RAS 14530

Army Snyder Exh. # 5-D

[Originally Attached As EXHIBIT SNS #5 to Witness Snyder's pre-filed testimony]

U.S. NUCLEAR REGULATORY COMMISSION
In the Matter of US ARMY (JEFFERSON PROVING GROUND)
Docket No. 40-8838-MLA Official Exhibit No. ARMY EXH. # 5-D
OFFERED by: Applicant/Licensee Intervenor ______
NRC Staff Other ______
IDENTIFIED on ______ Witness/Panel ______
Action Taken: ADMITTED REJECTED WITHDRAWN
Reporter/Clerk ______

Fetter, C. W., 1988, "Applied Hydrogeology", Merrill Publishing Company, Columbus, Ohio.

DOCKETED
USNRC

October 25, 2007 (2:00pm)

OFFICE OF SECRETARY
RULEMAKINGS AND
ADJUDICATIONS STAFF

Docket No. 40-8838-ML

SECY-02

TEMPLATE = SECY-028

SECOND EDITION

APPLIED HYDROGEOLOGY

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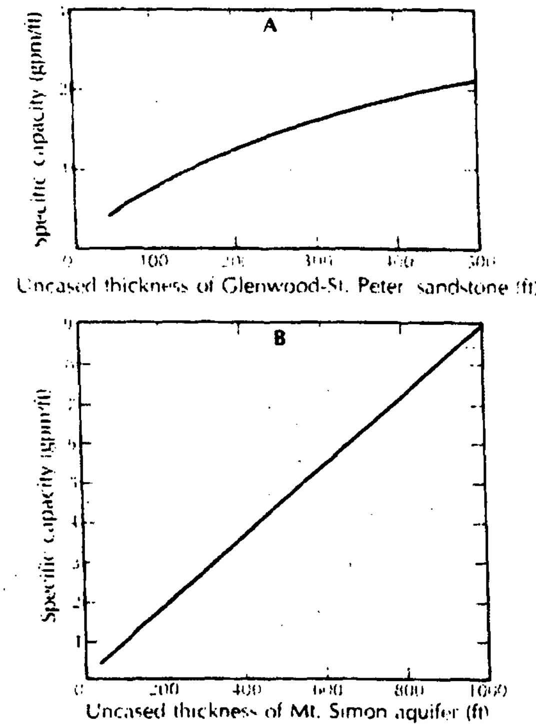

FIGURE 8.19 Relation between the specific capacity of a well (gallons per minute of yield per foot of drawdown) and the uncased thickness of the sandstone aquifer: **A.** Glenwood-St. Peter Sandstone. **B.** Mt. Simon Sandstone. Both of northern Illinois. Source: W. C. Walton and S. Csallany, Illinois State Water Survey Report of Investigation 43, 1962.

nation of explosives in the uncased hole. The shots are generally located opposite the most permeable zones of the sandstone. The loosened rock and sand is bailed from the well prior to the installation of the pump. The shooting process has two effects on the borehole: it enlarges the diameter of the well in the permeable zones and also breaks off the surface of the sandstone, which may have been clogged by fine material during drilling. Fractures near the well may be opened all the way to the borehole by shooting. Old wells may be rehabilitated by shooting if the yield has decreased owing to mineral deposition on the well face. Shooting has increased the specific capacities of sandstone wells in northern Illinois by an average of 22 to 38 percent, depending upon the formation (31).

8.3.4 CARBONATE ROCKS

The primary porosity of limestone and dolomite is variable. If the rock is clastic, the primary porosity can be high. Chemically precipitated rocks can have a very low porosity and permeability if they are crystalline. Bedding planes can be zones of high primary porosity and permeability.

Limestone and (to a much lesser extent) dolomite are soluble in water that is mildly acidic. In general, if the water is unsaturated with respect to calcite or dolomite, it will dissolve the mineral until it reaches about 99+ percent saturation with respect to calcite (32, 33). The rate of solution is linear with respect to increasing solute concentration until somewhere between 65 and 90 percent saturation, at which value the rate decreases dramatically. Figure 8.20 shows the general nature of the solution rate as a function of degree of saturation.

