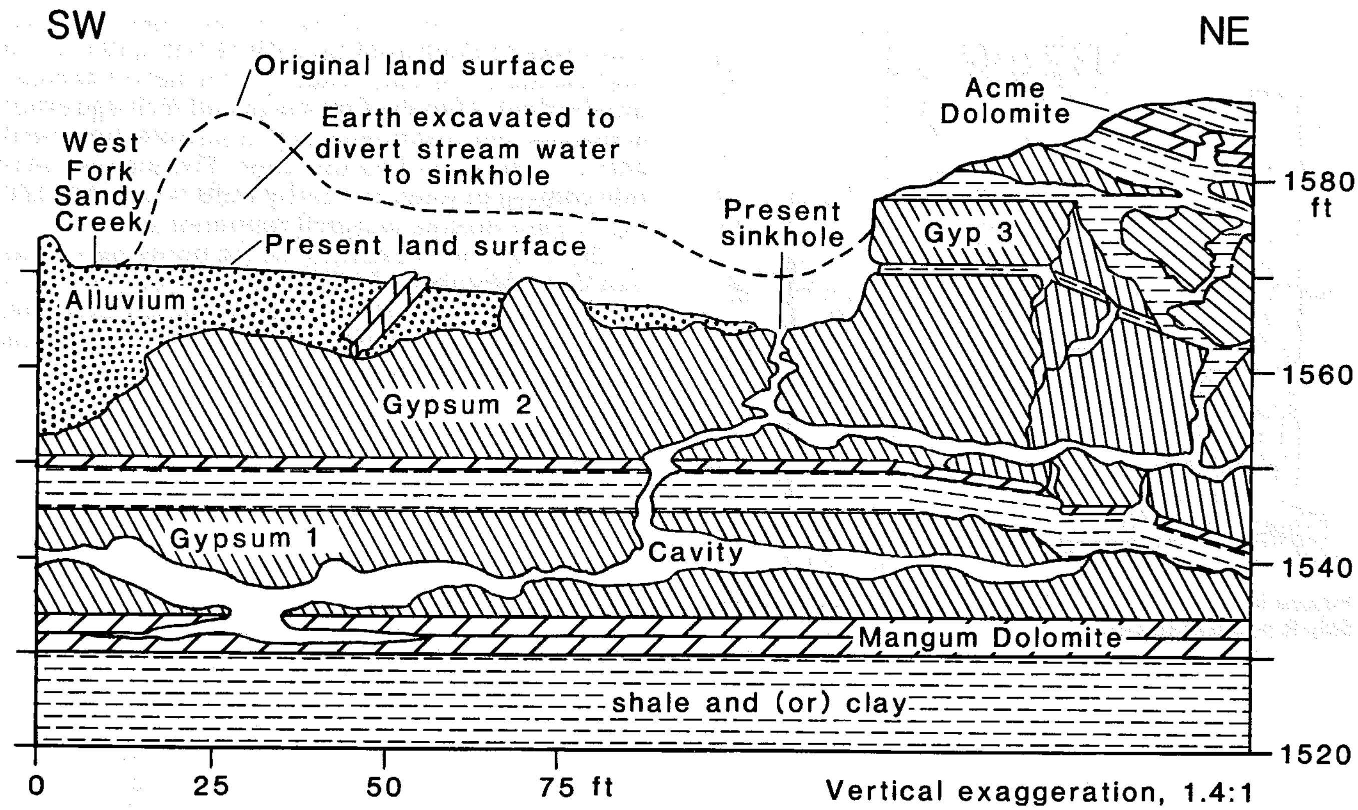


Figure 27. Schematic cross section showing diversion of excess stream flow from West Fork Sandy Creek into sinkhole at Stop 8. Subsurface geology and shape of cavity are speculative.

GEOLOGY BETWEEN STOP 8 AND STOP 9

The trip proceeds east about 5 mi to Oklahoma Highway 5, then south 4 mi, and back west 0.9 mi. Over most of the drive we traverse the lower Dog Creek Shale. For several miles on the highway south of Gould the depth to the Blaine is 50–100 ft, and the Blaine is not a well-developed aquifer; most of the sulfate beds are anhydrite, and the dolomites have limited development of porosity and permeability.

STOP 9

RECHARGE SINKHOLE AT BULLINGTON DAM, SOUTH OF GOULD

SW¼SW¼ sec. 36, T. 2 N., R. 25 W., Harmon County. Land is owned by Mrs. Martin Bullington, Rt. 1, Gould, OK 73544 (phone: 405/676-2835); she lives along Oklahoma Highway 5 on the east side of sec. 36.

