H. H. Hobbs III

BULLETIN

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NATIONAL SPELEOLOGICAL SOCIETY

VOLUME 28 NUMBER 3

LIMESTONE HYDROLOGY

A SYMPOSIUM WITH DISCUSSION

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JULY 1966

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LIMESTONE HYDROLOGY—A SYMPOSIUM WITH DISCUSSION

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The Society serves as a central agency for the collection, preservation, and publication of information relating to speleology. It also seeks the preservation of the fauna, minerals, and natural beauty of caverns through proper conservation practices.

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Introduction to Limestone Hydrology

By George W. Moore

Abstract

At the symposium on limestone hydrology held in Berkeley in 1965, the following ideas were presented: The flow of ground water in limestone is largely independent of the flow in surface rivers and in surface drainage basins. Flow averages about eight meters per year in the Floridan aquifer. Solution channels initiate along joints and partings. Before the channels are large enough for water to move in them under normal hydraulic gradients, the water may move by pumping oscillations from earth tides and distant earthquakes. After channel flow is established, solution is most rapid in a narrow zone just below the water table where downward-percolating water mixes with ground water, because, although both waters may be saturated with respect to calcite, the mixture is usually undersaturated, owing to a nonlinear relation between calcite solubility and the partial pressure of carbon dioxide.

Knowledge of limestone hydrology up to the time of World War II was admirably summarized by A. C. Swinnerton in his "Hydrology of Limestone Terranes", published as a chapter of Meinzer's modern classic, *Hydrology*. After the hiatus caused by the war, interest in theoretical limestone hydrology was slow in rekindling. The impasse that resulted from the seemingly unreconcilable conflict between the Martel-Katzer theory that karst water behaves like water in pipes, and the Grund-Davis concept that it resembles flow in a semi-infinite permeable medium, led to further field investigations. By 1960, after the "Symposium on the Origin of Limestone Caves" was published in this *Bulletin*, the shape and spatial relations of the principal conduits for ground-water flow were better understood. Neither of the previous theories was entirely adequate. As implied by Swinnerton, and as documented by W. E. Davies and many other investigators, the principal channels for ground-water flow in unconfined limestone tend to form directly below the water table.

A new body of factual information has stimulated new interest in many aspects of limestone hydrology. This renewed interest has led to several recent symposia: one at Miami Beach, Florida, in November 1964; one at Dubrovnik, Yugoslavia, in October 1965; and this symposium at Berkeley, California, on December 29, 1965.

The meeting in Berkeley was held under the auspices of the American Association for the Advancement of Science. It was jointly sponsored by the Geological Society of America and the National Speleological Society. A. D. Johnson, Chairman of the San Francisco Bay Chapter of the National Speleological Society, was in charge of local arrangements, and the discussion was recorded by D. R. McClurg. Both the discussion and the individual papers presented at the symposium are published here.

In the symposium, Bedinger provided new evidence, through an electric-analog study, that in limestone most of the water flow and calcite solution occurs directly below the water table. Watson, citing data from the classic karst area of central Kentucky, illustrated that the flow is continuous from the input points to the discharge points—it is little affected by back-up water from surface rivers. Maxey and Mifflin showed that in Nevada the flow is indeed a regional phenomenon, entirely disregarding the surface topography and often passing through limestone ranges between undrained basins.

Back, Cherry, and Hanshaw defined the patterns of calcite undersaturation in Florida and illustrated how the various mineral phases present in the environment can affect the dissolving of limestone. They also showed by radiocarbon-dating methods that the ground-water velocity in the Floridan limestone aquifer averages about eight meters per year.

Ewers illustrated by laboratory and field studies that the first channels to form are braided solution tubes several millimeters in diameter. Davis focused his attention on the problem of joints and partings that initially are too narrow for pressures along natural hydraulic gradients to transmit enough water to initiate solution channels. He proposes that the water is forced through the fractures by differential movement of the two walls caused by earth tides and world-wide vibrations from seismic events. This permits the water in fine fractures to be replenished by undersaturated water from larger cracks nearer the surface.

The writer of this article pointed out another reason why most water flow and consequent solution in limestone occurs just below the water table, rather than at random depth within the water-saturated zone. A nonlinear relation exists between calcite solubility and the partial pressure of carbon dioxide. As a consequence of this relation, known as the Bogli effect, the mixing of any two waters, either saturated or undersaturated with respect to calcite, will always produce an undersaturated mixture capable of dissolving more calcite, so long as they initially differ in carbon dioxide partial pressure. Therefore, solution is most intense in the zone just below the water table, where downward-percolating water constantly mixes with slowly moving ground water.

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Initiation of Ground-Water Flow in Jointed Limestone

By Stanley N. Davis

Abstract

Dense unfractured limestone generally has a permeability of less than 10^{-12} cm^2 and will not transmit significant amounts of water under natural hydraulic gradients. In contrast, limestone fractured by faulting may have a permeability of more than 10^{-6} cm^2 which will allow a natural circulation of ground water with subsequent development of large solution openings. Many limestone caverns, however, are found in areas devoid of extensive faulting. These caverns are commonly localized along joints that must have been originally only hairline fractures. If this is true, then a basic question presents itself. Are the original fractures wide enough to transmit significant amounts of water? In-place measurements of joint widths in otherwise solid rock have not been reported in the literature, but general observations of exposed joints strongly suggest that little water can flow through newly formed joints. Measurements in Wool Hollow Cave, California, indicate that the rocks on opposite sides of some joints are in constant motion probably caused by periodic stresses such as those that are produced by tidal forces and seasonal temperature changes together with nonperiodic stresses such as those that are produced by world-wide seismic events. Differential movement along joints will both inhibit cementation and will increase the hydraulic conductivity of the joints by wearing away minor irregularities along the joint surfaces. Movement along the joints will also cause a reciprocating motion of the water in the joints, and it may even produce a net movement of water in one direction. Thus ground-water pumping produced by the movement of rocks may be an important driving force that supplements the driving force of the regional hydraulic gradient. Although partly speculative, this theory is supported by observations of tide-induced water-level fluctuations in deep water wells.

Introduction

The fact that joints control the geometry of many solution features in limestone has been so widely reported in the literature (for example, Bretz and Harris, 1961; Brod, 1964; Howard, 1963; Zeizel and others, 1962) that further comment on the subject may seem superfluous. The earliest stages of solution along joints, nevertheless, have received very little attention. Unfortunately, inquiry into this matter is hampered by an almost complete lack of basic data. Little is published, for example, on the permeability of dense crystalline limestone. Of the countless thousands of limestone cores tested by petroleum companies, almost all are of stratigraphic horizons that give some promise of being permeable enough for petroleum or gas production. Lack of information concerning the original width of joint openings is even more serious. Joints exposed in caves and artificial excavations do not represent the initial conditions in the rock because stress in the rock is released as natural or artificial openings are made. At the surface, conditions are even less favorable for the study of the initial width of joints. Not only has the stress in the rock been altered by erosion, but the joints near the surface commonly display signs of incipient solution.

In view of the acute lack of data, it is somewhat premature to attempt a discussion of the initiation of ground-water flow in limestone. Nevertheless, some profit can be drawn from a review of available information as well as a formulation of a hypothesis to explain observed features in the rock. Authors of most papers hope for some degree of longevity in their contributions; in contrast, this paper will have served its greatest purpose if it stimulates fieldwork so that it will be superseded as rapidly as possible.

Flow Velocities in Limestone

Continued solution of limestone with the resultant formation of void space is obviously not possible unless water moves freely through the rock. The amount of water coming into contact with the limestone is roughly proportional to the velocity of the water, provided the porosity is not changed radically. As a first approximation, it is assumed that the circulating ground water always has about the same capacity to dissolve limestone, so the formation of solution openings is related directly to the velocity of the water. Admittedly, this is an oversimplification, inasmuch as the pore geometry is modified slowly by solution, and the rate of solution is dependent on the chemical nature of the water. For the present, however, probable ground-water velocities will be taken as an indication of the likelihood of cave formation within a given limestone.

Dense crystalline limestone generally has a permeability of less than 10^{-12} cm^2 (approximately 10^{-4} darcy) and a porosity of less than 10 percent (Archie, 1952). Effective porosities are probably a few percent lower than the measured total porosities. Slightly porous limestone most commonly will have a permeability of less than 10^{-10} cm^2 (McComas, 1963). Regional hydraulic gradients that might be assumed to be present in dense limestone can be estimated by flow nets (fig. 1). Such gradients in all except mountainous regions and near hydrologic boundaries should range between about 0.01 and 0.0005. By the use of Darcy's law it is easily seen that reasonable combinations of effective porosity, permeability, and hydraulic gradient yield average flow velocities of less than 3 mm per year in dense limestone to as much as 3 m per year in slightly porous limestone. Even the higher velocity is barely significant in terms of formation of solution openings because of the extremely large amount of water that must percolate through the rock in order to dissolve a significant amount of limestone. As an example, water flowing through limestone at a rate of 3 m per year would take almost 100,000 years to increase the porosity of the rock by only one percent if the rock had an initial

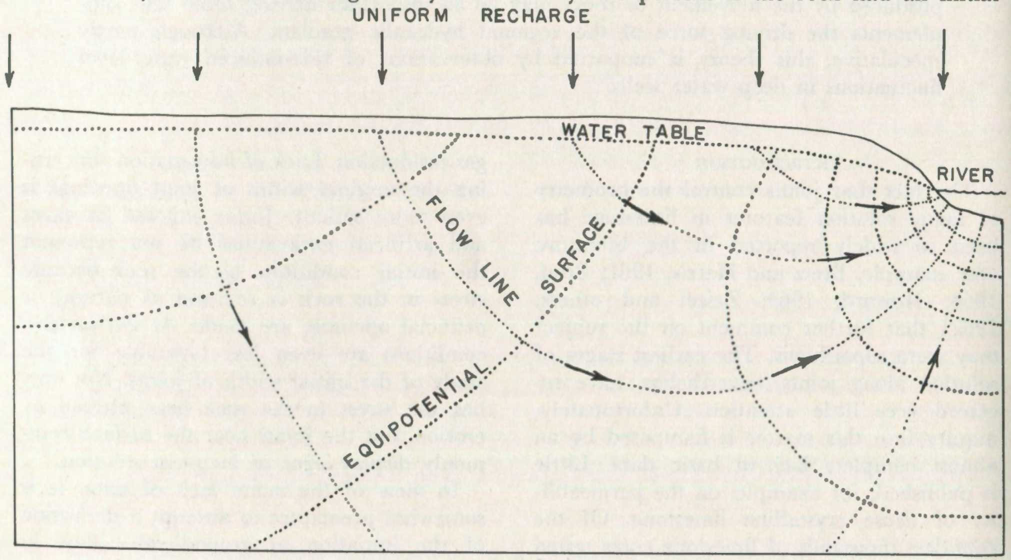

Figure 1.
Ground-water flow in a homogeneous and isotropic medium. Spacing between equipotential surfaces is inversely proportional to average water velocity.

effective porosity of 10 percent and if the water were to dissolve 1 ppm of $CaCO_3$ for every meter of travel. This rate of solution is probably liberal in comparison with all natural conditions except those near a soil-limestone interface. The unavoidable conclusion from the preceding remarks is that large solution openings in dense limestone are not caused by the percolation of water through the solid matrix of the limestone.

Limestone fractured by faulting will yield large quantities of water to wells, tunnels, and open excavations. Few reliable measures of permeabilities of such faulted zones are given in the literature, but values as large as 10^{-5} cm^2 would be reasonable. Flow will be turbulent in larger openings, so velocities estimated on the basis of Darcy's law can be misleading. Nevertheless, even rough estimates based on expressions for turbulent flow in pipes would suggest velocities measured in many kilometers per year. Little difficulty exists, therefore, in establishing hydraulic conditions prerequisite for extensive solution of faulted limestone.

The fact that caves and other solution features are developed in regions devoid of extensive faults and the fact that many of these solution features are linear has led to the almost universal conclusion that solution will follow preferentially joints that may be present. The critical question of the hydraulic characteristics of the joints, however, has not received much attention. Published references to joint widths are too general to be of much use. Typically, it is stated merely that openings are less than 1.0 mm wide. My own efforts to obtain better data have not been too fruitful. Information from test holes and water wells demonstrates that joints and other fractures in crystalline rocks tend to close with depth (Davis and Turk, 1964). The study was of noncarbonate rocks, but the same physical forces should act on thick massive limestone. Median well yields in carbonate rocks are generally so small that most joints penetrated must be less than 1.0 mm wide (see Palmquist and Hall, 1961, for records of typical wells in dense but jointed limestone).

I have made some attempts to make direct measurements of joint openings in fresh outcrops of rock along the beaches south of San Francisco, California. Rock types included sandstone, limestone, and granodiorite. Virtually all joints refused to receive the smallest probe used which was an 80 micron diameter wire. Study of the joints with a hand lense showed that the joints were at least ¼ the width of the wire probe. Although these measurements are not indicative of true conditions at depth, the fact that most joints close with depth would strongly suggest that joints in dense rock are generally less than 20 microns and probably less than 10 microns in width. Experiments with artificially fractured rocks tend to support the idea that joints can form which have less than 10 microns width. One piece of highly weathered marble, two pieces of schist, and two pieces of gneiss were broken in the laboratory. After they were broken they were forced together again by hand. Measurements of length taken before and after fracturing gave some indication of the actual width of the crack at the time the rock was being held together by moderate pressure of the hand. The limestone crumbled along the crack and gained 1 mm in length. The schist samples both gained about 0.1 mm. One gneiss sample gained about 0.2 mm; the other gained less than 0.01 mm, or 10 microns, which was the accuracy of the measurements. These crude experiments do not duplicate conditions of joint formation, but they do suggest that rocks can be traversed by incipient fractures much narrower than those measured because confining pressures much higher than those imposed by hand exist in the rocks at depth.

Flow velocities of ground water through joint openings cannot be predicted at present because of lack of information concerning joint widths and the hydraulic effects of the micro-texture of the rock surfaces. Nevertheless, a first approximation can be obtained of the flow velocities by assuming reasonable joint widths and by considering that the joints are bounded by smooth planes. In this case the following equation (DeWiest, 1965) can be used:

$$v = -\frac{b^2}{12}\frac{g\rho}{\mu}\frac{\partial h}{\partial x}$$

in which,

v is the average velocity in a fixed direction,
b is the spacing between smooth surfaces,
g is the acceleration of gravity,
ρ is the density of the fluid,
μ is the viscosity of the fluid, and
$\frac{\partial h}{\partial x}$ is the hydraulic gradient, a dimensionless ratio of lengths.

For water at about 20°C the equation reduces to,

$$v = -8.16 \times 10^3 (b^2) \frac{\partial h}{\partial x}$$

in which velocity is in cm/sec and spacing or width of the joint is in cm. One important aspect of the equation to note is the effect of width of the crack on velocity (fig. 2).

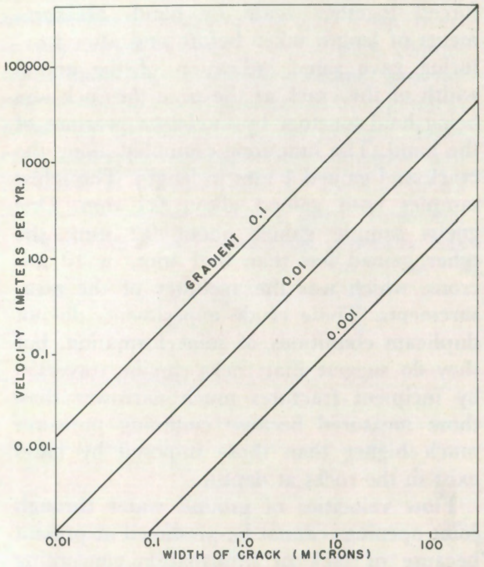

Figure 2.

Relation between water velocity, hydraulic gradient, and width of crack for flow between smooth parallel plates.

For example, water velocity will be 10^6 times as large in a crack 100 microns wide as it will be in a crack 0.1 micron wide. As a consequence, cracks appreciably smaller than about 10 microns will add very little if any to the bulk permeability of dense limestone.

Factors Related to Solution Rates in Jointed Limestone

Several possibilities exist that might explain the puzzling fact that solution cavities develop rapidly along joints that are not significantly more permeable than the enclosing rocks. First, the scant data available may be misleading. Many joints may indeed be virtually closed but they may be associated with other joints that are open and that eventually become enlarged by solution. Also, published values of permeability of dense limestone may be biased in favor of samples of higher permeability. If so, the contrast in permeability between joints and limestone may be highly significant, even with joints closed to less than 10 microns. Second, preferential solution along joints may be confined largely to shallow limestones, say less than 200 feet in depth. Joints in these shallow limestones may tend to open due to stress released by erosional unloading. Jewett and others (1965) have described an increase in the frequency of visible joints near the portals of a large underground limestone quarry near Kansas City, Kansas. Joints near the portals transmit water after surface precipitation; whereas joints in the interior of the mine do not transmit enough moisture to even leave visible mineral residues after being exposed for several years. Most of the mine is below the local water table which has been defined by exploratory drilling from the surface. Third, as mentioned elsewhere in this symposium, mineral surfaces under a higher stress undergo solution more rapidly than unstressed minerals. It is quite likely that stress is concentrated along some parts of joints so that these parts may be removed by solution more rapidly than the adjacent rock. Fourth, forces other than those supplied by the regional hydraulic gradient may serve to move water through joints. This last possibility will be explored further in the following section.

Differential Movement Along Joints

Earthquakes, temperature variations, earth tides, and loading of the ground surface by rain water as well as man-made objects will cause measurable strain in subsurface rocks (Benioff, 1959). Near the surface of

the ground the strain should be localized along joints, faults, and other discontinuities within rocks. Increased friction between rocks at depth produced by the weight of the overlying rocks will probably reduce the differential movement along tight joints so that it may be virtually eliminated at depths of a few thousand feet. Measurements across a joint in Wool Hollow Cave, California, indicated a semidiurnal displacement of 0.8 micron (Davis and Moore, 1965). The irregular nature of this movement is illustrated in figure 3 and also in the original paper. Although the effect of a distant seismic disturbance was detected, most of the movement was attributed to strains from earth tides. Whether or not all joints are subject to differential movement remains to be investigated. It is quite likely that some joints are more firmly locked together than others.