Massive chemically precipitated limestones can have very low primary porosity and permeability. Secondary permeability in carbonate aquifers is due to the solutional enlargement of bedding planes, fractures, and faults (34). The rate of solution is a function of the amount of ground water moving through the system and the degree of saturation (with respect to the particular carbonate rock present) but it is nearly independent of the velocity of flow (33). The width of the initial fracture is one of the factors controlling how long the flow path is until the water reaches 99+ percent saturation and dissolution ceases (33).

Initially, more ground water flows through the larger fractures and bedding planes, which have a greater hydraulic conductivity. These become enlarged with respect to lesser fractures; hence, even more water flows through them. Solution mechanisms of carbonate rocks favor the development of larger openings at the expense of smaller ones. Carbonate aquifers can be highly anisotropic and nonhomogeneous if water moves only through fractures and bed-

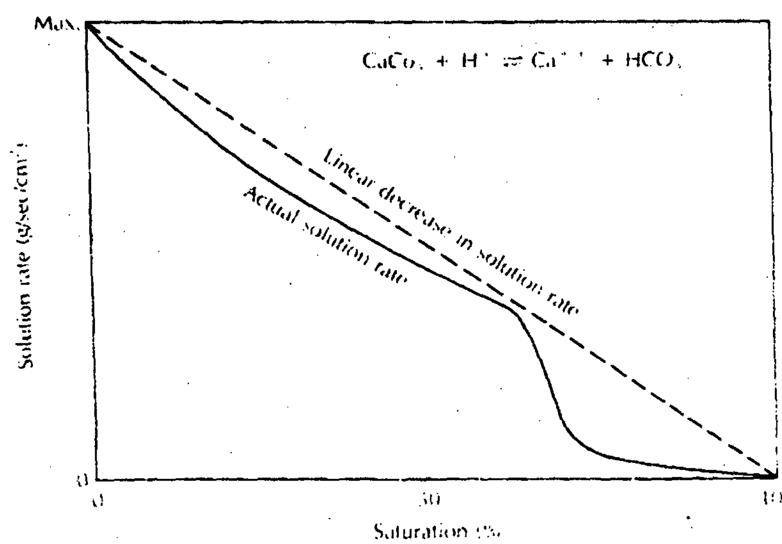

FIGURE 8.20 Solution rate vs. degree of saturation. Instead of decreasing linearly, the solution rate drops sharply to a low level at 65–90 percent saturation. Source: A. N. Palmer, *Journal of Geological Education*, 32 (1984):247–53.

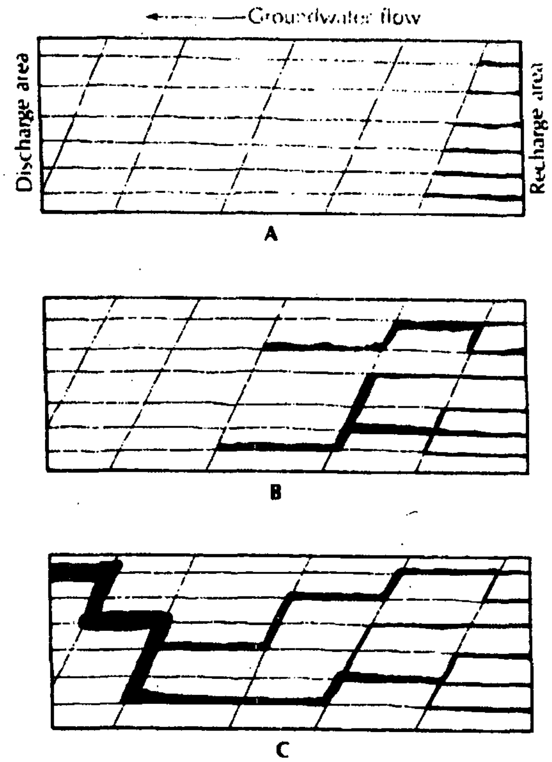

FIGURE 8.21 Growth of a carbonate aquifer drainage system starting in the recharge area and growing toward the discharge area. **A.** At first, most joints in the recharge area undergo solution enlargement. **B.** As the solution passages grow, they join and become fewer. **C.** Eventually, one outlet appears at the discharge zone.

ding planes that have been preferentially enlarged. Water entering the carbonate rock is typically unsaturated. As it flows through the aquifer, it approaches saturation, and dissolution slows and finally ceases. It has been shown experimentally that solution passages form from the recharge area to the discharge area and that, as they follow fracture patterns, many smaller solution openings join to form fewer but larger ones (35) (Figure 8.21). Eventually, many passages join to form one outlet. Greater ground-water movement—hence, solution—takes place along the intersection of two joints or a joint and a bedding plane. Ground water moving along a bedding plane tends to follow the strike of intersecting joints.