Bullington Dam was constructed (about 1970) by SWSCD in order to capture stream water and divert it to a nearby natural sinkhole: it was one of the first artificial-recharge programs carried out by SWSCD (Fig. 28). The stream, which drains about 25 mi^2, originally flowed about 300 ft north of the sinkhole. A diversion ditch was excavated to carry base flow to the sinkhole on the south, and then a dam was built across the stream in order to store storm runoff and allow it to be funneled down the sinkhole. When the local water table is below the base of the sink, the lake that results from storm runoff will empty within 2–3 days; when the water table is up to the swallow hole, water can remain in the lake for several weeks.

The effect of this artificial-recharge system on local ground water was dramatic. Prior to construction,

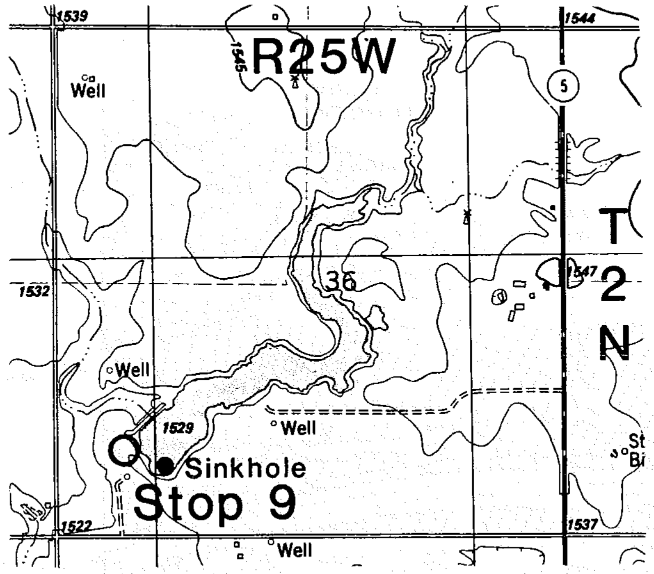


Figure 28. Location of Bullington dam and sinkhole at Stop 9, south of Gould.

water in local wells (such as the one adjacent to the dam) was of poor quality, with about 3,000–4,000 mg/L sodium chloride (Paul Horton, personal communication). After the first storm and recharge event at this site, the quality of local ground water improved and it could be used for irrigation. The sodium chloride content of water in nearby wells is now 200–500 mg/L (Paul Horton, personal communication).

The sinkhole is developed in the upper part of the Van Vacter Member, which dips to the east about 40 ft/mi. The precise beds at the swallow hole are unknown, but the Mangum Dolomite Bed is about 60 ft below ground level in the adjacent irrigation well.

GEOLOGY BETWEEN STOP 9 AND STOP 10

Return to Oklahoma Highway 5 and proceed south to the Oklahoma Highways 34 and 44 bridge over Red River. Our route takes us across lower Dog Creek and Van Vacter outcrops. In most areas the gypsum is extensively dissolved, because of proximity to Sandy Creek. The last 2 mi traverses windblown sands at the top of Red River terrace deposits.

STOP 10

GROUND-WATER DISCHARGE TO RED RIVER, SOUTHWEST OF ELDORADO

Vicinity of sec. 5, T. 2 S., R. 24 W., Jackson County. A good vantage point for this stop is in the parking lot of the Cow Barn Tavern, at the northeast end of the bridge across Red River.

A total of eight test wells were drilled along Red River (Figs. 29 and 30) in order to study the vertical movement of ground water between the various aquifers. Red River, when it flows, carries high-salinity water derived from natural-brine springs formed by dissolution of bedded salt in the Flowerpot, Blaine, and Dog Creek formations about 50 mi upstream. It was uncertain whether the Red River brine (13,000–23,000 mg/L TDS, and 6,000–12,000 mg/L chloride; Runkle and others, in review) might pass into the alluvium and then be discharged into the terrace and/or Blaine aquifers, thus contaminating the ground water in parts of the Hollis basin, or whether the various aquifers were discharging into Red River.

The eight wells were drilled in June and early July 1987, and water-level data (Fig. 30) were collected February 23, 1988. The well locations are marked by PVC pipe that projects about 3 ft above ground level. Wells 3 and 4 were drilled 10 ft apart and were completed to test alluvium and the Elm Fork Member, respectively. Wells 5, 6A, 6B, and 7 also were drilled 10 ft apart and were completed to test terrace deposits, Van Vacter Member, Elm Fork Member, and the Flowerpot Shale, respectively.