Hypothesis of Ground-Water Pumping

The initial development of solution openings in jointed limestone is tentatively attributed to ground-water pumping. Although the hypothesis is presented here in a simplified form, chemical and other factors must certainly operate together with the pumping in a very complex manner. Pumping effects are most likely important only during the initial stages of solution.

At the start the limestone is pictured as being jointed but with a very low overall permeability. The tendency for water to move through the rock under a natural hydraulic gradient will be present but the quantity of flow will be so small as to be negligible. Water from a stream, fault zone, or preexisting cavern is assumed to be in contact with the joints. Differential movement of rocks on either side of the joint will tend to smooth microscopic irregularities along the joint, inhibit cementation of the joint, and induce reciprocal motion of the small amount of water along the joint. The highly irregular motion of the joint will probably produce some unidirectional component to the water so that in some localities the slow movement of the water under the natural hydraulic gradient will be reinforced by the pumping.

As an example, a joint 10 microns wide

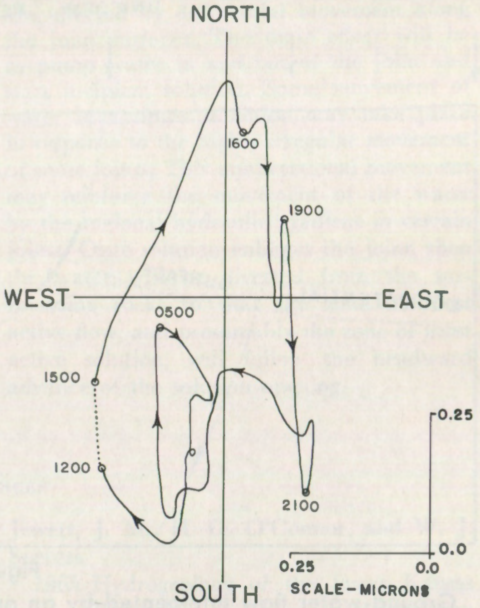

Figure 3.

Relative motion of limestone on either side of a joint during one day in Wool Hollow Cave, California (data from Davis and Moore, 1965).

extends 10 meters from a rock-water interface into the solid rock. If earth tides produce a 0.5 micron movement perpendicular to the joint, then average velocities of as much as 200 cm/day will be produced where the joint terminates at the rock-water boundary. Only a 1 percent unidirectional component of the flow will produce a net movement of 2 cm/day, or considerably more than would be produced by the normal regional hydraulic gradient to be expected under natural conditions. The large amount of energy required to force water through small cracks means that groundwater pumping cannot be effective in cracks much smaller than 1 micron nor in cracks that extend several hundred meters or more from the point where the water discharges from the crack into a larger body of water. This conclusion comes from a consideration of the equation for the flow of water between parallel plates. A joint 10 meters long having an opening only 1 micron wide and undergoing 0.5 micron movement will need a

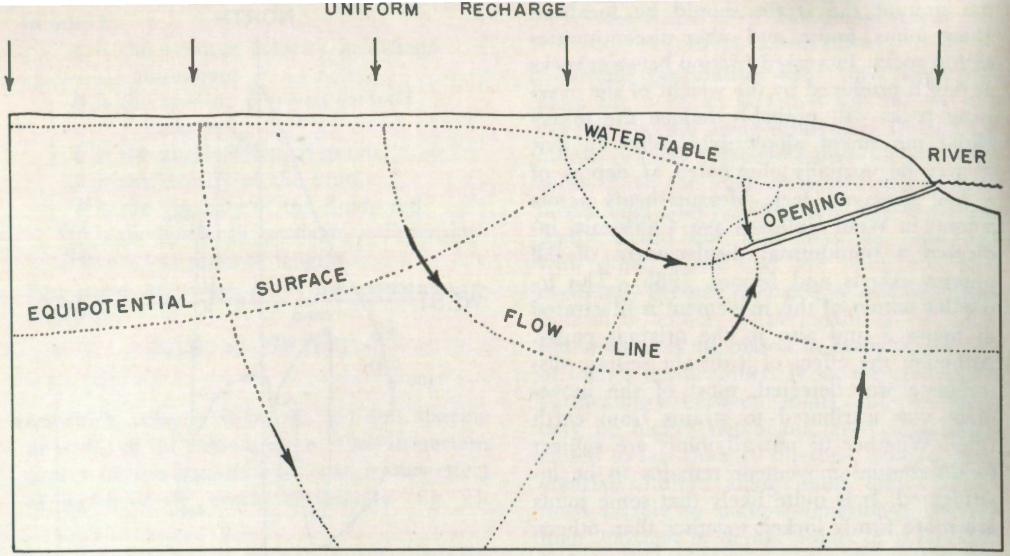

Figure 4.

Ground-water flow intercepted by an opening that is assumed to extend for some distance perpendicular to the plane of the figure. Comparison with figure 1 will indicate the great reduction of ground-water flow in the vicinity of the river.

hydraulic head of more than a thousand meters to force the water out. This is not possible because such high heads will be greater than the corresponding lithostatic pressure from the overlying rocks. What probably happens is that only a small fraction of the 0.5 micron of movement takes place and that most of the tidally induced stress is transmitted across the crack. In other words, the water-filled crack acts mechanically like part of the solid rock. As the crack width increases, the required head is decreased drastically. The initial example having a crack opening of 10 microns would need no more than 20 meters of hydraulic head to move the water.

Not all joints are visualized to act efficiently as pumps. Accidents of nature will favor only a few selected joints that react more vigorously to earth tides and other induced stresses. These joints will enlarge quite rapidly to widths of 100 microns or more. The relative importance of ground-water pumping, however, will decrease rapidly as the joints and cracks are opened so that they become efficient channels for normal ground-water flow. The enlarged joint subsequently tends to capture the surrounding ground-water flow so that large hydraulic gradients tend to be concentrated only along the leading edge of the enlarging joint (fig. 4). This water diversion will slow the flow in other lateral joints and will favor the continued growth of the most active joint. This can be seen in a comparison of figures 1 and 4. In the first figure the converging flowlines and crowded equipotential lines indicate the potential for rapid flow near the river. In contrast, the second figure shows the capture of most of the ground water by the joint and greatly diminished ground-water velocities near the river.

Although quite speculative, the foregoing hypothesis is not only supported by the data from Wool Hollow Cave but also by observations of earthquake- and tide-induced water-level fluctuations in wells. Several liters af water can be forced out of cracks and joints in dense rock to produce water-level fluctuations of as much as 1 meter in response to distant earthquakes (Vorhis, 1964; Parker and Stringfield, 1950). More frequent but less drastic fluctuations have been correlated with earth tides (Robinson, 1939). Twice-

daily fluctuations of as much as 7 cm suggest that more than a liter of water can be forced in and out of a well 15 cm in diameter and 120 meters deep (Stewart, 1964). Inasmuch as the total strain along joints and fractures must be very small, a large volume of fractured but otherwise dense rock must contribute to the ground-water pumping effect.

CONCLUSIONS

Review of the limited amount of data available suggests that jointed limestone under normal conditions is quite impervious to ground-water flow. Joints adjacent to existing bodies of surface or subsurface water are affected by differential movement along the joint surfaces. The main effect will be to pump water in and out of the joint and start incipient solution. Some movement of water in a single direction may take place in response to the highly irregular movement of some joints. This unidirectional movement may reinforce the movement of the water by the regional hydraulic gradient in certain joints. Once solution enlarges the joint, then the water will be diverted from the surrounding rocks so that the zone of most active flow, and presumably the zone of most active solution, will follow the headward advance of the solution opening.

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Discussion

Kenneth L. Edwards, *Chabot College, Hayward, California*: I would like to ask whether you have ever looked at the joints for recrystallization right along the joint surfaces.

Davis: The joint that we measured at Wool Hollow Cave extended through deposits of travertine on the cave wall, indicating that it actually was breaking through geologically Recent deposits. Some joints do tend to be cemented together, so probably they have no greater permeability than the solid rock, but I have just started work in this line of inquiry. I have been interested in water transmission through joints for some time, but, as I have already indicated, the collection of such data is rather frustrating.

Arthur L. Lange, *Stanford Research Institute, Menlo Park, California*: Where did you measure the thicknesses of the joints?

Davis: Along the coast south of San Francisco, where the ocean has actively eroded into the limestone, so the joints appear on very fresh surfaces.

Lange: Cenozoic limestone?

Davis: No, the Calera Limestone of Cretaceous age was the limestone I worked with.

Lange: You didn't measure any in the cave?

Davis: No, I became enthusiastic about this after we had measured the movements in the cave, and I haven't been back.

Lange: My impression is that joints are much more open in caves.

Davis: You would expect this because a cave is usually a zone of stress release, and you do get incipient movement, cave ins, and so forth. You would expect this of most cracks in tunnels as well as in caves.

Raymond E. deSaussure, *Lawrence Radiation Laboratory, Livermore, California*: Why do you exclude barometric and temperature changes in your pumping action?

Davis: I don't exclude them. I just didn't have time to discuss them. Any differential stresses that may be transmitted to the area along the joints would theoretically, at least, cause some movement. So several types of loading and strains are induced.

deSaussure: Have you made any attempt to work with the relative magnitudes of the effects?

Davis: No, our measurements are just getting started. We've actually only made the 22-hour series of measurements in Wool Hollow Cave. Now, related data from strain measurements that span several joints, which have been measured by seismologists, would seem to indicate that fairly large earthquakes have magnitudes of movement larger than those of earth tides. These may occur once every two or three months. Also, the sun coming up in the morning and hitting the rocks at the surface causes a thermal expansion right at the surface. This stress is transmitted to depth, but it is less than the stress produced by earth tides. So we do have some data of that type, but not data for a single joint.

Dan Sokol, *Hazleton Nuclear Science Corporation, Palo Alto, California*: Would you expect the solution cavities to develop more easily in confined aquifers where diurnal and semidiurnal fluctuations would be more likely to induce movement into and out of the cracks than under unconfined conditions?

Davis: This pumping mechanism that I postulate would probably be more effective in confined zones, not only due to the factors you mention but also due to the fact that under confined conditions much higher pressure differentials commonly exist, and very high gradients occur right at the margins of the solution openings.

Chemical Equilibrium Between the Water and Minerals of a Carbonate Aquifer*

By William Back, Rodney N. Cherry, and Bruce B. Hanshaw

Abstract

The departure of ground water from chemical equilibrium with several carbonate minerals is being studied in the major artesian aquifer of central Florida. The aquifer is composed predominantly of Tertiary age limestone with small amounts of dolomite and gypsum. Field measurements of pH, temperature, and concentration of bicarbonate are used in conjunction with standard laboratory analyses of water samples to calculate an ion-activity product for calcite, aragonite, and dolomite. The water in the major area of recharge is characterized by calcium bicarbonate type water, by low concentrations of dissolved solids, and by undersaturation with respect to calcite and dolomite. Water from a depth of several hundred feet is undersaturated within a large part of the recharge area. This indicates that solution of dolomite and limestone can occur at depths several hundred feet below the water table to form caverns and channels.

Concentration of total dissolved solids increases with length of flow path and duration of residence time in the aquifer. At a distance away from the recharge area, the water is supersaturated with respect to calcite and dolomite. Throughout most of the area studied, the water has not reached equilibrium with aragonite. Supersaturation with respect to dolomite downgradient from the recharge area is due to an increase in the magnesium content of the water probably derived from the solution of magnesian calcite. Radiocarbon ages indicate that within the area of little recharge, the ground-water velocities are about 8 meters per year. The combined study of mineral equilibrium and isotopic composition of the water provides an understanding of its geologic history from which the rate of solution in various parts of a hydrologic system may be determined.

This paper summarizes some of the work that we have been doing during the past few years that bears directly on the formation of caves and caverns. Owing to the greater solubility of carbonate minerals compared to silicate minerals, limestone generally is more permeable than other jointed rocks, such as granite or sandstone. In order to understand controls on secondary permeability and porosity distribution, including recementation of limestone aquifers, we need to identify the controls of solution and precipitation of carbonate minerals.

We are all aware that the solution of carbonate minerals is influenced by such things as the pH, temperature, effect of other ions in solution, and carbon dioxide content. The question is, how should the interrelations of these factors be studied to evaluate their relative effectiveness under natural conditions. One way to approach the problem is through the application of thermodynamics concepts to ground-water conditions. Our hypothesis has been that the law of mass action provides a thermodynamic model which can be tested in a carbonate aquifer (Back, 1963). To determine whether water is in equilibrium with a particular mineral, we need to know only the equilibrium constant K_{mineral}, obtained from physical-chemical laboratories; the other variables necessary to calculate an equivalent constant are obtained from the chemical analysis of water, which has commonly been

*Publication authorized by the Director, U. S. Geological Survey.

called the ion-activity product K_{iap}. For example, in the solution of calcite we have

$$CaCO_3 = Ca^{+2} + CO_3^{-2} \quad (1)$$

and the law of mass action gives us

$$K_{calcite} = \frac{\alpha Ca^{+2} \bullet \alpha CO_3^{-2}}{\alpha CaCO_3} \quad (2)$$

here α = the activity in molal units, which is the thermodynamic concentration of each component and is equal to the molality multiplied by a correction factor called the activity coefficient. The activity coefficient γ is obtained from the Debye-Hückel equation

$$\log \gamma_i = \frac{-Az_i^2 \ \Gamma^{1/2}}{1 + Ba_i \ \Gamma^{1/2}}$$

where A and B are temperature-dependent constants, z_i is the charge on the ion, a_i represents the effective diameter of the ion in solution, and Γ is the ionic strength of the solution. $\Gamma = \frac{1}{2}\Sigma m_i z_i^2$

where m_i is the molality of each ion.

By definition, the activity of pure water or of a solid phase is equal to 1.

In a like manner, the ion-activity product is

$$K_{iap} = \alpha Ca^{+2} \bullet \alpha CO_3^{-2} \quad (3)$$

The activities used in equation 3 are calculated from the chemical analysis of the water. However, because the dominant carbonate species in solution in water is bicarbonate rather than carbonate ion, we use the reaction

$$HCO_3^- = H^+ + CO_3^{-2} \quad (4)$$

The equilibrium constant for equation is given by

$$K_{HCO_3} = \frac{\alpha H^+ \bullet \alpha CO_3^{-2}}{\alpha HCO_3^-} \quad (5)$$

From equations 3 and 5, we obtain

$$\alpha CO_3^{-2} = \frac{K_{HCO_3^-} \bullet \alpha HCO_3^-}{\alpha H^+} \quad (6)$$

where $K_{HCO_3^-}$ is a thermodynamic constant at a given temperature, the activity of the bicarbonate ion is calculated from the chemical analysis of the water, and the activity of the hydrogen ion is the pH of the water. The K_{iap}, if compared with the equilibrium constant, determines departure from equilibrium of the water with respect to calcite. If the ratio K_{iap} to $K_{calcite}$ is equal to 1 (100 percent), the water is saturated with respect to calcite; a ratio less than 1 (less than 100 percent), denotes undersaturation; and if the ratio is greater than 1, the water is supersaturated. From the form of equation

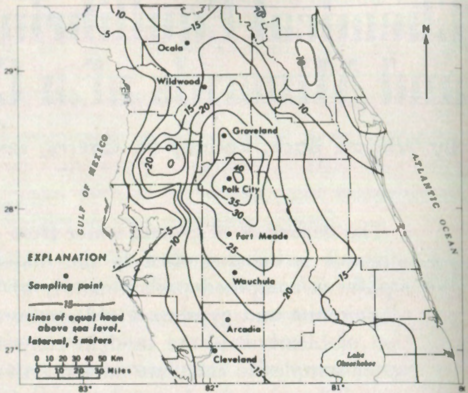

Figure 1.
Piezometric map of the principal artesian aquifer in central Florida showing location of sampling points shown on cross sections (after Stringfield, 1936).

3, it is seen that the K_{iap} for calcite and aragonite are the same.

Because all systems in nature tend toward equilibrium, we know that where water is undersaturated with respect to a mineral, this mineral is being dissolved. However, because of kinetic effects, we cannot categorically say that, where the water is supersaturated with respect to a mineral, the mineral is necessarily being precipitated.

By using ground water in Florida as an example, we can see the application of this approach to a hydrologic system in a carbonate aquifer (Back, 1963). The ground-water flow pattern of the system under study is controlled by the piezometric high shown in figure 1. Water tends to move perpendicular to, and toward, piezometric contours indicating progressively lower elevations. Although water is recharged in the highest part of the area underlain by the piezometric surface, greater volumes of water are recharged to the north in the vicinity between Wildwood and Groveland. The wells sampled range in depth from about 100 to 1500 feet and are completed in the Ocala Limestone and other formations of Eocene age which comprise the principal artesian aquifer (Floridan).

The total-dissolved-solids (TDS) content of the water (fig. 2) increases north and south away from the area of high potential and is lowest in the Polk City-Wildwood area, with a TDS content less than 200 milligrams per

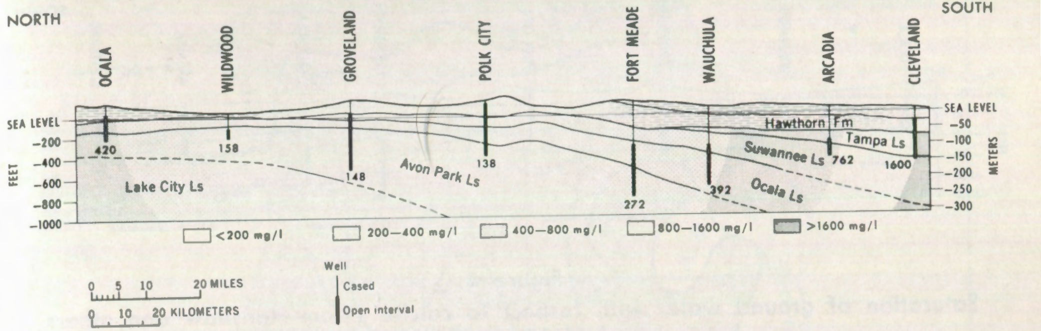

Figure 2.
Total-dissolved-solids content.

