HYDROGEOLOGY AND RECHARGE OF A GYPSUM – DOLOMITE
KARST AQUIFER IN SOUTHWESTERN OKLAHOMA, U.S.A.

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Abstract

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 2,500 sq. km making up the Hollis Basin of southwestern Oklahoma, U.S.A. The Blaine aquifer typically is 50 to 65 m thick, and it consists of a sequence of laterally persistent gypsum, dolomite, and shale interbeds. Gypsum and dolomite beds have been partly dissolved by circulating ground waters, thus creating the karstic system comprising the aquifer. Karst features include sinkholes, caves, disappearing streams, springs, and underground water courses. Irrigation wells in the district are typically 15 to 100 m deep. They commonly yield 1,000 to 8,000 lpm of water containing about 1,500 to 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 have practiced artificial recharge by diverting excess runoff and surface drainage to natural sinkholes or to recharge wells drilled 15 to 30 m deep into cavernous gypsum - dolomite units.

Introduction

The Blaine Formation is a major aquifer in the Hollis Basin of southwestern Oklahoma (Fig. 1). Large quantities of irrigation water are produced at depths of began in the district in 1942, and it has grown to the extent that there are probably more than 800 irrigation wells in the 2,500 sq. km that make up the Hollis Basin. In addition, there are probably more than 2,000 additional water wells in the basin that tap the Blaine aquifer, but the yields are relatively small and these wells are devoted to household use or to providing water for livestock on pastureland. All water wells in the basin are extremely important to continued development of the region, because the relatively low annual precipitation (an average of about 60 cm) requires a heavy dependence upon ground water in the Blaine aquifer.

In earlier reports the gypsum - dolomite aquifer was referred to as the "Blaine - Dog Creek aquifer." Johnson (1967), however, showed that the massive gypsum and dolomite beds that yield almost all the water are restricted to the Blaine Formation; he proposed, therefore, that the aquifer be referred to as the "Blaine aquifer" (Johnson, in press).
Fig. 1 — Map of United States showing location (in black) of karstic Blaine aquifer in Hollis Basin area of southwestern Oklahoma.

Earlier reports on the hydrology of the Blaine aquifer in the Hollis Basin were done by Schoff (1948), Steele and Barclay (1965), Havens (1977), and Johnson (1983, and in press). Principal reports on the geology of the Blaine Formation and associated strata in the area include those of Johnson (1967, and in press), Johnson and Denison (1973), and Havens (1977).

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Geology of Aquifer

The Blaine aquifer is 50 to 65 m thick in southwestern Oklahoma and it consists of a sequence of laterally persistent interbeds of gypsum, dolomite, and shale (Fig. 2). Typically, each of the 9 white gypsum beds in the Blaine is underlain by a light - gray dolomite, and each is overlain by a shale that is mainly reddish brown but is gray at the top. Individual gypsum beds are 3 to 8 m thick, whereas the dolomite beds range from 0.1 to 1.5 m thick and the shale beds range from 0.3 to 8 m thick. Thus, the evaporite - redbed sequence
making up the Blaine is, in fact, a series of cyclic units, each of which begins (at the base) with a thin dolomite, overlain successively by a thick gypsum, a thin to thick red shale, and (at the top) a thin gray shale; above the gray shale is another dolomite, which begins the next cycle.

Gypsum and dolomite beds of the Blaine are widespread and laterally persistent throughout the Hollis Basin irrigation district (Fig. 2). They crop out in much of the region (Fig. 3A), and the top of the formation is nowhere more than 120 m below the land surface in the irrigation district (Fig. 4). 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).

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 ranges from about 55 to 60 m in the north and east to about 60 to 63 m in the southwest. The Van Vacter Member, comprising the upper half of the Blaine, consists of 6 thick gypsum-dolomite units, separated by thin shales. Gypsum beds in the Van Vacter are informally referred to as Gyp–1 through Gyp–6 (Fig. 2). The Elm Fork Member, the lower half of the Blaine, consists of only 3 thick gypsum-dolomite units, separated by thick shales (Fig. 5).

Individual gypsum beds of the Blaine typically range from 1.5 to 5 m thick in the east to about 3 to 8 m thick in the west; shale beds, on the other hand, typically thin westward across the basin. The thickest shale in the Blaine in near the middle of the formation, at the top of the Elm Fork Member (Fig. 2); this shale is about 15 m thick in the east and 10 m 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.