A second mode for the entry of unsaturated water into a carbonate aquifer occurs near valley bottoms. In **karst*** regions, flow in valleys with permanent

*Karst is a term applied to topography formed over limestone, dolomite, or gypsum: characterized by sinkholes, caverns, and lack of surface streams.

streams is usually discharged from carbonate aquifers recharged beneath highlands. Water tables in many karst areas are almost flat owing to the high hydraulic conductivity. Floodwaters from surface streams can enter the carbonate aquifers and reverse the normal flow. If the floodwaters are unsaturated with respect to the mineral in the aquifer, solution will occur (36).

Swallow holes, or shafts leading from surface streams, can carry surface water underground into caverns. Swallow holes can drain an entire stream or only a small portion of one.

Geochemical studies have shown that there are two types of ground water found in complex carbonate aquifer systems (37). The joints and bedding planes that are not enlarged by solution contain water that is saturated with respect to calcite (or dolomite). Because of the low hydraulic conductivity of these openings, this water mass moves slowly. Another mass of water, generally undersaturated, moves more rapidly through well-defined solution channels close to the water table. It is this second body that forms the passageways.

Cave systems can be formed above, at, or below the water table. They form when free-flowing water enlarges a fracture or bedding plane sufficiently for non-Darcian flow to occur. This can be above the water table if a surface stream enters the ground in the unsaturated zone **(vadose cave),** below the water table if the joint or bedding plane through which flow is occurring dips below the water table **(phreatic cave),** or at the water table itself **(water-table cave)** (34). The pattern of cave passages is controlled by the pattern and density of the joints and/or bedding planes in the carbonate rock (34). Figure 8.22 shows the influence of fissure density and orientation on cave formation. With widely spaced fissures, the cave can develop below the potentiometric surface because the fissure pattern is too coarse to allow the cave development to parallel the water table (Figures 8.22A and 8.22B). If the fissure density is great enough, cave development can occur along the water table (Figures 8.22C and 8.22D). Vertical shafts can form in the vadose zone by undersaturated infiltrating water trickling down the rock surface (38). Caves that are presently dry were formed below the water table when the regional water table was higher. The regional base level of a karst region is typically a large river. If the river is downcutting, the regional water table will be lowering. The result will be a series of dry caves at different elevations, each formed when the regional water table was at a different level.

Carbonate aquifers show a very wide range of hydrologic characteristics. There are, to be sure, a number of "underground rivers" where a surface stream disappears and flows through caves as open channel flow. At the other extreme, some carbonate aquifers behave almost like a homogeneous, isotropic porous medium. Most lie between these extremes.

Three conceptual models for carbonate aquifers have been proposed (36). **Diffuse-flow carbonate aquifers** have had little solutional activity directed toward opening large channels; these are to some extent homogeneous. **Free-flow carbonate aquifers** receive diffused recharge, but have well-developed solution channels along which most flow occurs. Ground-water flow in free-flow