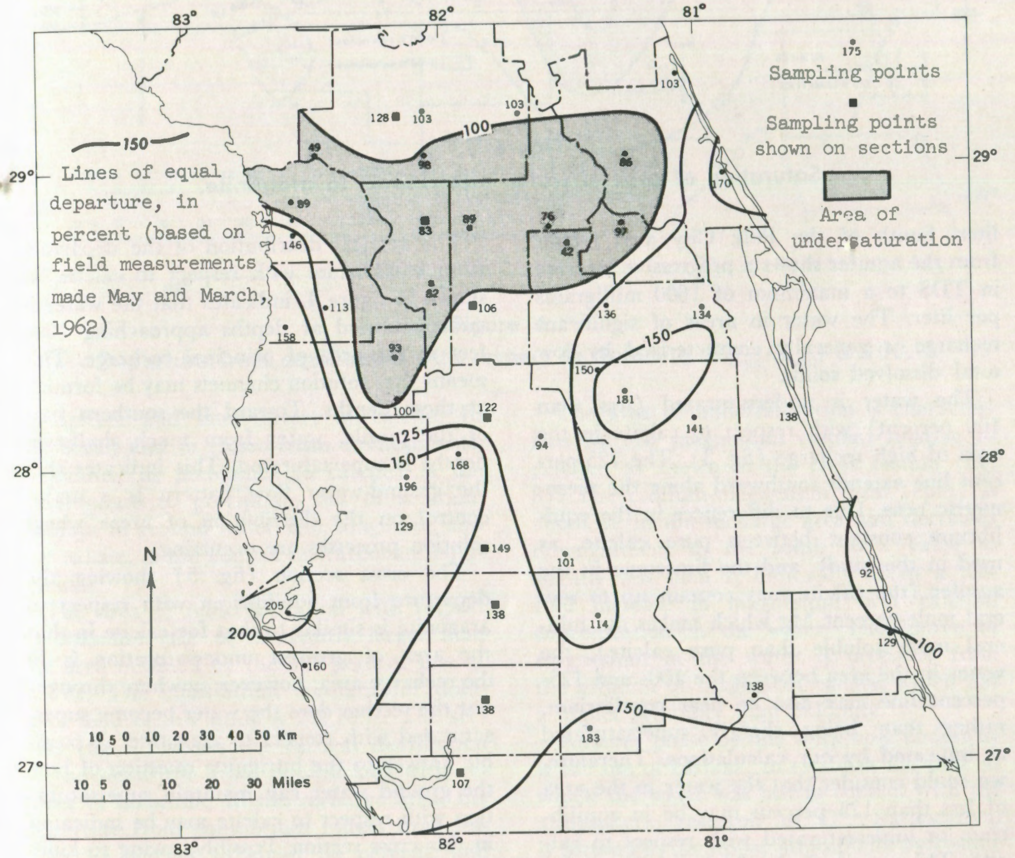

Figure 3.
Departure from equilibrium with respect to calcite ($K_{calcite}/K_{iap}$) x 100.

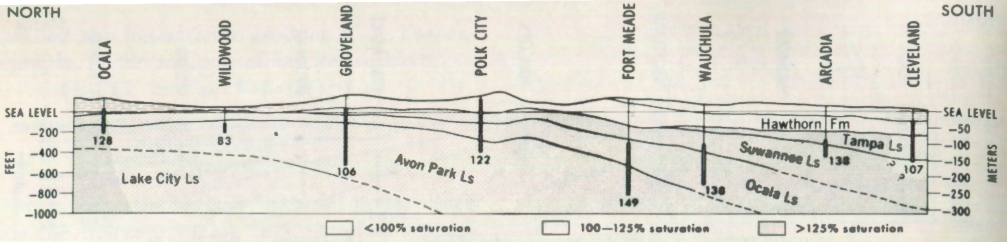

Figure 4.

Saturation of ground water with respect to calcite (from Hanshaw and others, 1966).

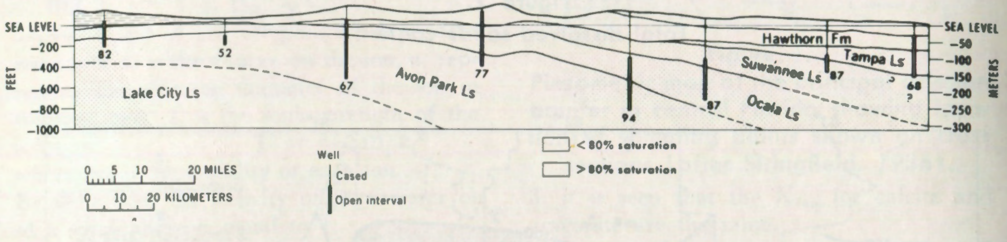

Figure 5.

Saturation of ground water with respect to aragonite.

liter. South of the Polk City area, water from the aquifer shows a progressive increase in TDS to a maximum of 1600 milligrams per liter. The water in areas of significant recharge is generally characterized by low total dissolved solids.

The water is undersaturated (less than 100 percent) with respect to calcite in the area of high recharge (fig. 3). The 125-percent line extends southward along the piezometric nose. Due to differences in the equilibrium constant between pure calcite, as used in this study, and the limestone in the aquifer (the calcite may contain up to several mole-percent Mg which makes the mineral more soluble than pure calcite), the water in the area between the 100- and 125-percent line may also be near equilibrium, rather than being slightly supersaturated as indicated by our calculations. Therefore, we could consider that the water in the area of less than 125 percent may be in equilibrium or undersaturated with respect to calcite. The water off the flanks of the piezometric high to the east, west, and south is supersaturated, with the highest values approaching 200 percent.

The vertical distribution of the departure from equilibrium with respect to calcite, as shown in figure 4, indicates that the water is undersaturated at depths approaching 1000 feet in the area of principal recharge. This means that solution channels may be forming at these depths. Toward the southern part of the profile, water from much shallower depths is supersaturated. This indicates that the ground-water flow pattern is a major control on the distribution of areas where solution processes are occurring.

The cross section (fig. 5) showing the departure from equilibrium with respect to aragonite is similar to that for calcite in that the area of greatest undersaturation is in the recharge area; however, nowhere throughout the section does the water become supersaturated with respect to aragonite. A possible answer to the intriguing question of how the ground water can maintain supersaturation with respect to calcite may be indicated in this cross section. Possibly, owing to kinetic effects, the water can maintain supersaturation with respect to calcite until saturation with respect to aragonite is reached. Once equilibrium is obtained, aragonite may

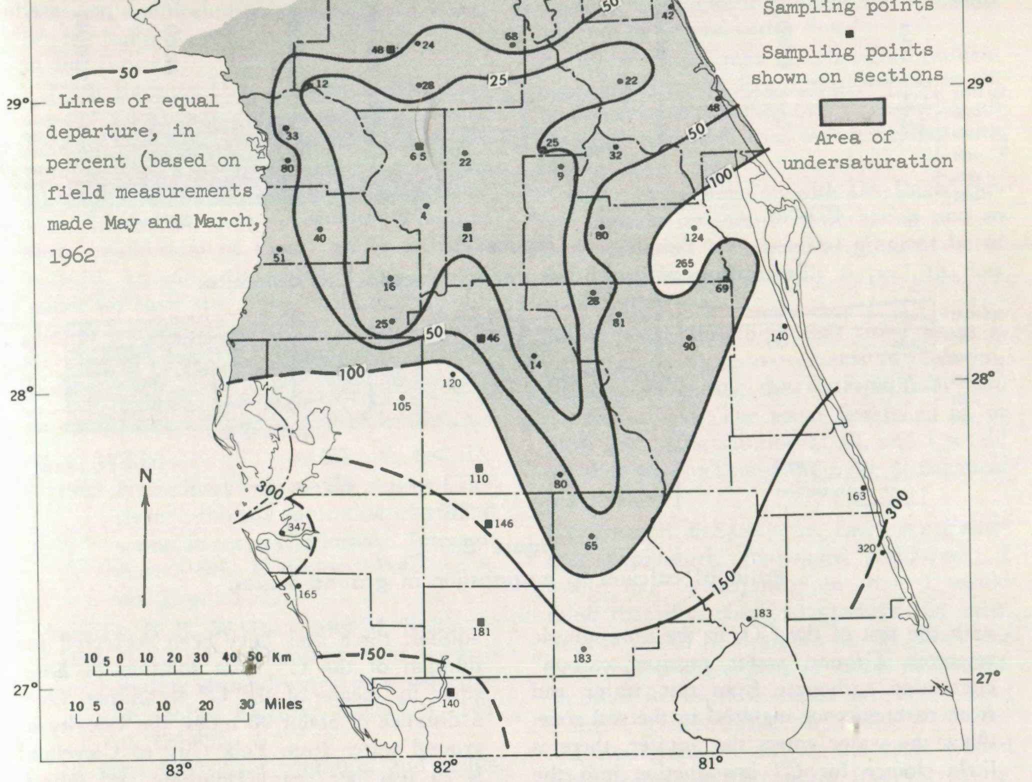

Figure 6.

Departure from equilibrium with respect to dolomite (K_{dolo}/K_{iap}) x 100.

precipitate and later invert to calcite. In the ocean and in closed-basin environments, aragonite (in preference to calcite) is the usual phase to precipitate out of solution. Perhaps in ground water also, aragonite, and not calcite, is the control on carbonate equilibrium.

The area of undersaturation with respect to dolomite (fig. 6) is slightly larger than the area outlined by the 125-percent line of calcite. The equilibrium constant for dolomite used in this study is 2×10^{-17}. The section (fig. 7) shows the vertical distribution of the departure from equilibrium with respect to dolomite. In the recharge area, the departure from equilibrium ranges from less than 10 percent to about 50 percent, and farther downgradient, the water becomes supersaturated, with values greater than 150-percent saturation.

The area of undersaturation is characterized by low magnesium content relative to calcium as shown in the cross section (fig. 8). The calcium-magnesium ratio (epm) is about 12 in the recharge area and decreases downgradient to the south to a value to slightly less than 1, indicating about a tenfold increase in magnesium over calcium concentration in the water. The source of magnesium in the water is probably from the solution of magnesian calcite, which has a higher solubility than dolomite.

Another phase of this work now underway is the use of carbon^{14} to establish the time at which the water was last in contact with the atmospheric reservoir of C^{14}. Radiocarbon is formed in the upper atmosphere by cosmic radiation of nitrogen; it has a half life of approximately 5700 years. When this carbon oxidizes, it forms CO_2 which then mixes

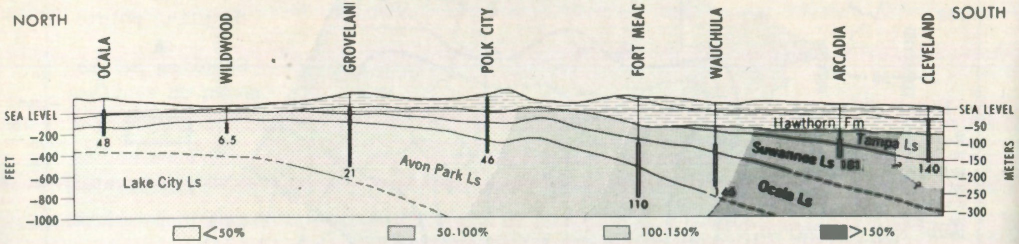

Figure 7.
Saturation of ground water with respect to dolomite.

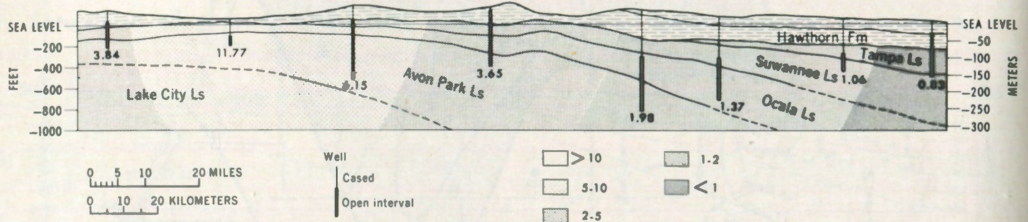

Figure 8.
Ratio of calcium to magnesium in ground water.

with the rest of the CO_2 in the atmospheric reservoir. Ground water receives carbon^{14} containing carbonate from rain water and from carbonaceous material in the soil zone. Once the water enters the aquifer, there is little chance for C^{14} introduction into the ground-water reservoir. One of the problems associated with radiocarbon dating of ground water is the many sources of nonradiogenic carbon which dilute the initial concentration of C^{14} in the water. Ingerson and Pearson (1964) have shown that it is possible to use a stable carbon isotope (C^{13}) to correct for dilution of C^{14} concentrations caused by the solution of carbonate minerals. The cross section (fig. 9) shows the radiocarbon ages of water produced from various places in the aquifer; these ages have been corrected for dilution of the C^{14} from solution of limestone by means of the C^{13} technique. Over a distance of about 90 miles, the velocity of ground water from Polk City to Cleveland is 25 feet per year (Hanshaw and others, 1965). With knowledge of the age of the water at various places in the aquifer, it is possible to gain an understanding of the kinetics of mineral reactions in natural environments. By determining the residence time of the water in the aquifer, we have some guide lines for determining how long it takes water to come into chemical equilibrium with the various mineralogic phases comprising the aquifer. The interrelations of radiocarbon ages of water in the aquifer and

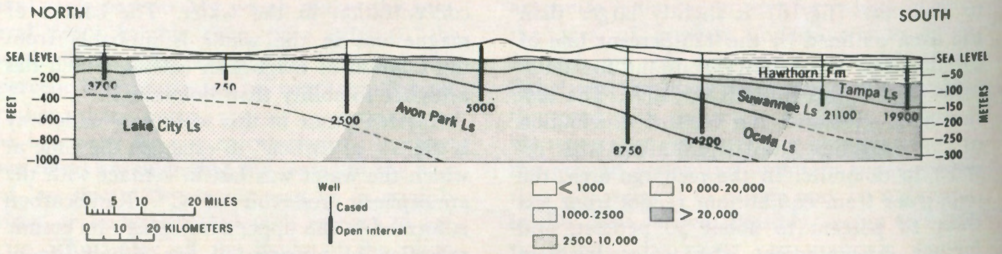

Figure 9.
Radiocarbon ages of ground water.

chemical equilibrium within the aquifer will be the subject of a more complete paper in the future.

Many intriguing questions have arisen during this work that can be answered only with detailed knowledge of the mineralogy of the aquifer. This will be the next phase of our study. We cannot obtain an explanation for the distribution pattern of the departure from equilibrium of these various minerals without knowing the mineralogy. In this paper we have attempted to show that solution of carbonate rock can occur not only in the recharge area close to the surface but may also occur in the deep subsurface.

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U. S. GEOLOGICAL SURVEY
WASHINGTON, D. C.

DISCUSSION

WILLIAM NORTH, *Berkeley, California*: If there is a chemical gradient, would this supplement the hydraulic gradient and add some impetus to ground-water flow?

BACK: We think that the chemical gradient in an environment like this is just too small. It would be masked or overwhelmed by all the other potentials that we are dealing with here.

NORTH: In conjunction with Dr. Davis' earlier remarks on microseismic action and so forth, wouldn't the chemical gradient be of the same magnitude with respect to flow in fractures?

BACK: Well, the thing that Stan Davis is saying about these movements is their importance in getting the water initially into the tiny cracks. The water has to go in, or there will be no solution at all, and I would say that we don't need the force of chemical potential.

RAYMOND E. DESAUSSURE, *Lawrence Radiation Laboratory, Livermore, California*: I would like to comment on that. I would think that if a chemical gradient did exist, then it would be opposite in sign from one that you would expect to cause movement. In other words, flow would have to be from a higher concentration to a lower.

I was going to ask, in the cross section where you show the concentration increasing, how you account for the fact that it apparently continues to increase after it reaches saturation? A change in conditions must occur between the time when the calcite is dissolved up to the point of saturation, and the time when it goes into the realm of supersaturation.

BACK: Our concept is that we are not dealing with pure calcium carbonate but are dealing with a magnesian carbonate which has a somewhat higher free energy and is, therefore, more soluble. This would account for part of the apparent supersaturation with respect to calcium carbonate. However, we believe further that it is not calcite which is controlling the precipitation of a calcium carbonate phase but, instead, aragonite. Recall that on one of the figures where we show both calcite and aragonite saturation, even though the water may be supersaturated with respect to calcite by as much as 60 percent, we never reach supersaturation with respect to aragonite. Apparently, magnesian calcite continues to dissolve to increase the magnesium content of the water so that equilibrium with respect to dolomite may be attained. At the same time, we have to get rid of the calcium component in the water, and we believe that the precipitation of aragonite takes care of this.

DESAUSSURE: In your first equation, you showed the effect of the ionic strength. This is very important. For example, some work on sea caves on the Island of Majorca related solution to the salt concentration from spray in the cave air.

BACK: Yes, much more solution of limestone occurs in areas of salt water than in areas of fresh water. This is because the additional concentration increases the ionic strength which decreases the activity coefficient and therefore allows more calcite to go into solution.

THOMAS YANCEY, *University of California, Berkeley*: How stable are these equilibrium constants over a variety of other physical conditions such as temperature?

BACK: They are all temperature dependent, and that is taken into account in our calculations. We use the equilibrium constant at the temperature of the water and use the Gibbs-Helmoltz equation for the correction. We don't compare the ion-activity products that we calculate to $K_{calcite}$ at 25°C, but rather at the temperature of the water. Temperature has an appreciable effect.

YANCEY: Does concentration of other dissolved minerals have any effect on this?

BACK: Only through the ionic strength, which is a major factor.