Water levels in the alluvium, terrace deposits, Van Vacter Member, and Elm Fork Member all are well above the base-flow level of Red River (Fig. 30), showing that all these significant aquifers are contributing

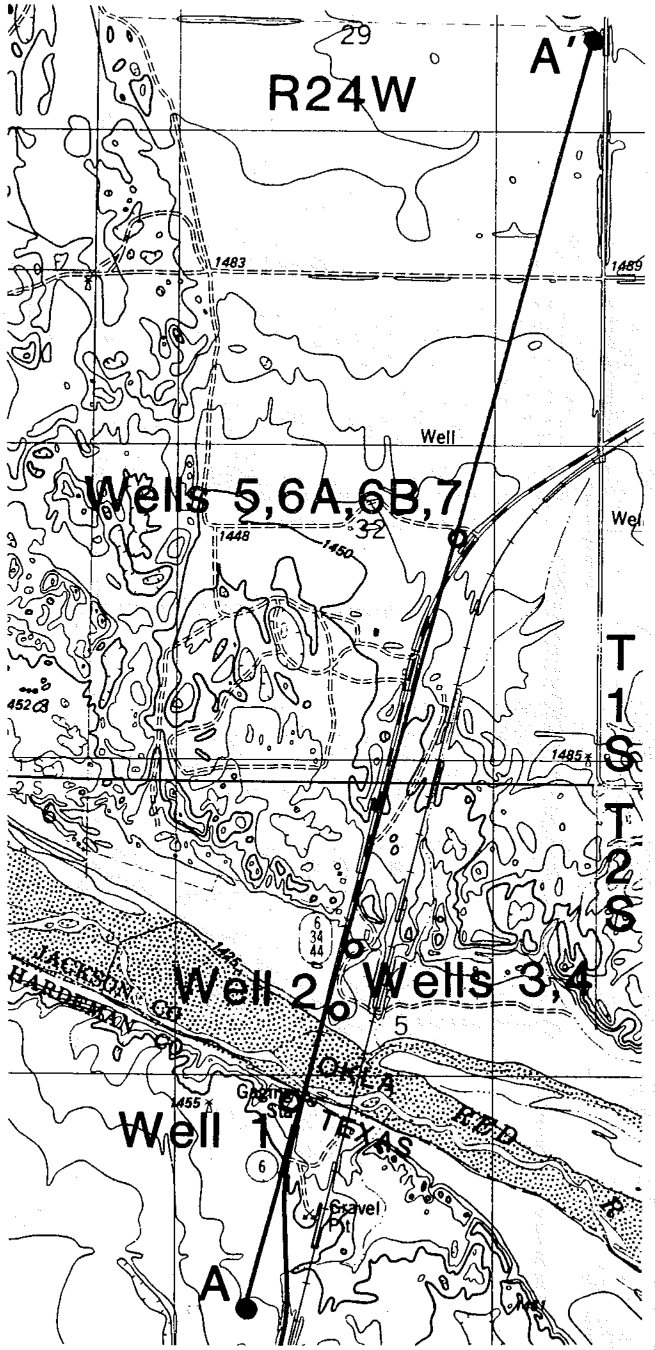


Figure 29. Location map for Stop 10, southwest of Eldorado. Cross section *A–A'* shown in Figure 30.

water to Red River in this area. Alluvium water levels (wells 2 and 3) are 3–4 ft above river level. Terrace-deposit water levels (wells 1 and 5) are 10–21 ft above river level: similar water levels in the terrace deposits and Van Vacter Member (wells 5 and 6A) confirm the ready hydraulic communication between these aquifers, where they are in contact. The 6-ft loss of head in the Elm Fork Member between wells 6B and 4 shows flow of water toward Red River and probable discharge into the base of the alluvium. The 15-ft difference in water levels between the Van Vacter and Elm Fork Members (wells 6A and 6B) indicates that there is no ready communication here between these two parts of the Blaine aquifer, and that the thick shales in the Elm Fork Member here are an effective aquitard. The water level for the Flowerpot Shale (well 7) is approximately the same as river level, and it appears to have little or no connection with any of the overlying aquifers.

Water quality in the alluvium is poor (20,000–30,000 mg/L TDS; Runkle and others, in review), as would be expected from the upstream emission of brine formed by dissolution of bedded salt. Terrace-deposit water in well 1 has only 800 mg/L TDS. Partly intermixed terrace-deposit and Van Vacter waters to the north (wells 5 and 6A) have 3,000–4,700 mg/L TDS; Van Vacter water contains mainly calcium sulfate, whereas the terrace-deposit water has significant amounts of both sodium chloride and calcium sulfate. Waters in the Elm Fork Member and Flowerpot Shale are of poor quality (35,000–234,000 mg/L TDS), with sodium chloride being the predominant constituent.