Overlying the Blaine Formation in the western part of the area is the Dog Creek Shale (Figs. 2, 3A, and 4), which consists of 30 to 55 m of red-brown shale with several gypsum-dolomite units in the lower 15 m of the formation (Fig. 2). The dolomite beds commonly are 0.1 to 0.7 thick and the gypsum beds are commonly 0.7 to 3 m thick.

Underlying the Blaine in all parts of the area is the Flowerpot Shale (Figs. 2 and 4). The Flowerpot consists mainly of red-brown shale with several interbeds of gray shale and gypsum that are each 0.1 to 1 m thick. Total thickness of the Flowerpot is about 45 to 60 m in the southeast and about 90 m throughout the rest of the irrigation district (Johnson, 1967). Interbedded with the red-brown shale in the northern part of the area are layers of rock salt (halite), referred to as the Flowerpot salt. The Flowerpot salt is generally 15 to 60 m thick, and its top is about 6 to 30 m below the base of the Blaine Formation (Fig. 4). 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 outside of these limits.

Where the Blaine Formation is deeply buried it normally contains some beds of anhydrite instead of gypsum (Fig. 4). Anhydrite, which is anhydrous calcium sulfate, is
Fig. 2 – Stratigraphy of Blaine Formation and associated strata in Hollis Basin area of southwestern Oklahoma (modified from Johnson, 1967). On left is standard stratigraphic column, and on right is a north-south stratigraphic cross section along the line of cross section X–X’ in Figs. 3 and 4.
Fig. 3 – Maps of Hollis Basin area of southwestern Oklahoma showing generalized surface geology (Map A, on left) and ground-water yields of the Blaine aquifer (Map B, on right). Map A modified from Havens (1977); map B modified from Johnson (in press). Cross sections X–X’ are shown on Fig. 4.
Fig. 4 — Cross sections showing Blaine aquifer in Hollis Basin area of southwestern Oklahoma (modified from Johnson, in press). Locations of cross sections are shown on Fig. 3.

Fig. 5 — Aerial view of the Blaine Formation (Elm Fork Member) showing three laterally persistent, white gypsum beds (each 3 to 6 m thick) separated by thick shales.
Fig. 6 — Sinkhole developed in gypsum beds of the Blaine Formation (Van Vacter Member). Sinkhole is about 15 m deep.

Fig. 7 — System of small caves developed in lower part of a gypsum bed in Blaine Formation. Horizontal dolomite bed, about 0.5 m thick, is at base of cavernous gypsum.
Fig. 8 — View inside a gypsum cave in the Blaine Formation. Cave is about 3 m high and 4 m wide at this place.

Fig. 9 — Cavity in Blaine gypsum bed is filled with sediment deposited from through flowing waters. Cavity is about 1.5 m high.
transformed to gypsum by hydration (combining with water) as a natural process in those areas where ground water circulates through shallow deposits. In general, hydration of anhydrite in the Hollis Basin has begun in sulfate beds less than about 60 m below the present surface, and is commonly completed in sulfate beds that are less than about 30 m below the surface.

Surface and near-surface Permian strata are essentially undisturbed structurally in most parts of the Hollis Basin. Strata dip gently towards the central part of the basin at a rate of about 2 to 4 m per km (0.1 to 0.2 degree). There are, however, about 12 faults and/or flexures in various parts of the basin where strata are offset about 3 to 30 m; these structures are not known to have any particular influence on groundwater circulation or distribution.

Karst Features

Gypsum and dolomite beds of the Blaine have been partly dissolved by circulating groundwater, thus creating the cavernous and karstic system comprising the aquifer. Karst features include sinkholes, caves, disappearing streams, springs, and underground water courses (Figs. 6, 7, and 8). These features are most common in areas where the Van Vacter Member is at or near the surface (Fig. 3A), because the Van Vacter consists predominantly of thick gypsum beds, and the low-permeability shale interbeds that it contains are quite thin. Note also that areas where the karstic Van Vacter Member crops out, or is in shallow subsurface beneath the Dog Creek Shale, are also the areas where the Blaine aquifer consistently yields large quantities of water to irrigation wells (Fig. 3B).