GEORGE L. BECK, *Finleyville, Pennsylvania*: Approximately what value do you have for the constant for dolomite?

BACK: We are using 2×10^{-17}.

BECK: The reason I ask is that two years ago there was a paper on the evolution of cave water by Holland. In prior work, the values did not correlate well together, and the reason was uncertain.

BACK: There was a big difference in the literature. We have about six different values by reputable workers, and the full range is almost three orders of magnitude, but recently there has been additional work. Don Langmuir's value, in his thesis at Harvard, is about 2×10^{-17}, and there has been a lot of indication that this is correct. Other values range from 1.0 to 2.5×10^{-17}, so I don't think that this is the problem it was two years ago. I believe only Garrels' and Robie's values differ significantly from the one we are using.

ISAAC BARSHAD, *Department of Soils, University of California, Berkeley*: In studying the solubility of minerals, what justification do you make for applying a thermodynamics which is essentially for reversible systems to irreversible systems, because the solution of minerals is a process of dissolution rather than a reflection of the solubility of the minerals?

BACK: We are dealing with a simple, almost monomineralic terrain where the solution is congruent and where the components in solution are always in contact with the mineral that we are considering. I think that in this case it is quite proper to use reversible thermodynamics. Jake, can you give us some additional help in answering this question.

JACOB RUBIN, *U. S. Geological Survey, Menlo Park, California*: The measurements test whether the system can be described using equilibrium concepts. To test this hypothesis, the system was assumed to be reversible.

BARSHAD: But we know it isn't, because all kinds of things change the system; otherwise the whole process of weathering would be understood, very simple, and open-and-shut. This has always bothered me.

BACK: The use of reversible thermodynamics in a weathering environment is open to some question, because we are dealing in that case with incongruent solution, and reversible thermodynamics would apply only to the solution of the minerals. Once some of the components have gone into solution and others have been left behind, we can no longer deal strictly with reversible thermodynamics. In our situation, however, I don't think this is a serious problem, because we are dealing with the phases and components which always remain in contact with each other.

Electric-Analog Study of Cave Formation*

By M. S. Bedinger

ABSTRACT

This study of limestone solution leading to cavern development is based on the following conditions: (1) the permeability of the limestone is low, but it contains and transmits water in joints, fractures, bedding-plane partings, and solution-channels; (2) at depth, the limestone aquifer is underlain by impermeable rock; (3) ground water in the limestone is under water-table conditions; (4) recharge to the limestone is by infiltration of precipitation through the overlying rock to the zone of saturation; (5) discharge from the aquifer is by seeps and springs; (6) ground water dissolves the limestone through which it flows, continuously modifying the flow pattern and the hydraulic properties of the medium. These conditions commonly are found in terrains of limestone of Paleozoic age in the eastern and central United States. An electrical analog conforming to these conditions has been constructed and used to define the pattern and density of ground-water flow and the relative length of time that water is in contact with the aquifer. Successive models are used to illustrate progressive limestone solution and changes in ground-water flow in the aquifer. The initial analog indicates a strongly convex water table with the greatest density of flow at shallow depth beneath the water table near the point of discharge. Successive models indicate greater concentration of flow near and on the level of ground-water discharge, an over-all lowering of the water table, and a pronounced flattening of the water table near the discharge point.

Results of the analog study support the following conclusions: (1) solution channels generally decrease in size with depth and with lateral distance from the point of ground-water discharge; (2) the larger solution channels are above or near the elevation of the point of discharge and have greater lateral than vertical extent. This second observation supports the theory that the major zone of cave formation is at shallow depth beneath the water table.

INTRODUCTION

The predominance of observational data supports the theory that limestone solution leading to the development of caves principally is by ground-water flow in the zone of saturation (Davis, 1930; Bretz, 1942 and 1956; Davies, 1960; White, 1960; Moore, 1960; Moore and Nicholas, 1964). Considerably less unanimity of opinion exists as to whether the greater amount of solution is deep within the aquifer or at relatively shallow depth beneath the water table. Davis (1930) and Bretz (1942) held that major cavern development is at relatively great depth below the water table. Recently, Davies (1960) and White (1960), while subscribing to the view that major cavern development is within the zone of saturation, have presented evidence, based on morphology, spatial relations, structure, and lithology, that major solution is at shallow depth.

This paper describes the use of electrical models analogous to the ground-water conditions prevailing in many limestone areas of the eastern and central United States. From these models the temporal and spatial characteristics of ground-water flow and of solution in a limestone aquifer are inferred. The patterns of limestone solution developed by the models compare favorably with ob-

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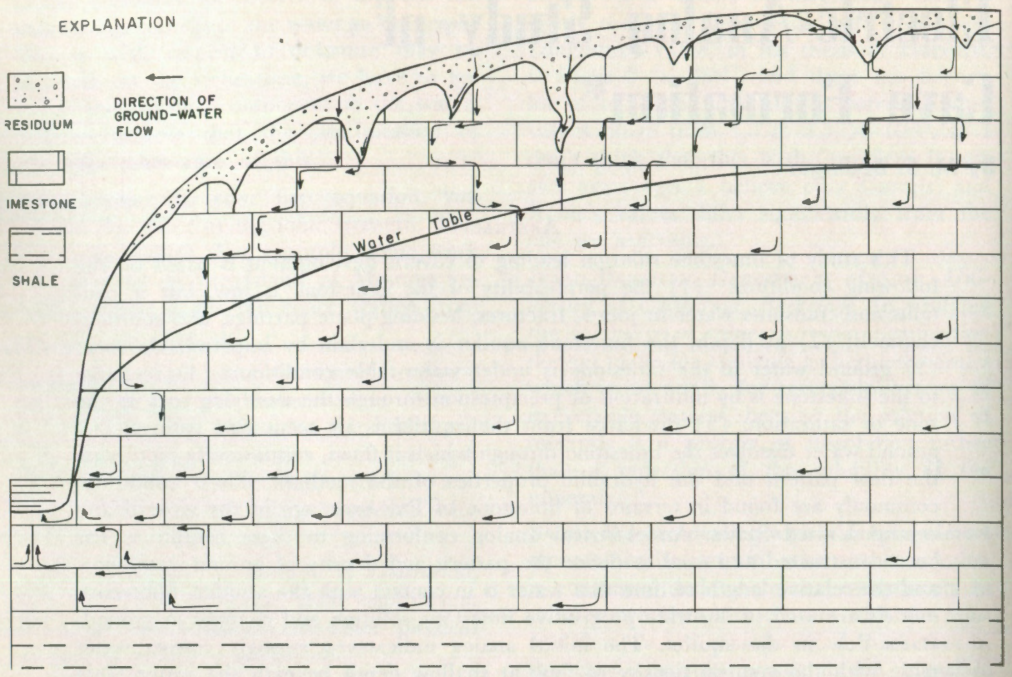

Figure 1.
Geohydrologic section through a limestone aquifer.

servations of limestone-solution features in the Appalachian region. The analog study supports the view that the major zone of solution is at shallow depth beneath the water table.

Ground-Water Flow System

The ground-water flow system chosen for modeling comprises characteristics of flow systems common to many areas underlain by limestone of Paleozoic age in the central and eastern United States. The conditions of the system may be summarized as follows: (1) the primary permeability and porosity of the limestone is low, and it is an aquifer by virtue of joints, fractures, bedding-plane partings, and solution enlargement of these secondary features; (2) at depth the aquifer is underlain by impermeable material; (3) water in the aquifer is under water-table conditions; (4) recharge to the aquifer is by infiltration of precipitation through the overlying rock and soil to the zone of saturation; (5) discharge from the aquifer is by seeps and springs to streams.

A geohydrologic section through part of a limestone aquifer, illustrating a complete flow system before appreciable solution has taken place, is shown in figure 1. The aquifer is areally extensive; however, no water is exchanged with the similar system to the right because of the ground-water divide. Likewise, there is no exchange of water with the flow system to the left because, as the head conditions indicate, all flow is discharged to the stream.

Electric-Analog Model

The foregoing criteria describe a ground-water flow system that can be duplicated by an analogous electrical flow system. The analogy between electrical and ground-water flow permits close study of an electric equivalent of the ground-water flow system reduced to laboratory dimensions (Skibitzke, 1960).

Conceivably, an unending number of combinations of anisotropic and nonhomogeneous aquifers could have been chosen for modeling. However, an isotropic prototype was

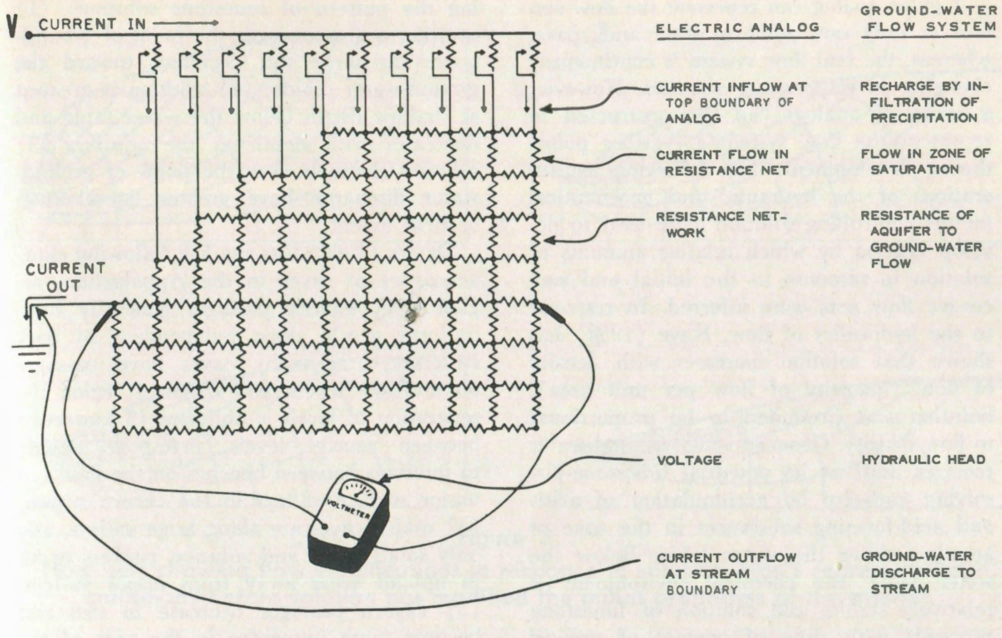

Figure 2.
Schematic drawing of electric analog of the ground-water flow system.

selected for initial analysis because it provided for greater facility in modeling and presentation of results. The analog results, however, can be interpreted as a series of anisotropic, but homogeneous, models having various ratios of horizontal to vertical permeability. For example, if it is assumed that the vertical scales in the cross sections (fig. 1 and 3) are exaggerated, the flow nets become valid for a prototype aquifer having a greater horizontal than vertical permeability. The ratio of horizontal to vertical permeability is increased in proportion to the square of the vertical exaggeration.

A schematic drawing of the electric circuit and the analogous relations between ground water and electricity are shown in figure 2. The analog consists of a rectangular network of resistors representing the aquifer. The physical size of the analog is scaled to the prototype aquifer, and in the initial analog each resistor was of the same value so as to represent an isotropic condition. Recharge to the aquifer was simulated by electric currents at each resistor junction on the upper boundary. Under water-table conditions, the upper boundary of the aquifer is coincident with the water table. Also, the hydraulic head at the water table is equivalent to the elevation of the water table above the point of discharge. The upper boundary of the model could not, therefore, be predetermined, but was determined by trial and error as part of the solution of the ground-water flow problem.

After the shape of the water table was determined, the voltage (analogous to hydraulic head) in the network was read. Head lines were constructed to represent 10 equal drops in hydraulic head from the point of highest head at the ground-water divide to the point of lowest head at the stream. Similarly, flow lines were constructed to separate the total flow into 10 equal parts. The analysis of the initial analog (flow net 1 in fig. 3) provides data on density of flow and time of flow under homogeneous conditions. Flow net 1 in figure 3, therefore, represents the flow conditions before solution has modified the initial hydraulic properties of the aquifer.

A given analog can represent the flow system at only one point in time and space, whereas, the real flow system is continuously changing in response to solution. However, a series of analogs can be constructed to represent the flow system at various points during development. The following considerations of the hydraulic and geochemical factors controlling solution were used to develop criteria by which relative amounts of solution in response to the initial and successive flow nets were inferred. In response to the hydraulics of flow, Kaye (1957) has shown that solution increases with density of flow (quantity of flow per unit area). Solution was presumed to be proportional to flow density. Geochemically, ground water receives most of its potential limestone-dissolving capacity by accumulation of acids and acid-forming substances in the zone of aeration above the water table. Below the water table, the chemical environment is relatively stable, and solution of limestone proceeds with time of contact of ground water with the aquifer until chemical equilibrium is reached between the water and its environment. In response to the geochemical factor, it was assumed that solution decreased in proportion to time of contact of ground water with the aquifer.

The resistance values in each successive analog were based on calculations of flow density and time of travel from the preceding analog. Resistances were decreased (change in resistance being inversely proportional to amount of solution) in proportion to flow density and inversely in proportion to time of travel.

Flow nets for the initial aquifer and three successive aquifers modified by solution are shown on figure 3. Head lines have been constructed to represent 10 equal drops in hydraulic head from the point of highest head at the ground-water divide to the point of lowest head at the stream. Similarly, flow lines separate the total flow into 10 equal parts.

Pattern of Solution

The relative amounts of solution in the aquifer are shown by the line widths in figure 4. Based on the analog study, the following generalizations can be made regarding the pattern of limestone solution: (1) solution is greatest near the point of ground-water discharge and decreases toward the ground-water divide; (2) solution is greatest at shallow depth below the water table and decreases with depth in the aquifer; (3) solution channels near the point of ground-water discharge have greater lateral than vertical extent.

Davies (1960) lists the five following characteristics of caves in the Appalachian region: (1) cavern passages generally have uniform gentle slope independent of rock structure; (2) many caves have passages on multiple levels, and within a region the separation of levels is uniform; (3) intervals between passage levels correspond closely to intervals between benches on the flanks of major surface valleys in the cavern region; (4) major caves are along large valleys, and only small caves and solution pockets occur in upland areas away from major valleys; (5) cavern passages decrease in size and become more numerous in the part of the cave away from major surface valleys. In addition, caves in the Appalachian region have a much greater lateral than vertical extent.

The pattern of limestone solution inferred from the analog is compatible with the characteristics of Appalachian caves observed by Davies. The fact that many of Davies' generalizations are tied to surface-water drainage features is significant. As shown by the prototype model, the major surface stream is the discharge level of the ground-water flow system, and is, therefore, the major factor that determines the vertical position of caves.

Davies' observations indicate that solution channels are larger and fewer in number near the major streams. This observation involves, in part, a third dimension, parallel to the stream and ground-water divide. This dimension was not modeled with the analog. Strict interpretation of the analog results would indicate that, with the uniform conditions chosen for analysis, solution at right angles to the plane shown in the analog would be uniform. However, it is in the third dimension that rock structure and inhomogeneities have the greatest influence.

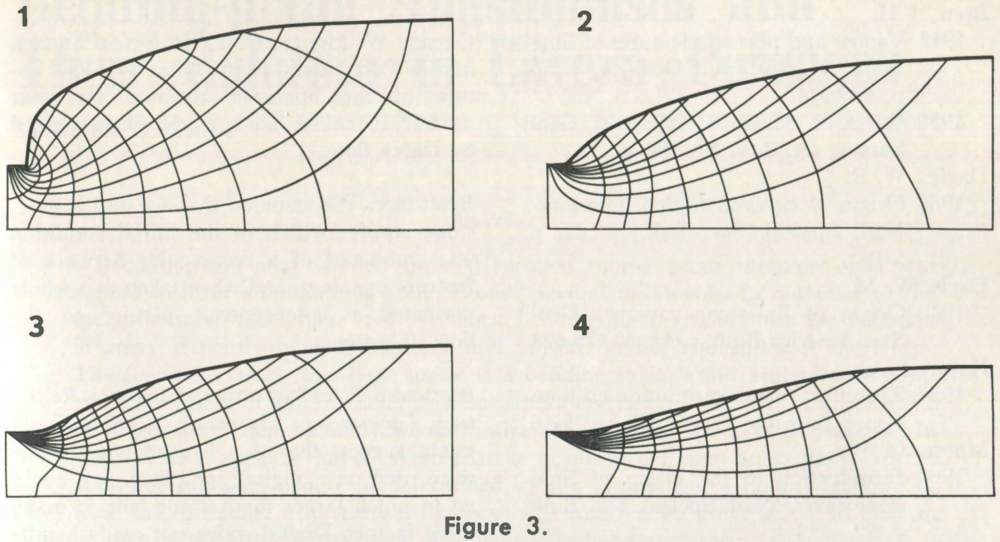

Figure 3.
Flow nets showing flow modifications in response to solution. Net 1 represents flow conditions before solution has modified the initial properties of the aquifer.

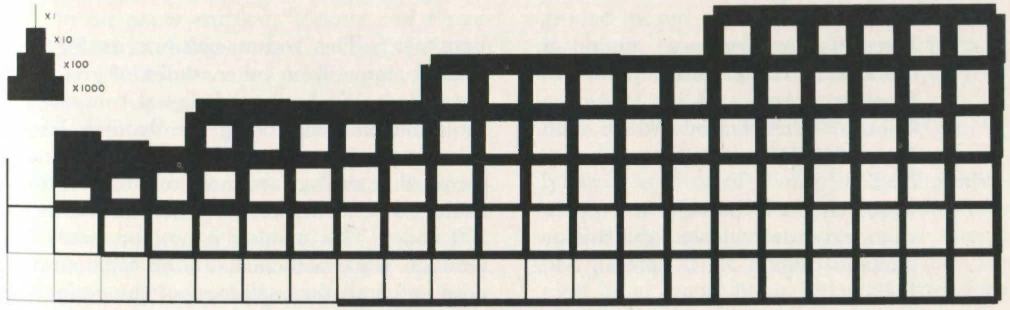

Figure 4.
Diagram showing relative amounts of limestone solution. The explanation at the upper left indicates the increase in permeability relative to the initial analog model.