Test data at Stop 10 show that there is discharge from each of the principal aquifers (alluvium, terrace deposits, and Blaine aquifer) into Red River. Data also affirm the hydraulic connection between terrace deposits and the Van Vacter Member, where they are in contact, and the lack of significant hydraulic connection between the Van Vacter Member and either the Elm Fork Member or Flowerpot Shale.

This site is also significant as the first "watering hole" north of Hardeman County, a "dry" county in Texas. The area, formerly known as Trash Hill Taverns, had as many as five taverns operating in the 1940s to help slake the thirst of parched Texans. And now, with the field trip completed, we journey south into "dry" Texas toward Dallas, carrying our own supply of refreshments.

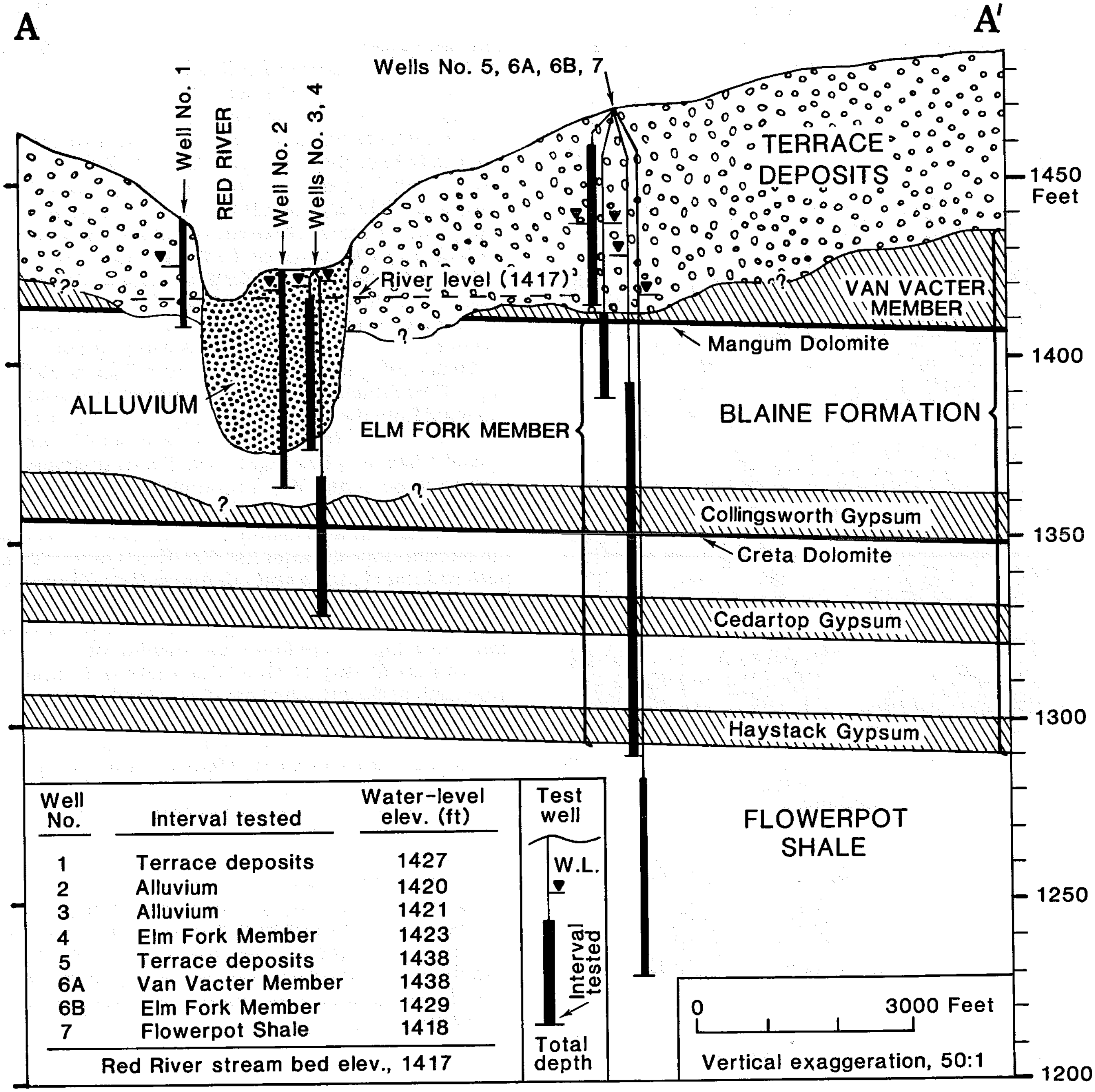


Figure 30. Schematic cross section showing eight test wells and water levels at Stop 10. Water-level data measured on February 23, 1988 (Runkle and others, in review).