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 locally is due to early dissolution of fossils (mainly pelecypod shells), of cement around oolites and pellets, or of small nodules of gypsum; in other places it is intercrystalline porosity. 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 concentration of gypsum dissolution at the base of the gypsum beds. Thus the cavern system most commonly encountered is at and near the contact between gypsum and dolomite layers (Fig. 7). 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 a few cm to about 3 m. Enlargement of individual cavities and caverns occurs due to dissolution of the soluble rock and also due to abrasion of the rock by gravel, sand, and silt carried along by through-flowing waters. The sediment carried by groundwater is deposited locally in the underground caverns, and it partly or totally fills some of the openings (Fig. 9).

In some areas the underground caverns have become so wide that their roofs have collapsed to partly close the caverns (Fig. 10). The collapse structures and fractures thus enable vertical movement of the 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 apparently occupy their stratigraphic position.

Karst features are generally sparse to nonexistent in those areas where the Blaine is buried at depths in excess of 60 m below the surface. In these areas there has been little or no hydration of massive beds of anhydrite to gypsum, and the pathways for ground-water movement appear to be mainly in the dolomite beds.

Geohydrology

The Blaine aquifer has been yielding significant amounts of irrigation water for more than 40 years. High-yield wells produce water that is not very potable, but it is fully suitable for producing crops, and programs of artificial recharge are now adding to the supply of water that has been naturally recharging the aquifer each year.

Fig. 10 — Chaotic structure of outcropping strata in background where the roof has collapsed into caverns developed in underlying gypsum beds.
Groundwater Production—Discovery of the high-yielding Blaine aquifer occurred in 1942, when a 40 cm diameter hole drilled to a depth of 27.5 m was capable of being pumped at a yield of about 7,000 Lpm (liters per minute) (Schoff, 1948). Although the shallowest water was encountered at a depth of 15 m below the land surface in that well, the static water level after completion of the well was about 9 m below the surface. From the beginning, therefore, it was clear that the Blaine Formation contained water under artesian conditions. When a pump on the well was operated at a rate of 4,000 lpm, the water level in the well dropped 3 m almost instantaneously, and then it remained constant.

Areas with yields generally in excess of 400 lpm are shown on Fig. 3B. This map, generalized from Johnson (in press), was compiled from data provided mainly by well drillers who have worked in the Hollis Basin irrigation district for 10 to 40 years. Yields generally are the highest in the middle parts of the high-yield areas, where major streams are incised deeper into the upper part of the Blaine Formation and where solution channels are better developed.

Areas shown on Fig. 3B where the Blaine is expected to yield less than 400 Lpm are those areas where the aquifer generally is more than 60 m deep and is more likely to contain anhydrite rather than gypsum (Fig. 4). It is also true, however, that the land surface above some of these areas is very rough and rocky, and locally it has little or no soil cover; thus landowners in these areas have not attempted to complete high-yielding irrigation wells, and we must look at many of these areas as having not yet been adequately tested.

Water wells in the district normally are drilled to depths of 15 to 100 m using rotary or cable-tool drilling rigs. Most wells are completed in the shallowest gypsum-dolomite that is cavernous and yields 1,000 Lpm or more. Many of the wells yield more than 4,000 Lpm, and a few yield more than 8,000 Lpm. Several wells that yielded 8,000 Lpm (the pump's capacity) had less than 6 m of drawdown, and thus were capable of yielding at an even greater rate. The highest yield, about 10,000 Lpm, was reported for one well during a short period in 1952 (Steele and Barclay, 1965).

If an adequate supply of water is not encountered in the upper part of the Blaine aquifer, or if a shallow producing zone is pumped dry during the pumping season, wells are generally deepened another 10 to 30 m in search of cavernous zones in the deeper layers. The amount of deepening is dependent upon the local stratigraphy of the Blaine: drillers generally known the depth to each of the principal gypsum-dolomite units in the main parts of the district, but of course there is no advanced knowledge about whether a particular gypsum-dolomite unit will be cavernous and water bearing at a particular well location.

Few wells are flowing wells, and most are produced by use of gas-powered or electrically-powered turbine pumps. The main period of groundwater pumping is June, July, and August, the hot summer months when rainfall is at a minimum. Commonly the water table is lowered some 10 to 30 m during the summer season, thus requiring the aquifer.