Conclusions

This study supports the following conclusions: (1) The most active zone of solution is at shallow depth beneath the water table. Consequently, solution channels generally decrease with depth. (2) Solution channels increase in size and decrease in number with proximity to the point of discharge. (3) Solution channels are more numerous, though smaller, near the ground-water divide. (4) Solution channels generally have greater lateral than vertical extent. (5) The lateral trends of caves are controlled by the regional direction of ground-water movement, and the vertical position of caves is determined by the hydraulics of ground-water flow and the chemistry of the water. The flow of ground water in both the horizontal and vertical planes is in turn controlled by perennial streams. Complexities of rock structure and inhomogeneities result in minor variations in solution within the latitude of control exerted by ground-water flow and account for the infinite variety found in individual caves.

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Discussion

George W. Moore, *U. S. Geological Survey, Menlo Park, California*: What was the nature of the openings assumed for your model? It looked like a model characterized by Darcy flow.

Bedinger: We assumed that we had a limestone aquifer which in the initial condition was composed of a rectangular network of fracture openings, and when taken as a whole presented a homogeneous picture to the flow of water.

Raymond E. deSaussure, *Lawrence Radiation Laboratory, Livermore, California*: I might suggest that if a hydrodynamic code were used on a digital computer, you could go to much larger meshes and put in many more factors. Digital codes can easily handle meshes of 1000 that are laborious when you are taking point measurements from an analog system.

Bedinger: The resistor-network analog is used in many of our other studies of groundwater flow. We have used digital computers for some problems but generally find, considering time, cost, and accuracy requirements, that analogs are more suitable. In the analogs described here, there were fewer than 200 nodes. This number is not too cumbersome to work with and, at the same time, does not limit the usefulness of the analysis.

Bedding-plane Anastomoses and Their Relation to Cavern Passages

By Ralph O. Ewers

Abstract

Bedding-plane anastomoses (braided solution tubes) occur in many sizes and appear to form a continuum from channels several millimeters in diameter to the spaces between the largest roof pendants. Bedding-plane anastomoses are common in areas of poorly jointed limestone and appear on the undersides of the strata. The features extend over large areas of a bedding surface and are strongly influenced by minor fractures. Bedding-plane anastomoses are unquestionably phreatic in origin and often certainly predate adjacent or confluent cavern passages. In many cases it appears that a cavern passage results either from an extension of the anastomoses along a route predetermined by the presence of a minute fracture or from the breaching of a stratum by growth of anastomoses from below where two or more sets exist superimposed on adjacent bedding surfaces.

Introduction

In his paper entitled *"Vadose and Phreatic Features of Limestone Caverns"*, J Harlen Bretz (1942) described bedding-plane anastomoses from the St. Genevieve Limestone, and pointed out their wide distribution in that formation. He proposed a phreatic origin for these features and noted their apparent relation to subsequent phreatic and vadose activity.

Bedding-plane anastomoses are braided, freely interconnecting, networks of solution tubes which appear on the undersides of soluble sedimentary strata. The smallest of these tubes are roughly circular in cross section with diameters as small as several millimeters. These form a continuum with the largest type, which are channels of ovoid to trapezoidal cross section having broad bases and narrow tops, occasionally exceeding several feet in width. These form the spaces between features that are commonly called roof pendants after Bretz (although Bretz wished to reserve this term for features of vadose origin). Anastomosing solution tubes of intermediate size are usually characterized by vertically elongated cross sections with narrow bases and broad rounded tops (fig. 1-2). These features generally follow bedding surfaces, but any nearly horizontal plane of weakness capable of conducting ground water may be exploited by them, including cross-bedding surfaces. Because the anastomoses apparently have dissolved upward from these surfaces, there can be little doubt about their phreatic origin.

Bedding-plane anastomoses most frequently occur in areas of poorly jointed limestone where subsurface drainage is dependent upon bedding-plane-oriented routes for lateral percolation. In the cavern situation they are most often found as small, lateral extensions of meandering tubular passages. These extensions follow bedding planes which intersect the passages (fig. 3).

How a freely interconnecting network of similar-sized tubes can be developed, preserved and enlarged with the evolution of no large-scale integration is an important question. The answer to this question seems to lie in the assumption that the enlarging process must be operating uniformly throughout the system, regardless of the position of a tube in the network. If the hydraulic head and flow rate were extremely large and remained so in spite of the growth of the system, considerable flow would be induced in all of its parts, and conditions at the interface between the ground water and the rock would be similar

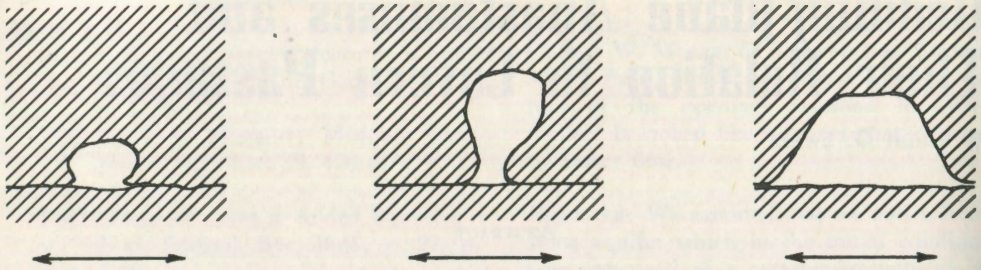

Figure 1.
Typical cross sections of solution tubes in anastomoses.

throughout. An alternate and more likely circumstance involving very slow ground-water movement, which was first suggested by Bretz (1942), may also produce uniform enlargement. Where the water movement in the system is sufficiently slow, the difference in resistance of the various paths may become insignificant, leading to similar rates of flow in all parts of the system.

Experimental Work

Early model experiments with phreatic solutional features along horizontal planes in salt blocks (Ewers, 1964) suggested that a similar technique employing a different scale might prove successful in exploring the mode of occurrence and significance of bedding-plane anastomoses. The early experiments and field evidence indicated that the presence of soluble material beneath the bedding plane is not pertinent to the primary development of phreatic bedding-plane features. Tubular cavern passages often do not significantly penetrate the stratum beneath the plane of weakness along which they develop. The same case is usual for all observed types of anastomoses. The only effect of the lower stratum in the dissolving process seems to be the restriction of solvent circulation. In view of this fact, it seemed possible that a transparent material could be substituted for the lower salt stratum in a model experiment to provide direct and continuous observation of the developing solutional features and processes. As of the date of this writing, however, continuous observation has not been possible. The major difficulty lies in producing a parting between two strata which closely approximates, on a reasonable scale, the partings which exist in nature between limestone strata.

In this series of experiments two methods of producing partings were used. The first involved casting the salt block in plaster of Paris. This method produced variable results due to premature solution of the salt by moisture in the plaster. In the second and most successful method, partings were produced by vacuum molding an acetate film over the bedding surface and backing it with a resilient substance to insure the maintenance of close contact with the salt.

Following the initial filling with a saturated salt solution, a solution of 80 percent saturation was introduced instead of the fresh water used in the earlier experiments. This saturation change produced a scale increase ten times that of the previous experiments. The features developed under these conditions appear to simulate natural-cavern features on the order of 100 times larger (fig. 4).

Anastomoses almost certainly predate the adjacent or confluent cavern passages that facilitate their observation. In most limestone quarrys, anastomoses are exposed which have no associated larger tubes. This situation occurs with such frequency that association with larger tubes may be the exception rather than the rule. Model studies support the premise that anastomoses primarily form independently of larger tubes. A large, low-resistance, phreatic path in a salt block fails to generate any but the

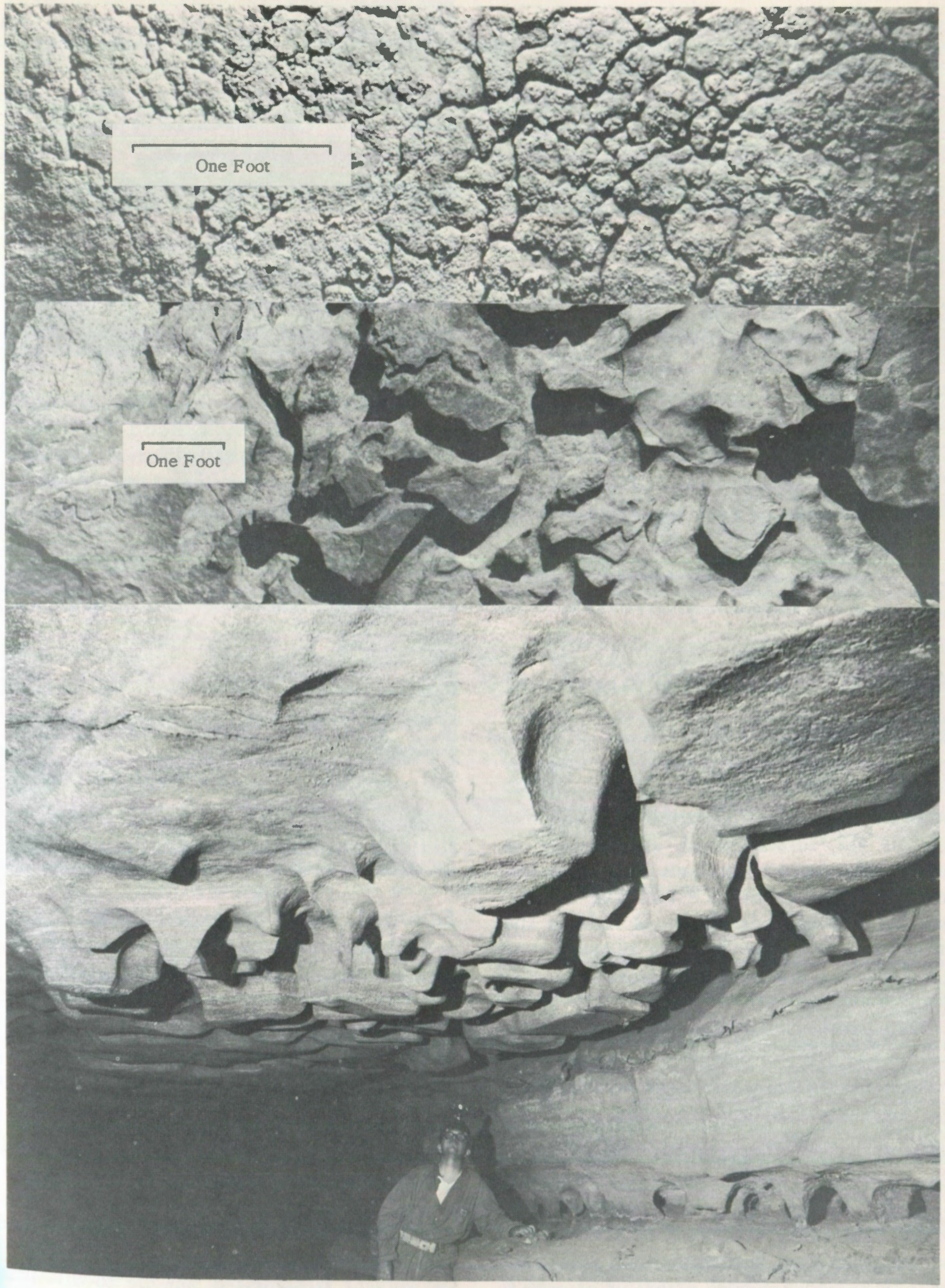

Figure 2.
Anastomoses of small, intermediate, and large size. Ceiling views, Crystal Cave, Mammoth Cave National Park, Kentucky.

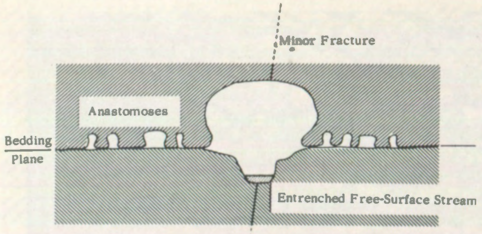

Figure 3.
Typical tubular-passage cross section showing anastomoses.

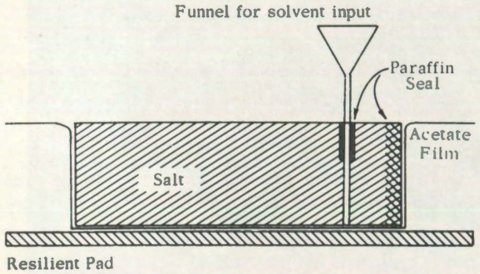

Figure 4.
Apparatus for anastomosis experiments.

simplest anastomoses in surrounding areas, and these are rapidly incorporated into the enlarging main tube. A free-surface path for solvent movement in an adjacent joint also inhibits the formation of anastomoses in model experiments. In both cases, the lessening of flow resistance and the accompanying loss of hydraulic head reduces the circulation in nearby bedding planes below a value which will provide adequate flow to develop anastomoses.

The experiments which succeeded in producing anastomoses over broad areas of bedding surfaces were confined to those models in which a hydrostatic head of at least 50 millimeters could be maintained for periods of 12 hours or more, with an available solvent input of less than one cubic centimeter per minute. This corresponds to more than a thousand times the head, four times the period, and one seventieth the flow rate used in previous model experiments which gave cavern-like results. The pressure required to produce circulation and significant solution along bedding planes can be re-

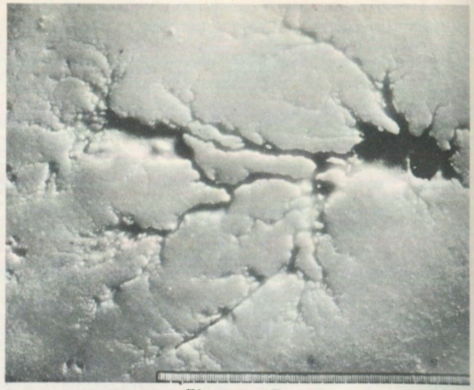

Figure 5.
Experimentally-produced anastomoses in salt, prior to hydraulic head reduction. Flow, right to left; smallest scale divisions are millimeters.

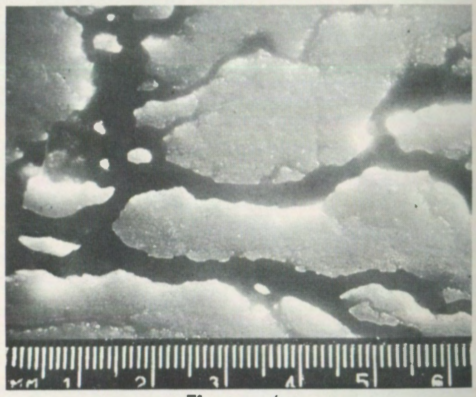

Figure 6.
Enlarged view of experimentally-produced anastomoses in salt. Flow, left to right.

duced by providing a poor mating of the strata. In such cases anastomoses will not develop; instead wide, flat channels with irregular borders form, which bear no resemblance to any cavern feature so far observed.

Anastomosis development in the models proceeds from the point of solvent input in a subdendritic pattern (fig. 5). Beyond the extremities of the advancing solution tubes, the solvent is saturated, unable to dissolve additional pathways and moves to the discharge point by slow seepage along the bedding surface. As the tubes enlarge behind

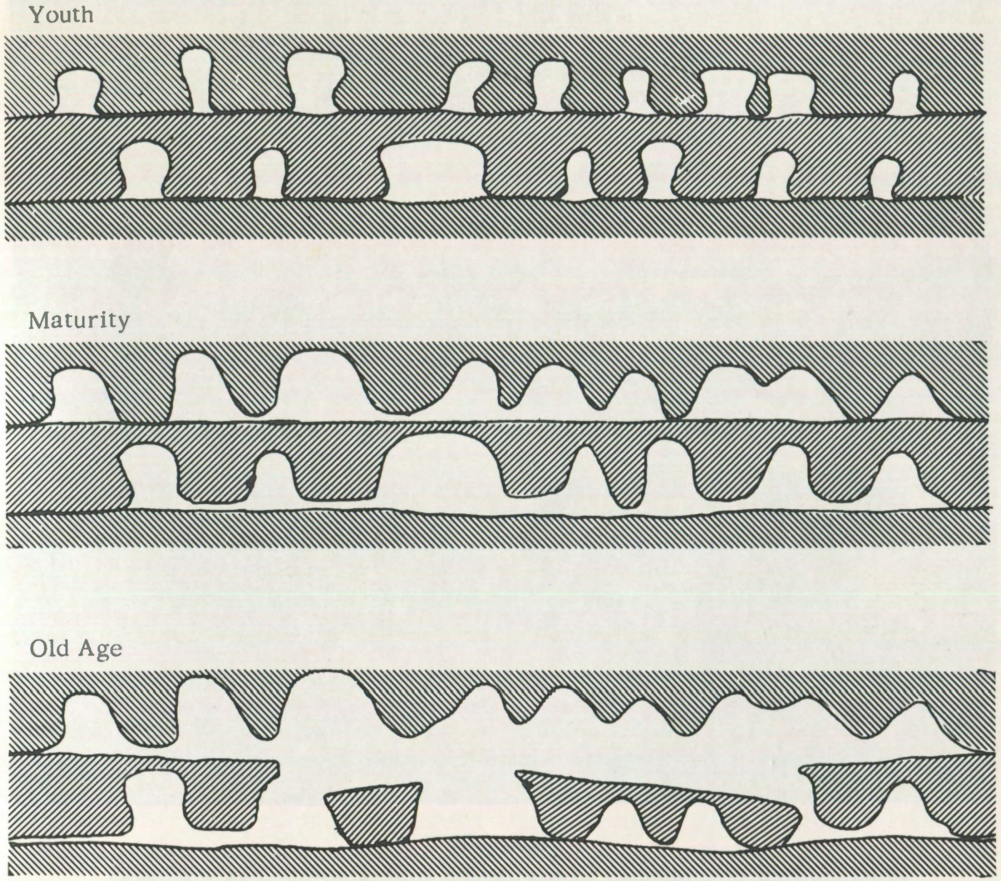

Figure 7.
Stages in the development of anastomoses.

the advancing perimeter of undersaturation, they become freely interconnecting, and the braided pattern of typical anastomoses develops (fig. 6). Provided that the flow rate remains small and a large hydraulic head is maintained, the anastomoses will grow and the diameters of the tubes at a given distance from the input point will remain similar. When the anastomoses reach the edge of the block, the pressure drops to an unmeasurable figure, and the most efficient route to the discharge point carries the bulk of the solvent. The anastomoses then cease to grow, and a rapidly enlarging central channel engulfs them. These tubes are very similar to naturally occurring anastomoses

Figure 8.
A cavern passage produced by collapse. Anastomosis remnants are visible on the undersides of the fallen blocks. Crystal Cave, Kentucky.