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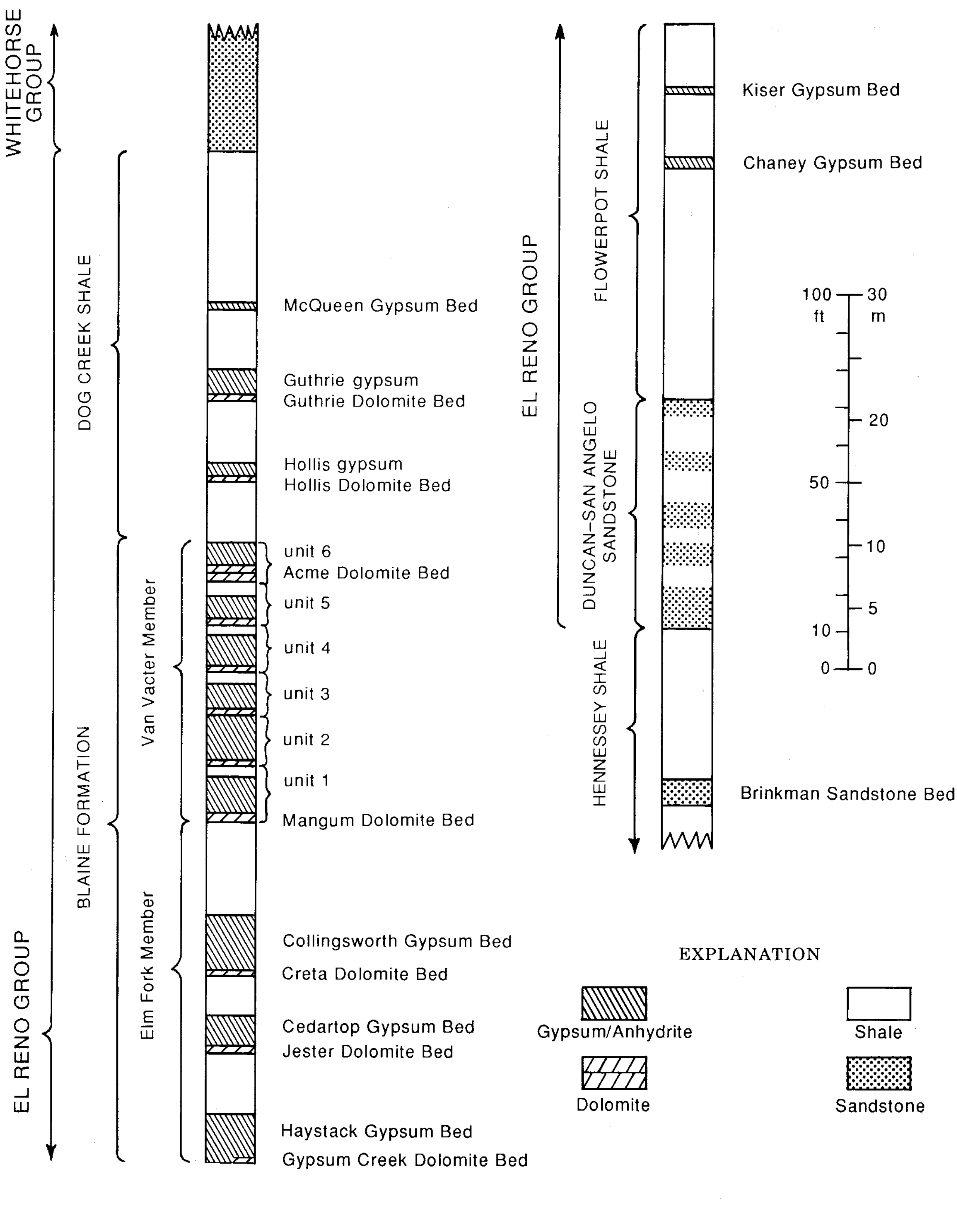
NOTES



Standard outcrop section of Permian Blaine Formation and associated strata in southwestern Oklahoma (from Johnson, 1990).