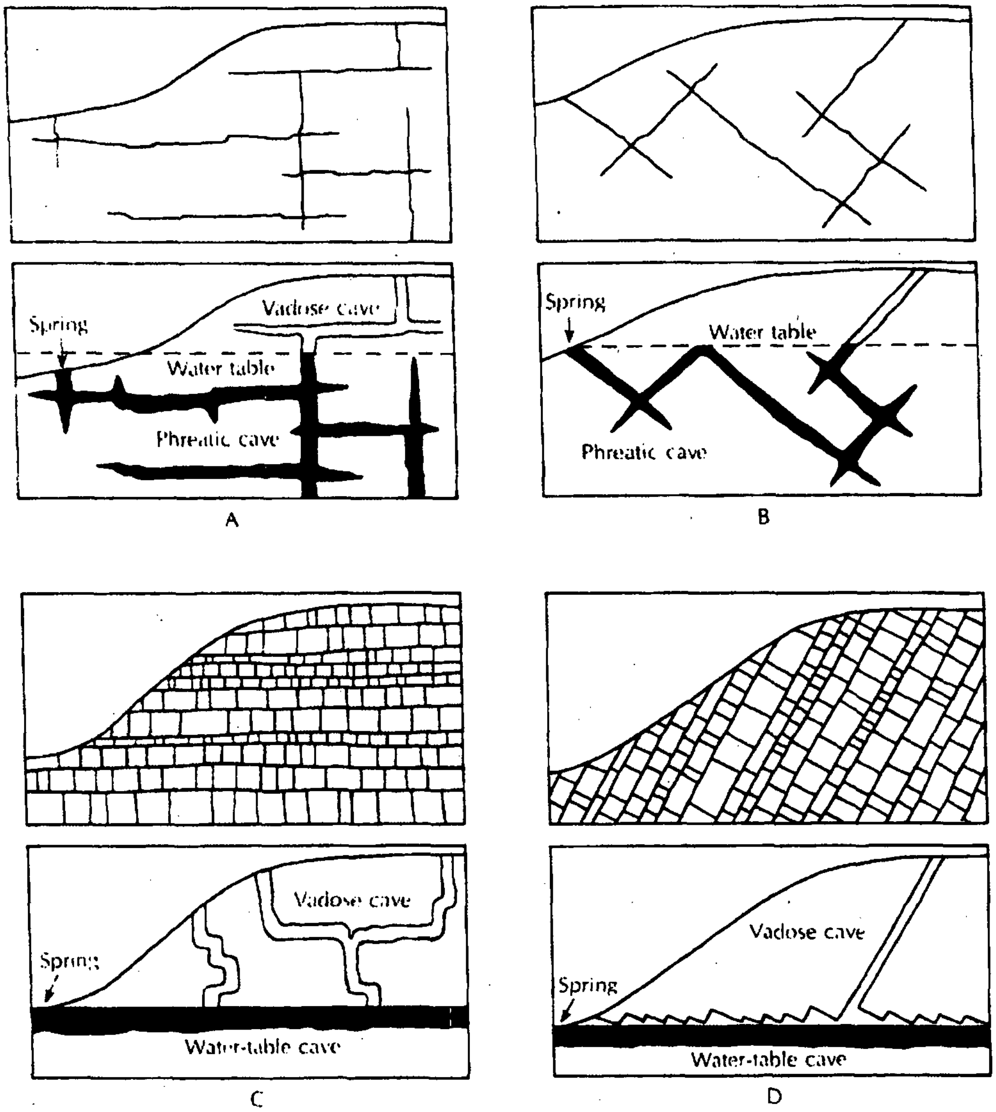

FIGURE 8.22 Effects of fissure density and orientation on the development of caverns. Source: Modified from D. C. Ford and R. O. Ewers, *Canadian Journal of Earth Science*, 15 (1978).

aquifers is controlled by the orientation of the bedding planes and fractures that determine the locations of solutional conduits, but not by any confining beds. **Confined-flow carbonate aquifers** have solution openings in the carbonate units, but low-permeability noncarbonate beds exert control over the direction of ground-water movement.

Diffuse-flow aquifers are typically found in dolomitic rocks or shaly limestones, neither of which is easily soluble. Water movement is along joints and bedding planes that have been only modestly affected by solution. Moving ground water is not concentrated in certain zones in the aquifer and, if caves are present, they are small and not interconnected. Discharge is likely to be through a number of small springs and seeps. The Silurian-age dolomite aquifer of the Door Peninsula of Wisconsin is an example. Well tests have shown that the horizontal flow of water is along seven different bedding planes in the dolomite. Vertical recharge is through fractures. The bedding-plane zones can be identified by borehole geophysical means (caliper logs) and correlated across several miles (39). Because water movement takes place along broad bedding planes, the yield of wells is fairly constant from place to place. Wells in vertical fractures have a higher yield, as they possess both vertical and horizontal conductivity. The water table in diffuse-flow aquifers is well defined and can rise to a substantial elevation above the regional base level.