Figure 9.
Superimposed anastomoses in a passage wall. Crystal Cave, Kentucky.

in pattern and cross section, but their decrease in size with increasing distance from the ground-water source has yet to be demonstrated in nature on a large scale.

Discussion

In their relation to the stratum in which they are developed, a series of anastomoses exhibits a similarity to the cycle of stream erosion, and it is often convenient to speak of them in terms of their youth or maturity (fig. 7). In their early or youthful stage the tubes are small and cover a small fraction of the bedding surface. As they mature, they may cover nearly all of the under surface of their stratum. With the onset of old age their stratum is largely unsupported and, in the case of thin strata, breaching and collapse may occur. Where anastomoses exist superposed on adjacent bedding planes, breaching and collapse may play an important role in the development of cavern passages (fig. 8-9). This erosional sequence should not, however, be construed to be cyclic. It is a straight-line route to destruction and is not repetitive in a given area due to the large, low-resistance water course that it produces.

Several factors, aside from the gross permeability and the ratio of bedding-plane to joint permeability of a sequence of limestone strata, seem to influence the production of anastomoses. Small-scale irregularities of the bedding surface, including stylolites, may materially influence the course and complexity of these tubes. When these irregularities were simulated in models by the application to the bedding surface of small

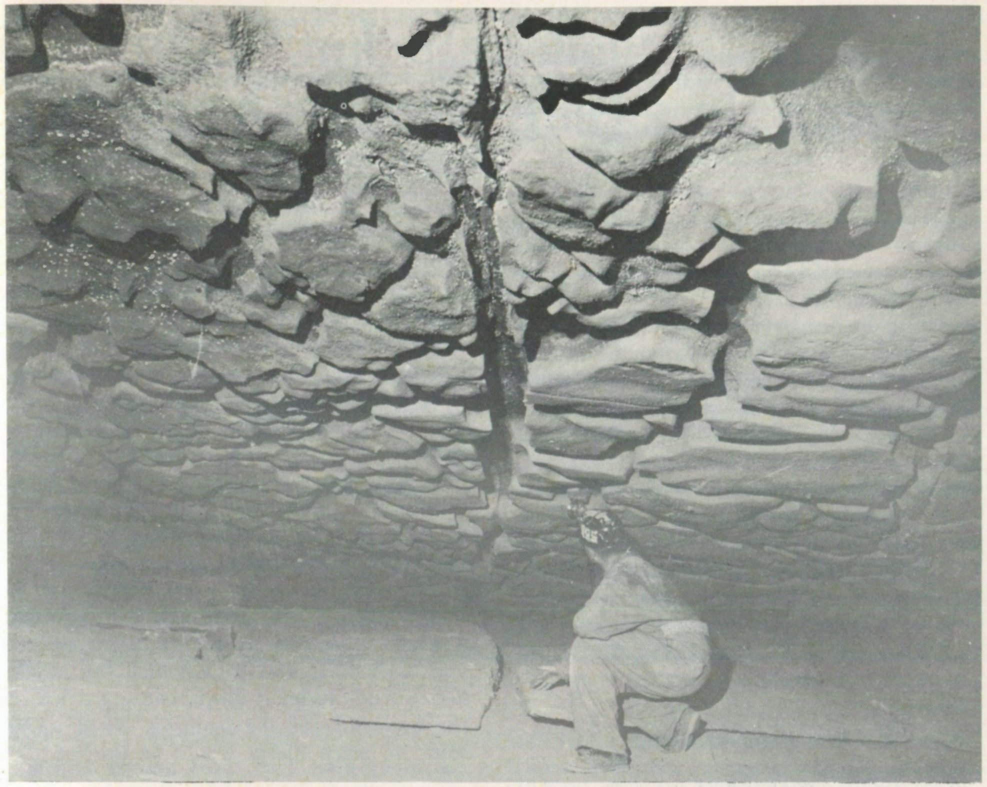

Figure 10.
Minor fracture control of an anastomosis. Crystal Cave, Kentucky.

amounts of petroleum jelly in a finely divided, random pattern, a more complex anastomosis was produced. Minor fractures strongly influence the pattern of an anastomosis, usually producing a long, straight tube along the intersection of the fracture plane and the bedding plane with a diameter that is similar to the average tube in that area. These straight tubes may become the most efficient routes for ground-water movement, following loss of resistance at the discharge point of the system (fig. 10). Large-scale irregularities, such as high primary permeability may influence the position of the anastomoses. A slightly greater tendency for tube formation at the higher side of a tilted salt block may indicate a preference for structural highs in the development of anastomoses.

Conclusions

On the basis of natural evidence and salt-block model experiments the following conclusions seem to be in order: (1) Bedding-plane anastomoses are probably among the earliest solutional openings in soluble sedimentary strata. (2) They form in strata where bedding planes provide the most important avenues for ground-water percolation. (3) They continue to develop while flow rates are very small and the hydraulic head is large. (4) They cease to grow significantly when the system resistance becomes low. (5) Efficient flow paths through anastomoses may develop into cavern passages after head loss in the system. (6) Long, relatively straight tubes, resulting from the presence of minor fractures (joints), often provide the most efficient flow paths. (The solution-tube, anastomosis, minor-fracture relationship of figure 3 is common.) (7) Anastomoses superposed on adjacent bedding planes may contribute to the formation of cavern passages through breaching and collapse.

Until this point, no mention has been made of the mechanism by which the upward progress of solutional enlargement takes place. Bretz (1942) proposed that insoluble residue from the dissolving limestone accumulating in the bottom of the tube, under conditions of very slow water movement, protected the limestone below from solution. The upward progress of solution in salt blocks is clearly produced by a saturation gradient. The more concentrated solution, due to its density, accumulates at the bottom of the tube and protects the lower stratum and lower walls from dissolving. Whether the density-gradient phenomenon is simply an analog of the accumulating-sediment process, or whether it is a process that itself contributes to the formation of natural anastomoses is not known at this time.

Acknowledgments

For their contributions to the present study, I should like to thank C. Thomas Klekamp for his assistance in conducting the experiments; the Cave Research Foundation for help in securing the photographs and for partial financial support; and Lynda M. Jones and Pamela Gray for assistance in preparing the manuscript.

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Discussion

Philip R. Pennington, *Department of Metallurgy, University of California, Berkeley*: In the dissolution of crystalline materials, crystal imperfections such as dislocations are proving to be more important than generally thought. Some years ago I did some experiments with silicon which produced forms strikingly similar to what you have here. We took crystals and heated them and quenched them. There was no apparent surface damage under high magnification, but then we etched them to dissolve the silicon and got patterns which are the type you show here. These were due to regions of very high dislocation density. Now, in regions of high dislocation density, dissolution is enhanced. This should be true in ionic crystals such as calcite as well as in silicon. The pattern which we got in the silicon was due to the stress distribution in the material. I can picture in your situation a similar mechanism taking place, where you have horizontal beds, which the weight above tends to bend down, causing stress on the underside of the bedding. Then under solution, anastomoses would be formed.

Leigh Readdy, *Kennecott Copper Co., Salt Lake City, Utah*: Along the same line, a number of years ago Kurt McLane did some work with plaster of Paris and found that solution was very definitely stress dependent. We're talking about a total geologic system here, and we see many evidences of the relationship between solution and stress.

Occurrence and Movement of Ground Water in Carbonate Rocks of Nevada

By George B. Maxey and Martin D. Mifflin

Abstract

Paleozoic and Mesozoic carbonate rocks underlie large parts of eastern and southern Nevada and constitute important aquifers there. In studies of carbonate-rock hydrology, two general classifications of ground-water flow systems can be made: (1) Local flow systems, where generally drainage areas are small, flow paths are relatively short, interbasin flow is not extensive, springs have large fluctuations in discharge, and the water is usually characterized by low temperature and low concentrations of Na, K, Cl, and SO_4. (2) Regional flow systems, where generally drainage areas are large, flow paths are long, interbasin flow is common, springs have steady discharge, and the water is usually characterized by relatively high temperature and higher concentrations of Na, K, Cl, and SO_4. Conventional speleologic and geologic investigations alone are not powerful enough tools to characterize most flow systems in Nevada, and several additional hydrologic approaches have been employed. These include water-budget studies, water-potential studies, and water-chemistry studies. Although detailed delineation of most flow systems in Nevada has not been accomplished, integration of geologic and hydrologic methods permits an approximate portrayal of many systems.

Introduction

Springs issuing from carbonate rocks in Nevada are estimated to supply in excess of 215,000 acre-feet of water a year, primarily in the eastern third of the state, from near the northern boundary to the latitude of Las Vegas (fig. 1). Apart from Las Vegas, Pahrump, and Diamond Valleys, where considerable water is pumped, these springs constitute essentially the only large perennial supply of water in a strip about 100 miles wide along the east boundary of the state, from the southern drainage divide of the Humboldt River Basin to Las Vegas. Therefore, most of the economy of the area is based on them. As far as is known, little or no water is produced from wells penetrating limestone, although wells finished in alluvial sediment may withdraw water that is recharged from limestone aquifers. Indeed, it is likely that in places in eastern Nevada extensive development of alluvial reservoirs would result in diminished spring flow.

The quality and temperature of water from the springs are characteristic of those of waters in other carbonate terrains, and, apart from being hard, the water is satisfactory for most common uses and is highly favorable for the present predominant use—irrigation.

Some of the springs are associated with large cavern systems, whereas others issue from fractures and crevices that seem to be only slightly enlarged by solution. In all instances the limestone and dolomite making up the country rock around the springs is fractured, creviced, and contains at least a few units displaying cavernation to some degree. Several of the springs mentioned in this paper—most notably Las Vegas, Pahrump, and Manse Springs—issue from alluvial sediment in the valleys but obtain recharge from predominantly carbonate formations that crop out in adjacent mountains and probably underlie the sediment. In several instances interbasin underflow of ground water through carbonate rock has been recognized.

Figure 1 illustrates the known occurrences of caves developed in carbonate rock in Nevada. However, it should be noted that because of limited accessibility and the sparse population in the mountainous terrain where most of the carbonate rocks crop

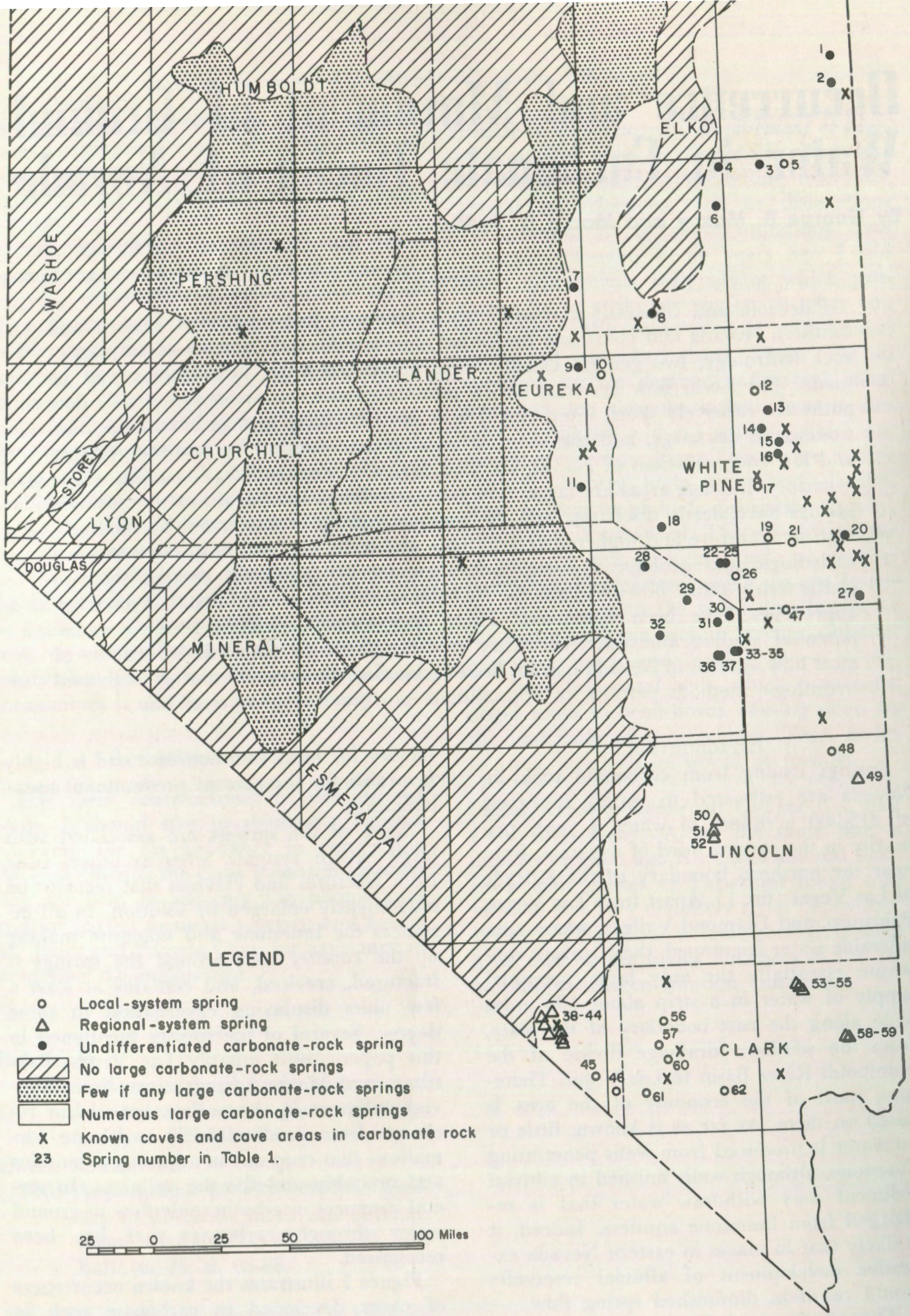

Figure 1.
Carbonate-rock distribution and associated springs and caves in Nevada.

out, many other caves are likely to be unreported and undiscovered. Of the caves known and explored, the majority occur above the zone of water saturation in the mountainous areas, and conventional exploration is of limited value in attempts to understand the present occurrence and motion of ground water in carbonate-rock terrain.

Detailed quantitative or definitive studies of the hydrology of carbonate rocks in Nevada are not available and have not been made because of several factors, the most important of which are: (1) the complex lithologic and structural pattern of the geology of Nevada; (2) lack of subsurface records; (3) lack of hydrologic records; (4) lack of economic pressure to develop further supplies of water in the region.

The results of work already done demonstrates that the present philosophy and practice of legal allocation of ground water in the state is not securely based. For many years in Nevada, as in other western states where the Appropriation Doctrine is recognized, the amount of water allocated for a given basin has usually been determined from estimates of the perennial recharge derived for the topographic drainage area of the basin. Obviously, demonstration that interbasin flow is not only possible but occurs in many instances both locally and regionally places this practice in doubt. Management and legal organizations must adopt practices which recognize the actual occurrence and motion of water in carbonate terrain.

ACKNOWLEDGEMENTS

Besides those investigators cited in this paper, the writers wish to acknowledge the work Alvin R. McLane contributed in the form of literature search and fieldwork in connection with spring and cave data. Our work has been supported by the Office of Water Resources Research, Department of the Interior, and by a special ground-water research fund appropriated by the legislature of the State of Nevada.

GROUND-WATER SYSTEMS

Characteristics of Ground-Water Systems—In eastern and southern Nevada two large categories of ground-water systems are now recognized which have been defined in general by studies of ground-water potential, geologic and physiographic characteristics, and hydrologic features, including areal budget studies. These categories include local systems and regional systems.

Local ground-water systems are commonly characterized by: (1) small drainage area and relatively short flow paths; (2) boundary controls consisting of local lithologic, tectonic, and topographic features; (3) spring-discharge points with water of relatively low temperature that often fluctuates considerably, and with discharges that vary broadly both seasonally and even with local storms; (4) many involve little or no interbasin transfer of ground water; (5) they may or may not be a part of a larger regional system, that is, they do not necessarily contribute water to a system where water motion may encompass a broad area.

Of the 62 springs included in this paper, at least 24 are recognized as being in discharge areas for local systems. In addition, many smaller springs not included in this paper are known to be associated with local systems.

Regional ground-water systems are commonly characterized by: (1) large drainage areas generally encompassing two or more topographic basins and relatively long flow paths; (2) boundary controls are regional; (3) discharge areas include springs and spring areas where discharge and temperature fluctuate little, spring discharges are large, and temperature is relatively high; (4) interbasin transfer of water occurs; (5) the system may or may not be hydraulically connected to local systems occurring in part of the same drainage area.

Of the springs discussed in this paper, 17 are considered to be discharge points in regional systems.

Present Knowledge of Ground-Water Systems—In only two areas of the state have adequate studies been published to define reasonably, in a general way, the nature of regional ground-water systems. One of these areas is reported upon by Eakin (1965), and the other is the Nevada Test Site and vicinity reported upon by Winograd (1962 and 1963). Several other studies by Eakin and Winograd (1965), Winograd and Eakin

(1965), Eakin (for example, 1963a, 1963b, 1963c), and Maxey and Eakin 1949, p. 40 and 45) indicate the probable existence of systems which involve transfer of water over long distances through previously presumed impermeable bedrock masses and across areas which form drainage divides on the land surface.