Water Quality—Water from the Blaine aquifer is commonly of fair to poor quality, with the dissolved solids content generally ranging from 1,500 to 6,000 mg/l (Table 1).
Principal impurities in the water are calcium, sulfate, and carbonate, which reflects the chemical constituents dissolved from the host gypsum and dolomite beds of the aquifer.

Clearly the water is highly mineralized and generally is not suitable for human consumption, but the water is proven to be suitable for long-term use in irrigation in the region. There is no evidence of buildup of harmful salts in lands irrigated for many years with water from the Blaine aquifer.

Steele and Barclay (1965) point out that although water from the Blaine has a high total salinity, the salts generally have a low sodium content and thus the water can be used successfully on these loamy soils that characteristically have good drainage properties. The crops grown in this region, including cotton, wheat, alfalfa, sorghum, and grasses, appear to thrive on the low-sodium water, even though the total salinity of the water may be high.

Table 1. Chemical Analyses of Ground Water from Blaine Aquifer and Terrace Deposits in Hollis Basin Area of Southwestern Oklahoma (from Steele and Barclay, 1969). Data (except pH) are in mg/l.

<table>
<thead>
<tr>
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<th>Blaine Aquifer</th>
<th>Terrace Deposits</th>
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**a** Data from 33 water samples (5 samples from unused wells with chloride concentrations of 1,400 to 85,000 mg/l are not included).

**b** Data from 4 water samples.

Transmissibility and Storage — Estimations of transmissibility and storage were made by Steele and Barclay (1965) by using data on quantities of water pumped in selected parts of the irrigation district, and by measuring water-level declines due to this pumping. The usual methods of conducting aquifer tests were not used by Steele and Barclay because they felt that the physical and hydrologic properties of the Blaine aquifer differ so widely from those assumed in developing formulas for determining coefficients of transmissibility and storage.
The coefficients of transmissibility calculated by Steele and Barclay (1965) ranged from 1,490 to 5,715 m$^2$/day, and averaged 3,230 m$^2$/day. The coefficients of storage ranged from about 0.0004 to 0.03, and averaged about 0.016. Steele and Barclay (1965) believe that these values of transmissibility and storage are reasonable for the aquifer in areas where solution channels are best developed, but that they probably are several times too high for the aquifer throughout the entire irrigation area (the entire high-yield area on Fig. 3B), and much too high for the aquifer as a whole throughout the entire ground-water basin (both the high-yield and low-yield areas on Fig. 3B).

Steele and Barclay (1965) therefore felt that a reasonable value for the coefficient of storage for the part of the aquifer outside the irrigation area (in the low-yield area) would be one-third of that where the solution channels are best developed, or about 0.005, and that the average coefficient of storage for all parts of the aquifer affected by pumping for irrigation (in the high-yield area) would be half way between the two values, or about 0.01. They estimated that the total amount of water in storage in the aquifer is about $3.9 \times 10^8$ m$^3$.

Recharge — Owing to the widespread karstic conditions in outcrops of the Blaine and lower Dog Creek Formations in the Hollis Basin area, there is a great deal of natural recharge to the aquifer. This occurs mainly due to direct precipitation onto the karst areas, and also due to percolation from streams that flow across the karst areas. After it rains, water can be seen flowing into many of the depressions and sinkholes in the area. Some of the sinkholes receive all the runoff from several hectares, whereas other receive runoff from smaller areas. Many natural depressions, which may be sinks that have become 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 20 years. Most of the projects have consisted of diverting surface drainage to sinkholes, caves, or other natural openings into the ground that would allow direct flow of surface waters to the subsurface. However, in areas of few sinkholes, landowners have drilled recharge wells 15 to 30 m deep into underlying cavernous gypsum-dolomite units. These wells are generally located in or near local depressions or drainage ways, or the surface flow is diverted from drainage ways directly to the recharge wells (Fig. 11). More than 75 separate projects have been completed to assist in artificial recharge of the aquifer, and this has helped refill the aquifer to its maximum capacity in those years when precipitation is well above normal.
Fig. 11 — Surface facilities of recharge well drilled 20 m deep into cavernous gypsum-dolomite unit of Blaine aquifer. Well was drilled in small depression to collect local runoff. Screen prevents sticks, logs, and large pieces of trash from flowing down into the aquifer.
REFERENCES


Johnson, K. S., 1983. Maps showing principal ground-water resources and recharge areas in Oklahoma : Oklahoma State Department of Health, 2 sheets, scale 1 : 500,000.