Free-flow carbonate aquifers have substantial development of solution passages. Not only are many joints and bedding planes enlarged, but some have formed large conduits. While all of the openings are saturated, the vast majority of flow occurs in the large channels; the flow behaves hydraulically as pipe flow. Velocities are similar to those of surface streams. Flow is turbulent, and the stream may carry a sediment load—as suspended material, bedload, and suspended bedload. Water quality is similar to that of surface water, and the regional discharge may occur through a few large springs. Because of the rapid drainage, the water table is nearly flat, having only a small elevation above the regional base level. The very low hydraulic gradient indicates that diffused flow through the unenlarged joints and fractures is exceedingly slow. Recharge to the subterranean drainage system is rapid, as water drains quickly through the vadose zone. The water level in the open pipe network may rise rapidly in a recharge event (40). The spring discharge will also increase in response to the amount of recharge, so that the spring hydrograph may resemble the flood peak of a surface stream. The water levels in the open-pipe network will also fall rapidly as the water drains. Caving expeditions have been known to end tragically when a "dry" cave passage became filled with surcharged water during a rapid recharge event.

The depth of major solution openings below base level is probably less than 200 feet (60 meters), unless artesian flow conditions are present (36). However, in areas where the regional base level was formerly at a lower level (for example, where a buried bedrock valley is present), cavern development may have taken place graded to that base level. In the coastal aquifers of the southeastern United States, the drilling fluid may suddenly drain from a well being drilled when it is 300 or more feet (100 meters) deep. Cavernous zones found at these depths are well below the present water table. They formed when mean sea level and the regional water table were lower during the Pleistocene. The development of a sinkhole in Hernando County, Florida, was initiated by drilling in the Suwannee Limestone. The drilling fluid was lost several times, and the drill-bit would drop through small caverns. At 200 feet (62 meters), the drill broke into a cavern and, within ten minutes, a large depression had formed, with the drill rig sinking into the ground and the drillers narrowly escaping the same fate. The present-day water level is close to the land surface (41).

Sinking surface streams may also feed the pipe-flow network of a karst region. Lost River of southern Indiana is a typical headwater surface stream flowing across a thick clay layer formed as a weathering residuum on the St. Louis and St. Genevieve limestones. Where the clay thins, a karst landscape is present, and Lost River sinks beneath an abandoned surface channel. It appears as a large spring some miles away. There is no surface drainage in the karst region other than some ephemeral streams flowing into sinkholes (42).

If the carbonate rock beneath the uplands between regional drainage systems is capped by a clastic rock, karst landforms will not form. Dry caves in the uplands capped with clastics are less likely to have collapsed than caves in areas that are not capped. Recharge to the phreatic zone occurs through vertical shafts located at the edge of the caprock outcrop (38). These shafts, which may be as large as 30 feet (10 meters) in diameter and more than 300 feet (100 meters) deep, extend only to the water table. Water flows from them through horizontally oriented drains.

The central Kentucky karst region, including the Mammoth Cave area, is a capped carbonate aquifer system in some areas (43, 44). A cross section through the Mammoth Cave Plateau is shown in Figure 8.23. The plateau is capped by the Big Clifty Sandstone Member of the Golconda Formation, with cavern development in the underlying limestone formations, including the Girkin, St. Genevieve, and St. Louis limestones. Contact springs are found at the margins of the top of the plateau, as there is a thin shale layer at the top of the Girkin Formation. Recharge to the main carbonate rock aquifer takes place through vertical shafts formed in the plateau where the shale layer is absent. Karst drainage in the Pennyroyal Plain also contributes to the regional water table. Drainage is to the Green River through large springs, such as the River Styx outlet and Echo River outlet. These streams may also be seen underground where they flow in cave passages. Wells in the area draw water from the Big Clifty Sandstone, which yields enough for domestic supplies. There are some perched water bodies in the limestone, but the amount of water in storage is limited. A more permanent supply is reached if the well penetrates the regional water table, below the level of the Green River. The water level in these deep wells can rise 25 feet (8 meters) in a few hours during heavy rains and then can fall almost as fast (43).