Eakin's approach involves, in addition to conventional geologic and hydrologic analysis, a water-budget study of individual valleys combined with balancing of a regional budget and determination of ground-water gradients from valley to valley to verify the possibility, direction, and amount of subsurface flow. He has shown that although budgets for individual valleys do not balance and that there must be underflow from them, the regional budget does balance, and the gradients allow flow of water from valley to valley within the region and in a generally southerly direction. Thus, he has demonstrated that water might be transferred from north to south through a system of topographically isolated basins extending from the latitude of Ruby Lake in Elko County to Moapa Valley in southern Nevada, a distance exceeding 250 miles.

In this paper, several springs within Eakin's postulated regional flow system are classified as regional. These are Hiko, Crystal, and Ash Springs in Pahranagat Valley, and Muddy River, Warm, and Iverson Springs in Moapa Valley. Perhaps Rogers and Blue Point Springs near Overton, Nevada, should also be included in this same system.

Winograd and his associates have determined from geologic mapping, test-drilling, and regional analysis that a broad area in Nye County and adjacent parts of Clark and Lincoln Counties are underlain by Paleozoic limestone at considerable depth which allows transfer of water in a system that extends at least from northern Yucca Flat to Ash Meadows, a distance of 50 or more miles in an area about 30 miles wide. The data available to Winograd demonstrates that ground water moves across some mountain ranges without regard to topographic divides. Also, it allows the definition of a ground-water system, the upper part of which is a recharge area receiving water from the overlying alluvial reservoirs, and the lower part of which is a discharge area recharging alluvial reservoirs and eventually discharging water at the surface through the springs in Ash Meadows. The system defined by Winograd has not been materially affected by withdrawals from wells or other activities of man, and is in equilibrium. Thus, the different gradients shown reflect the influence of different geologic controls on water movement and indicate areas of differing permeability. Along the east and south sides and the north edge of the potentiometric map (fig. 2), and, to some extent to the northwest, gradients steepen rapidly and, in effect, indicate low permeability boundaries. These boundaries, at least along the south side, coincide with thick deposits of clastic rocks as is shown in figure 2.

This example demonstrates, on the basis of local measurements of energy potential of the ground water, how a regional flow system is delineated when considerable subsurface information is available. In this paper, several springs associated with this regional system are classified as regional. These include Fairbanks, Rogers, Devils Hole, Crystal, King, Big, and Bole Springs, all of which are in the vicinity of Ash Meadows.

The Desert Research Institute is presently mapping the ground-water systems in the State of Nevada. As a basis for these studies, all available geologic, hydrologic, and geochemical data are being used, and the first concentrated effort deals with carbonate terrain. Preliminary results show that several regional systems including those described above do exist.

Present knowledge does not permit an accurate mapping of these systems. For example, although Eakin's work implies a possible hydraulic connection for a distance of perhaps 250 miles across several topographic divides, we cannot confidently delineate many hydrologic relationships and geologic controls within this large region. It seems plausible to speculate that as many as three regional systems could be present. On the other hand, we know or have good reason to believe that much more can be done to more accurately delineate and quantify our understanding of the carbonate-rock hydrology.

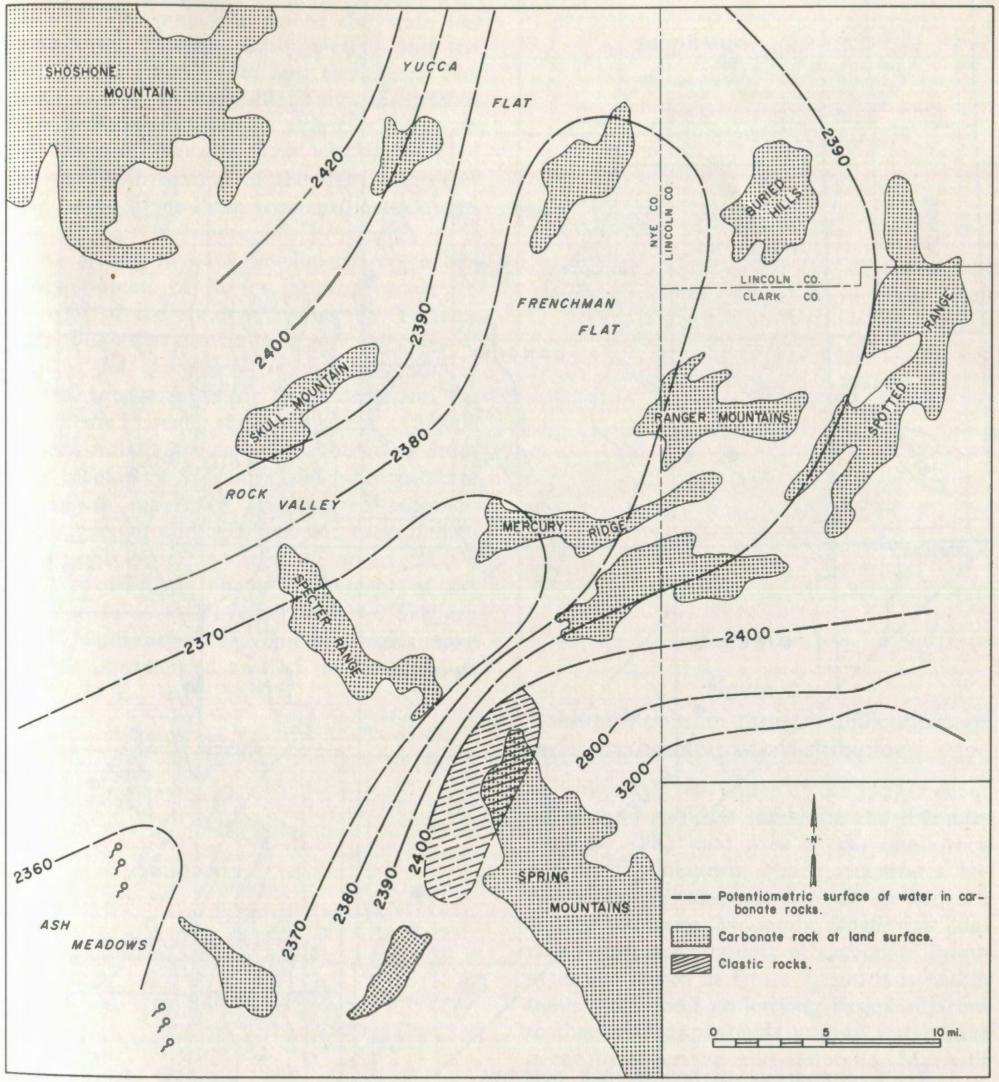

Figure 2.
Regional flow system of the Nevada Test Site.

One approach, presently under study, is careful and accurate mapping of the broad lithologic and major tectonic features that act as hydrologic boundaries. Much of this can be accomplished with field data presently available and with a minimum of new fieldwork. Presently available data may well allow delineation of not only the carbonate terrains but the predominantly clastic and crystalline terrains which Winograd's work has demonstrated to be likely to be major barriers to ground-water motion. Similarly, major tectonic features can be delineated.

Geologic Aspects

The State of Nevada lies almost wholly within the Great Basin section of the Basin and Range Physiographic Province and is characterized by isolated north-trending ranges separated by desert plains of aggra-

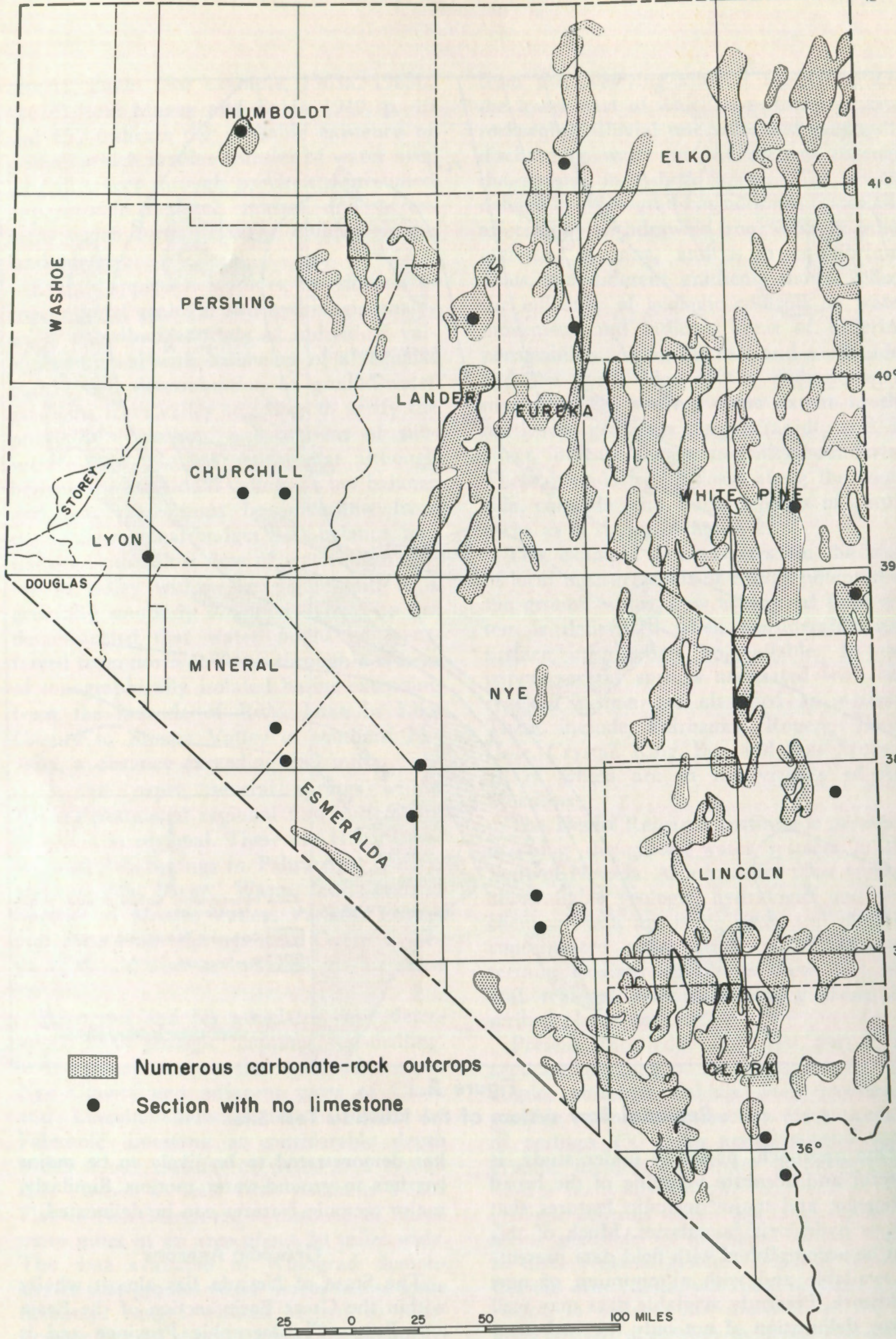

Figure 3.
Areas where carbonate rock is abundant.

dation. The southern tip of the state lies within the Sonoran Basin section and the western boundary cuts into the Sierra Nevada section in the Lake Tahoe area from the latitude of Reno to a few miles south of Minden, Nevada. The eastern part of the northern margin of the state is part of the Snake River Plain section of the Columbia Plateau Physiographic Province.

According to data developed by the Nevada Bureau of Mines, approximately 78 percent of the state is covered by Tertiary and Quaternary sediments and volcanics of which the volcanics cover approximately 33,000 square miles, or 30 percent, and the sediments cover the remainder (53,000 square miles; 48 percent). Protruding from this blanket of volcanics and sediments are numerous ranges of consolidated sedimentary, igneous, and metamorphic rocks, which crop out over an area of about 22,000 square miles, or about 20 percent of the area of the state. Together with the Tertiary and Quaternary rocks, these lithologies make up the unpatterned area of the state shown in figure 3.

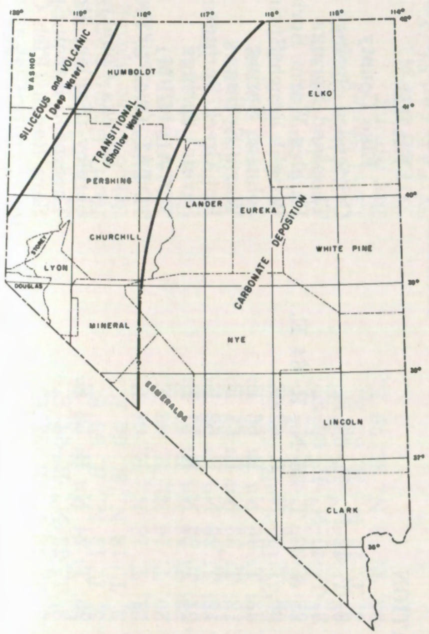

Figure 4.
Depositional environment from Middle Cambrian to Late Devonian time.

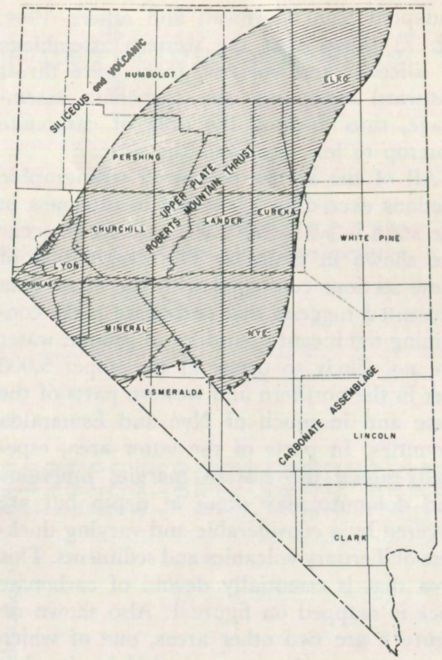

Figure 5.
Roberts Mountain thrust modifications of carbonate-rock distribution.

Within the remainder of the ranges occur all of the significant limestone and dolomite outcrops. The total area of carbonate-rock outcrop is unknown, but it constitutes between one-half and one-third of the bedrock in the mountains. Mountain ranges that contain significant amounts of carbonate lithologies are shown in figure 3, and it is within these ranges and underlying basins adjacent to them that appreciable ground water must occur in limestone and dolomite. Much of the carbonate rock is of Paleozoic age, primarily Devonian and older, although a considerable fraction is of later Paleozoic and Mesozoic age. Most of the Mesozoic carbonate lithologies occur in western Nevada and apparently are only of local importance as ground-water aquifers.

Figure 4 shows the original depositional environment in Nevada during Middle Cambrian to Late Devonian time according to the U. S. Geological Survey (Cornwall and others, 1964, p. 22-23). It clearly shows that carbonate sediments were deposited only in the eastern two-thirds of the state. Figure 5 (adapted from Cornwall and others, 1964, fig. 7) shows that the western assemblage of siliceous and volcanic rocks were thrust eastward overriding the carbonate assemblage, thus limiting the area of carbonate outcrop to less than half the state.

All of the known measured stratigraphic sections exceeding 1,500 feet in thickness in the state in which no carbonate rocks occur are shown in figure 3. The distribution of these sections combined with the other data presented suggests that carbonate rocks containing significant quantities of ground water are not likely to occur in the upper 5,000 feet in the northern and western parts of the state and in much of Nye and Esmeralda Counties. In parts of the latter area, especially along the eastern margin, limestone and dolomite may occur at depth but are covered by a considerable and varying thickness of Tertiary volcanics and sediments. This area that is essentially devoid of carbonate rock is mapped on figure 1. Also shown on figure 1 are two other areas, one of which contains some limestone and dolomite, but in which no regionally important ground-water systems apparently occur, and the remaining area constituting about a third of the state, where large expanses of limestone and dolomite are present both in outcrop and in the subsurface.

Hydrologic Aspects

Since nearly no data are available for wells drilled into the carbonate rocks, consideration of the hydrology must be based primarily on information from springs.

The occurrence and nature of springs issuing from carbonate rocks have been discussed in the introduction. The location of most of the large (discharge in excess of 1 cubic foot per second) carbonate-rocks springs is shown in figure 1 and the springs are listed in table 1. As might be expected from the geology, the large springs occur in the eastern part of the state east of the trace of the Roberts Mountain Thrust (fig. 5). The distribution of the springs clearly indicates that the major hydrologic systems associated with limestone and dolomite occur only in the area of principal outcrop of carbonate rocks and justifies the boundaries previously determined from geologic evidence.