If a carbonate rock is confined by strata of low hydraulic conductivity, those strata may control the rate and direction of ground-water flow (36). Such confined systems may have ground water flowing to great depths, with solution openings that are much deeper than those found in free-flow aquifers. Flow is not localized, and a greater density of joint solution takes place. The cavern formation of the Pahasapa Limestone of the Black Hills is apparently of this type, as water flows through the equivalent Madison Limestone aquifer eastward

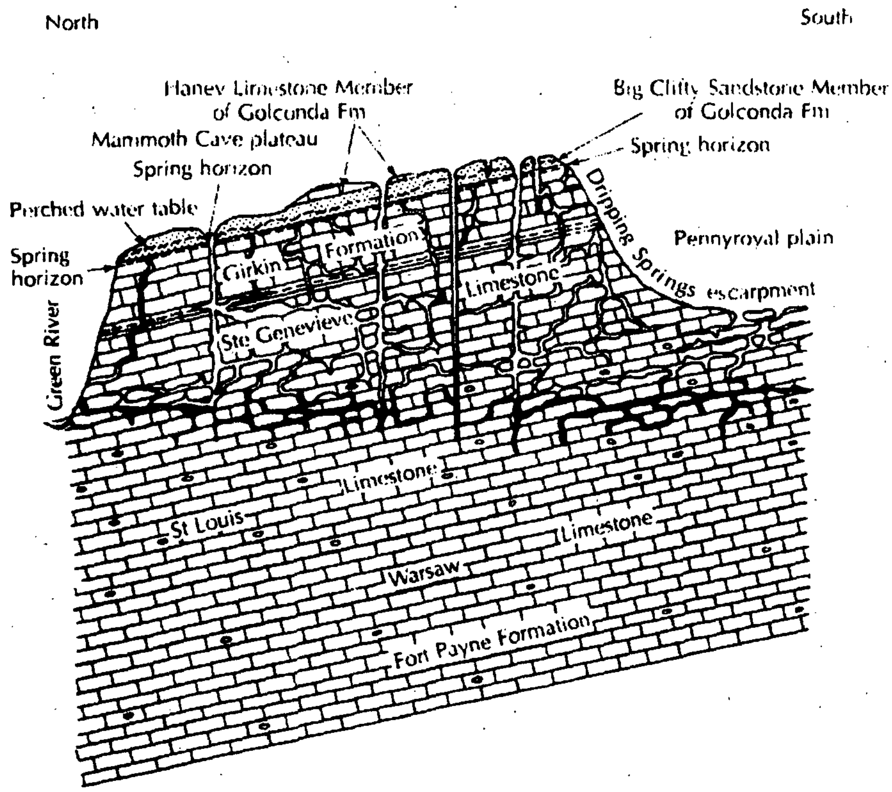

FIGURE 8.23 Diagrammatic cross section through the Mammoth Cave Plateau. Ground-water flow in the carbonate aquifer is from south to north. Source: R. F. Brown, U.S. Geological Survey Water-Supply Paper 1837, 1966.

from a recharge area at the eastern side of the Black Hills. It has been suggested that because of the low gradient (0.00022) of the potentiometric surface, the Madison Limestone is highly permeable, owing to solution openings, for at least 130 miles (200 kilometers) east of the Black Hills (45).

In the preceding discussion of karst hydrology, it was assumed that the various types of carbonate rock aquifers were isolated; however, this may not actually be the case. Highly soluble carbonate rock may be adjacent to a shaly carbonate unit with only slight solubility. The slope and position of the water table is a reliable indicator of the relative hydraulic conductivity of different carbonate rock units. In general, the water table will have a steeper gradient in rocks of lower hydraulic conductivity (46). This may be due to either a change in lithology or the degree of solution enlargement of joints. Figure 8.24 illustrates some conditions that might result in a change in the water-table gradient. In Part A, the hilltop is capped by sandstone, with a low-permeability shale between the sandstone and the underlying limestone. Only a limited amount of recharge oc-

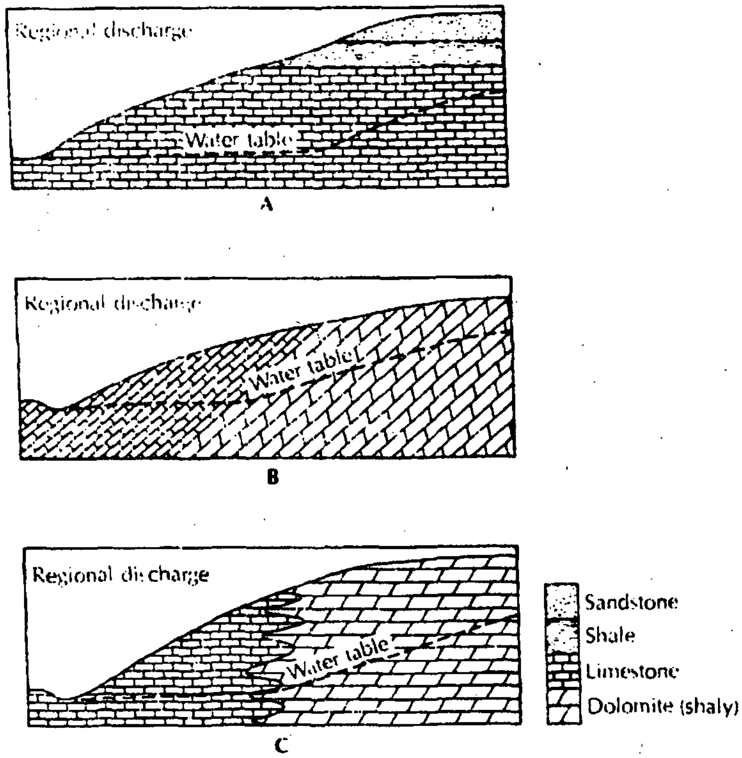

FIGURE 8.24 Geologic conditions resulting in a difference in hydraulic conductivity; hence, a difference in the water-table gradient.

curs through the shale. A spring horizon exists at the sandstone-shale contact, with small streams flowing across the shale outcrop area only to sink into the limestone terrane. The much greater amount of water circulating through the limestone in the area where the shale is absent has created a highly permeable, cavernous unit. Beneath the caprock, the limestone is less dissolved, owing to lower ground-water recharge. Because of the lower hydraulic conductivity, the water-table gradient is steeper. This type of situation has been reported in the central Kentucky karst, with the hydraulic gradient beneath the caprock near the drainage divide being much steeper (0.01) than that near the Green River (0.0005) (47). If an area has two rock units, one of which is more soluble, the more soluble rock may develop large solution passages and, hence, have greater conductivity. Parts B and C of Figure 8.24 illustrate two situations in which the upland is underlain by shaly dolomite and the lowland by cavernous limestone. The difference in rock solubility creates a change in hydraulic gradient.

The general concept of a water table in free-flowing karstic regions may be different from the model water table found in sandstones or sand and gravel aquifers. Because of the extremely high conductivity of some limestones, the water table can occur far beneath the land surface in mountains. Water can "perch" in solution depressions above the main water table. The level of free-flowing streams in caves is controlled by the regional water table, and the streams can have losing and gaining reaches, just as surface streams. Finally, because the solution of carbonate rock can be isolated along such features as fracture zones, the water table may be discontinuous. Wells drilled between fractures may not have any water in them, whereas nearby wells of the same depth may be in a fracture and therefore measure a water table (Figure 8.25).

The selection of well locations in carbonate terrane is one of the great challenges for the hydrogeologist. As the porosity and permeability may be localized, it is necessary to find the zones of high hydraulic conductivity. One of the most productive approaches to the task is the use of **fracture traces** (48, 49). Fracture traces (up to 1 mile, or 1.5 kilometers) and lineaments (1 to 100 miles, or 1.5 to 150 kilometers) are found in all types of geologic terrane. As they represent the surface expression of nearly vertical zones of fracture concentrations, they are often areas with hydraulic conductivity 10 to 1000 times that of adjacent rock. The fracture zones are from 6 to 65 feet (2 to 20 meters) wide and have surface expressions such as swales and sags in the land surface; vegetation differences, due to variations in soil moisture and depth to the water table; alignment of vegetation type; straight stream and valley segments; and alignment of sinkholes in karst. The surface features can reveal fracture traces covered by up to 300 feet (100 meters) of residual or transported soils. Fracture traces are found over carbonate rocks, siltstones, sandstones, and crystalline rocks.

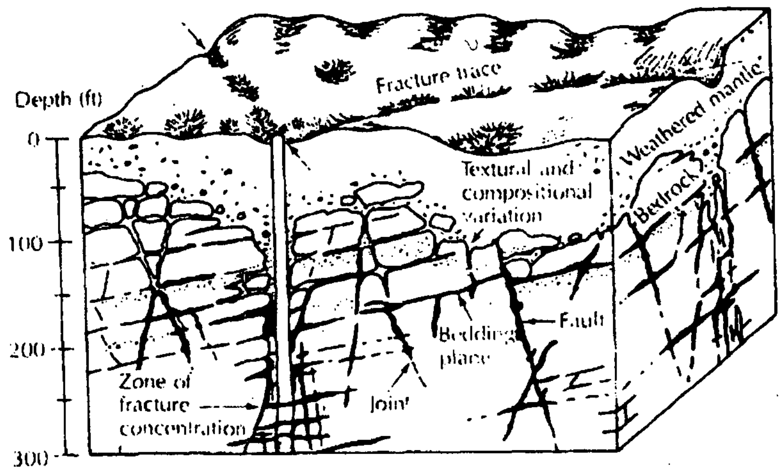

FIGURE 8.25 Concentration of ground water along zones of fracture concentrations in carbonate rock. Wells that do not intercept an enlarged fracture or a bedding plane may be dry, thus indicating a discontinuous water table. Source: L. H. Lattman and R. R. Parizek, *Journal of Hydrology*, 2 (1964):73–91. Used with permission of Elsevier Scientific Publishing Company, Amsterdam.

Because of differential solution in carbonate rocks, if a fracture zone has somewhat higher conductivity than that of the unfractured rock, flowing ground water will eventually create a much larger conductivity difference. Wells located in a fracture trace, or especially at the intersection of two fracture traces, have a statistically significant greater yield than wells not located on a fracture trace (50). The same relationship is apparently true for wells located on lineaments, as opposed to those not on lineaments (49).

In areas where topography is influenced by structure, valleys may form along fracture traces. In central Pennsylvania 60 to 90 percent of a valley may be underlain by fracture traces (50). Under such conditions, valley bottoms are good places to prospect for ground water. In the Valley and Ridge Province of Pennsylvania, the valleys are structural, and wells in valley bottoms are statistically more productive. In Illinois, just the reverse is true. For the shallow dolomite aquifer, the yields of wells in bedrock uplands are greater than those in bedrock valleys (51).

In central Pennsylvania, the yield of wells drilled into anticlines is greater than the yield of wells drilled into synclines. However, the proximity to a fault trace and rock type are more significant than structure in determining yield (50). In other karst areas, synclines have been noted as major water producers (52). The relation between structure and well yield is not clear; thus, local experience must be used as a guide.

One of the integral parts of carbonate terrane hydrogeology is the **regolith**—the layer of soil and weathered rock above bedrock. The regolith can be composed of weathering residuum of insoluble minerals remaining after solution of the carbonate minerals. This is typically reddish in color, owing to iron oxides, and contains a high proportion of clay minerals. The regolith may also include transported materials, such as glacial drift. If recharge must first pass through a low-permeability regolith, the rapid response of a carbonate aquifer will be reduced or eliminated. As the regolith slowly releases water from storage, spring discharge from areas overlain by a thick regolith will be more constant than if the regolith were absent. The regolith may also be a local aquifer. The weathering residuum of the Highland Rim area of Tennessee contains localized zones of chert, which can yield water in small amounts (53, 54).

8.3.5 COAL AND LIGNITE

Coal contains bedding planes cut by fractures that are termed **cleat.** Cleat is similar to joint sets in other rock. It is formed as a response to local or regional folding of the coal. There are typically two trends of cleat—normal to the bedding planes and cutting each other at about a 90-degree angle (55). Coal is often an aquifer and yields water from the cleat and bedding. The quality of water from coal aquifers is variable and sometimes can be poor. Such coals are typically anisotropic, with the maximum hydraulic conductivity oriented along the face cleat, which develops perpendicular to the axis of folding.

There is not a great deal of information on the hydraulic characteristics