TABLE 1 Carbonate-Rock Springs in Nevada

	LOCATION	NAME	DISCHARGE (GPM)	ANALYSIS	TEMP. (F°)
		Elko County			
1.	Sec. 8, T. 42 N., R. 69 E.	Crittenden Springs	1,000	No	—
2.	Sec. 8, T. 40 N., R. 69 E.	Thousand Springs	1,350	No	—
3.	Sec. 28 & 33, T. 36 N., R. 64 E.	Ralph's Warm Springs	1,193	No	70
4.	Sec. 30, T. 36 N., R. 62 E.	Wright Ranch Spring	450	No	55
5.	Sec. 33, T. 36 N., R. 66 E.	Johnson Spring	2,588	Yes	67
6.	Sec. 12, T. 33 N., R. 61 E.	Warm Springs	2,250	Yes	63
7.	Sec. 12, T. 28 N., R. 52 E.	Hot Creek Springs	5,900	No	84
8.	Sec. 24, T. 27 N., R. 57 E.	Several springs	6,000	No	44 - 60
		TOTAL (GPM)	20,731		
		TOTAL (Ac.Ft./Yr.)	33,400		
		Eureka County			
9.	Sec. 23, T. 24 N., R. 52 E.	Shipley (Sadler) Hot Spring	6,750	Yes	106
10.	Sec. 3, T. 23 N., R. 54 E.	Thompson Ranch Spring	900	No	71 - 75
11.	Sec. 7, T. 16 N., R. 53 E.	Sara Ranch (Fish Creek) Spr.	4,000	No	66
		TOTAL (GPM)	11,650		
		TOTAL (Ac.Ft./Yr.)	18,650		
		White Pine County			
12.	Sec. 16, T. 22 N., R. 63 E.	Borchert John Spring	447	Yes	64
13.	Sec. 24, T. 21 N., R. 63 E.	Monte Neva Spring	630	Yes	174
14.	Sec. 5, T. 19 N., R. 63 E.	North Group Springs	450(?)	Yes	77
15.	Sec. 3, T. 18 N., R. 64 E.	Schoolhouse Spring	450(?)	Yes	76
16.	Sec. 21, T. 18 N., R. 64 E.	McGill Warm Springs	450	Yes	76 - 84
17.	Sec. 20, T. 16 N., R. 63 E.	Murry Springs	3,300	Yes	—
18.	Sec. 33, T. 15 N., R. 57 E.	Green Spring	675	No	63
19.	Sec. 35, T. 14 N., R. 63 E.	Willow Creek Basin Spring	685	Yes	—
20.	Sec. 10, T. 13 N., R. 69 E.	Rowland Spring	1,915	No	48
21.	Sec. 10, T. 13 N., R. 65 E.	Rosebud Spring	22	Yes	—
22.	Sec. 2, T. 12 N., R. 61 E.	Preston Spring	3,900	No	70
23.	Sec. 12, T. 12 N., R. 61 E.	Cold Spring	780	No	70
24.	Sec. 12, T. 12 N., R. 61 E.	Nicholas Spring	1,125	No	71
25.	Sec. 12, T. 12 N., R. 61 E.	Arnoldsen Spring	1,380	No	72
26.	Sec. 4, T. 11 N., R. 62 E.	Lund Spring	2,800	Yes	66
27.	Sec. 33, T. 10 N., R. 70 E.	Big Spring	4,571	No	61
		TOTAL (GPM)	23,580		
		TOTAL (Ac.Ft./Yr.)	38,000		
		Nye County			
28.	Sec. 32, T. 13 N., R. 56 E.	Big Warm Springs (Duckwater)	6,300	No	—
29.	Sec. 25, T. 11 N., R. 58 E.	Currant Creek Springs	2,400	No	—
30.	Sec. 19, T. 9 N., R. 62 E.	Emigrant Springs	1,350	No	67
31.	Sec. 32, T. 9 N., R. 61 E.	Morman Warm Springs	1,900	Yes	100
32.	Sec. 1, T. 8 N., R. 57 E.	Blue Eagle Springs	2,270	Yes	82
33.	Sec. 28, 31, & 32, T. 7 N., R. 62E.	Several springs	2,000	No	65 - 67
34.	Sec. 28, T. 7 N., R. 62 E.	Butterfield Springs	1,125	No	—
35.	Sec. 33, T. 7 N., R. 62 E.	Flag Springs	1,125	No	—
36.	Sec. 25, T. 6 N., R. 60 E.	Moon River Spring	900	No	92
37.	Sec. 18, T. 6 N., R. 61 E.	Hot Creek Springs	6,885	Yes	92
38.	Sec. 9, T. 17 S., R. 50 E.	Fairbanks Spring	1,715	Yes	80
39.	Sec. 15, T. 17 S., R. 50 E.	Rogers Spring	725	Yes	67
40.	Sec. 36, T. 17 S., R. 50 E.	Devil's Hole	—	Yes	92
41.	Sec. 3, T. 18 S., R. 50 E.	Crystal Pool	2,824	Yes	89
42.	Sec. 7, T. 18 S., R. 51 E.	King Spring	1,162	Yes	89
43.	Sec. 19, T. 18 S., R. 51 E.	Big Spring	1,122	Yes	83
44.	Sec. 30, T. 18 S., R. 51 E.	Bole Spring	12	Yes	72
45.	Sec. 14, T. 20 S., R. 53 E.	Bennetts Springs	2,200*†	Yes	77
46.	Sec. 3, T. 21, S., R. 54 E.	Manse Springs	1,500*†	Yes	75
		TOTAL (GPM)	33,815		
		TOTAL (Ac.Ft./Yr.)	54,600		

LOCATION	NAME	DISCHARGE (GPM)	ANALYSIS	TEMP. (F°)
	Lincoln County			
47. Sec. 4, T. 9 N., R. 65 E.	Geyser Spring	538	Yes	68
48. Sec. 35, T. 1 N., R. 66 E.	Floral Spring	—	Yes	—
49. Sec. 4, T. 2 S., R. 68 E.	Panaca Warm Spring	4,883	Yes	85
50. Sec. 22, T. 4 S., R. 60 E.	Hiko Spring	2,400	Yes	80
51. Sec. 10, T. 5 S., R. 60 E.	Crystal Spring	5,300	Yes	82
52. Sec. 36, T. 5 S., R. 60 E.	Ash Spring	7,630	Yes	90
	TOTAL (GPM)	20,751		
	TOTAL (Ac.Ft./Yr.)	33,450		
	Clark County			
53. Sec. 15, T. 14 S., R. 65 E.	Muddy River Springs (Combined)	20,000	Yes	71
54. Sec. 16, T. 14 S., R. 65 E.	Warm Spring	—	Yes	90
55. Sec. 21, T. 14 S., R. 65 E.	Iverson Spring	—	Yes	89
56. Sec. 16, T. 16 S., R. 56 E.	Indian Springs	410	Yes	78
57. Sec. 13, T. 18 S., R. 55 E.	Whiskey & MacFarland Springs	48	Yes	49
58. Sec. 6, T. 18 S., R. 68 E.	Blue Point Spring	—	Yes	—
59. Sec. 12, T. 18 S., R. 67 E.	Rogers Spring	880	Yes	—
60. Sec. 15, T. 19 S., R. 56 E.	McWilliams Spring	34	Yes	44
61. Sec. 31, T. 20 S., R. 56 E.	Intermittent Spring	450 - 15,000†	Yes	57
62. Sec. 30, T. 20 S., R. 61 E.	Las Vegas Springs	1,135*†	Yes	73
	TOTAL (GPM)	21,822		
	TOTAL (Ac.Ft./Yr.)	35,200		
	GRAND TOTAL (GPM)	132,349		
	GRAND TOTAL (Ac.Ft./Yr.)	213,400		

* No longer yields cited discharge because of ground-water withdrawals from adjacent wells.
† Discharge not summed.

TABLE 2
SELECTED ANALYSES OF WATERS DISCHARGED FROM LIMESTONE SPRINGS IN NEVADA
(Constituents in parts per million)

Number	Name	Silica (SiO_2)	Iron (Fe)	Calcium (Ca)	Magnesium (Mg)	Sodium (Na) and Potassium (K)	Carbonate (CO_3)	Bicarbonate (HCO_3)	Sulfate (SO_4)	Chloride (Cl)	Fluoride (F)	Boron (B)	Nitrate (NO_3)	Dissolved solids	Temperature (°F)	Discharge (gpm)	Reference
5	Johnson Spring	11	0	48	15	8.3	0	160	13	11	0.2	--	0	191	67	2588	Eakin and others, 1951, p. 30
6	Warm Springs	35	--	52	20	63	--	334	39	23	1	0.1	0.8	398	63	2250	Eakin and others, 1951, p. 113
9	Shipley Hot Spring	40	0.01	57	21	35	0	279	35	21	.2	.26	0	346	106	6750	Eakin, 1962a, p. 27
12	Borchert John Spring	32	.18	47	19	4.4	0	227	13	3.5	--	--	.34	231	65	450	Clark and Riddell, 1920, p. 43
13	Monte Neva Spring	54	.19	67	21	26	0	324	25	6.6	--	--	.09	349	174	630	Clark and Riddell, 1920, p. 43
14	North Group Springs	32	.05	52	21	15	0	268	20	4.5	--	--	2.1	268	77	450?	Clark and Riddell, 1920, p. 43
15	Schoolhouse Spring	--	tr	--	--	24	6.7	147	27	9.4	--	--	--	240	76	350?	Clark and Riddell, 1920, p. 44
16	McGill Warm Springs	32	.1	54	21	12	0	267	21	4.3	--	--	1.2	266	84	450	Clark and Riddell, 1920, p. 43
17	Murry Springs	--	--	46	20	4.9	--	228	11	2.8	--	--	--	--	low	3300	Miller and others, 1953, p. 48-49
19	Willow Creek Spring	19	.15	49	19	1.2	0	225	13	3.5	--	--	tr	225	low	685	Clark and Riddell, 1920, p. 43
21	Rosebud Spring	22	.17	44	17	7	0	197	22	7.2	--	--	.29	214	low	22	Clark and Riddell, 1920, p. 43
26	Lund Spring	13	--	56	24	3.7	--	276	13	3	--	.02	3.2	252	66	2880	Maxey and Eakin, 1949, p. 47
31	Morman Warm Springs	--	--	63	22	23	0	290	46	8.9	--	.2	--	--	100	1890	Miller and others, 1953, p. 54-55
32	Blue Eagle Springs	17	.6	80	24	30	0	385	34	10	--	--	1.6	590	82	2270	Eakin and others, 1951, p. 153
37	Hot Creek Springs	32	--	58	22	32	--	294	45	12	--	.04	.3	346	92	6885	Maxey and Eakin, 1949, p. 47
38	Fairbanks Spring	--	--	55	18	126	0	367	74	26	--	--	--	--	80	1715	Miller and others, 1953, p. 62-63
39	Rogers Spring	23	.16	50	20	76	0	305	79	23	1.2	.28	.9	447	67	725	Walker and Eakin, 1963, p. 32a
40	Devil's Hole	23	.04	51	21	73	0	311	79	22	1.6	.38	.5	425	92	--	Walker and Eakin, 1963, p. 32a
41	Crystal Pool	--	--	47	21	102	--	310	88	24	--	.1	--	--	89	2824	Miller and others, 1953, p. 62-63
42	King Spring	23	.02	49	21	77	0	310	80	21	1.4	.1	.4	425	89	1162	Walker and Eakin, 1963, p. 32a
43	Big Spring	32	.11	45	18	107	0	314	110	25	1.4	.08	.3	468	83	1122	Walker and Eakin, 1963, p. 32a
44	Bole Spring	33	.03	38	19	115	0	306	113	27	1	--	1	500	72	12	Walker and Eakin, 1963, p. 32a
45	Bennetts Springs	--	--	50	22	8.2	--	244	33	.7	--	--	--	358	76	2520	Maxey and Jameson, 1948, app. 2, p. 43
46	Manse Springs	18	tr	55	29	tr	--	239	42	4.9	--	--	0	268	75	605	Maxey and Jameson, 1948, app. 2, p. 43
47	Geyser Spring	13	--	30	3.4	4	0	103	5	3	0	0	.6	115	68	538	Rush and Eakin, 1963, p. 17
48	Floral Spring	15	--	69	34	13	7	350	5	7	--	--	--	322	low	--	Phoenix, 1948, p. 88
49	Panaca Warm Spring	51	0	31	9.8	44.8	--	189	29	15	1.6	.1	2.6	118	85	4883	Rush, 1964, p. 27
50	Hiko Spring	33	--	44	23	44.6	0	260	36	11	.03	.01	1.24	--	80	2405	Eakin, 1963a, p. 24a
51	Crystal Spring	31	--	46	22.1	36.6	0	242	34	9.9	.15	.01	.6	277	82	5314	Eakin, 1963a, p. 24a
52	Ash Spring	31	--	39	18.1	48.4	0	231	34	9.6	.03	.1	1.24	--	88	7630	Eakin, 1963a, p. 24a
53	Muddy River Sprs. (combined)	32	--	71	33	139	0	303	216	75	2.4	.4	1.5	--	71	20925	Eakin, 1964, p. 28
54	Warm Spring	31	--	65	28	109	0	288	174	60	2.4	.3	2.3	614	90	--	Eakin, 1964, p. 28
55	Iverson Spring	29	--	70	26	112	0	274	179	64	2.3	.3	2.2	620	89	--	Eakin, 1964, p. 28
56	Indian Springs	17	.16	48	15	31	--	239	28	5	--	--	0	330	78	400	Maxey and Jameson, 1948, app. 2, p. 42
57	Whiskey & McFarland Springs	--	.05	62	13	3	--	249	6	3	--	--	--	230	49	48	This report
58	Blue Point Spring	--	--	482	169	428	--	122	1910	368	--	--	--	--	--	--	Miller and others, 1953, p. 58-59
59	Rogers Spring	--	--	451	149	395	--	185	1670	343	--	--	--	--	--	--	Miller and others, 1953, p. 58-59
60	McWilliams Spring	5	--	61	15	5	--	286	3.5	2	--	--	--	221	43.5	34	This report
61	Intermittent Spring	10	0	59	24	4.3	--	273	22	5	--	--	1.5	251	57	table 1	Maxey and Jameson, 1948, app. 2, p. 43
62	Las Vegas Springs	13	tr	56	23	17	--	239	43	2	--	--	6	267	73	1135	Maxey and Jameson, 1948, app. 2, p. 42

Spring Discharge—Magnitude of discharge of the springs varies considerably. The largest discharge (7,650 gallons a minute) for a single spring in available records is that of Ash Spring in Pahranagat Valley near Alamo, Nevada, and the largest discharge (20,000± gallons a minute) for a group of closely associated springs is that of the springs near Moapa, Nevada, which are the principal source of the Muddy River. Of the approximately 60 springs considered in this paper, 10 discharge 5,000 gallons a minute or more, 16 discharge 2,000 to 5,000 gallons a minute, 15 discharge 1,000 to 2,000 gallons a minute, and 15 discharge 450 to 1,000 gallons a minute. Springs with a discharge of less than 450 gallons a minute are generally not mentioned in this report but are numerous and are of importance in eastern Nevada as a source of domestic and stock water. Many are associated with local flow systems, perched water tables, and often are intermittent.

Some of the large springs maintain a relatively steady flow both seasonally and over long periods of time, whereas with others the discharge fluctuates considerably, especially on a seasonal basis, and some discharges fluctuate even as a result of local storms. Available records are not complete enough to accurately indicate discharge fluctuations for all of the springs either on a seasonal or long-term basis.

Spring Temperatures— Temperatures also vary from spring to spring, and of those considered, the range is from close to the

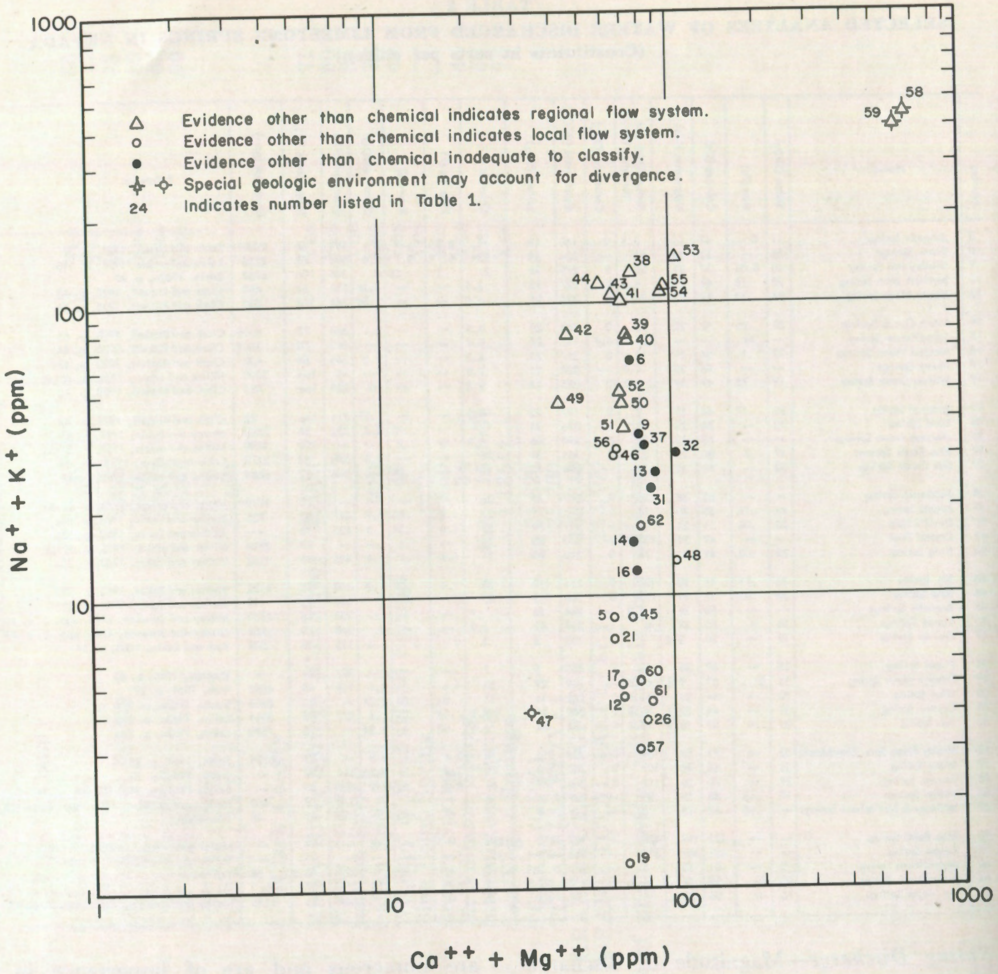

Figure 6.

Sodium and potassium versus calcium and magnesium in springs associated with carbonate rocks.

average annual air temperature to as high as 174°F. Adequate temperature records are not available for detailed analysis, but certain trends in relations between discharge and temperature are indicated. For example, the springs with relatively small discharge variation nearly all have relatively large discharges and high temperatures; whereas those with large or suspectedly large discharge variation have relatively low temperatures. In general, springs discharging water with lower temperatures are those associated with local systems, whereas springs discharging higher temperature water seem to be associated with regional systems.

Water Chemistry—Forty available chemical analyses of spring water are given in table 2. As is indicated in the table, these analyses were made by a number of different organizations over a period exceeding 40 years. Thus the variation in analytical accuracy is probably quite large. Also, many of the analyses were made for some specific objective and are not complete enough for

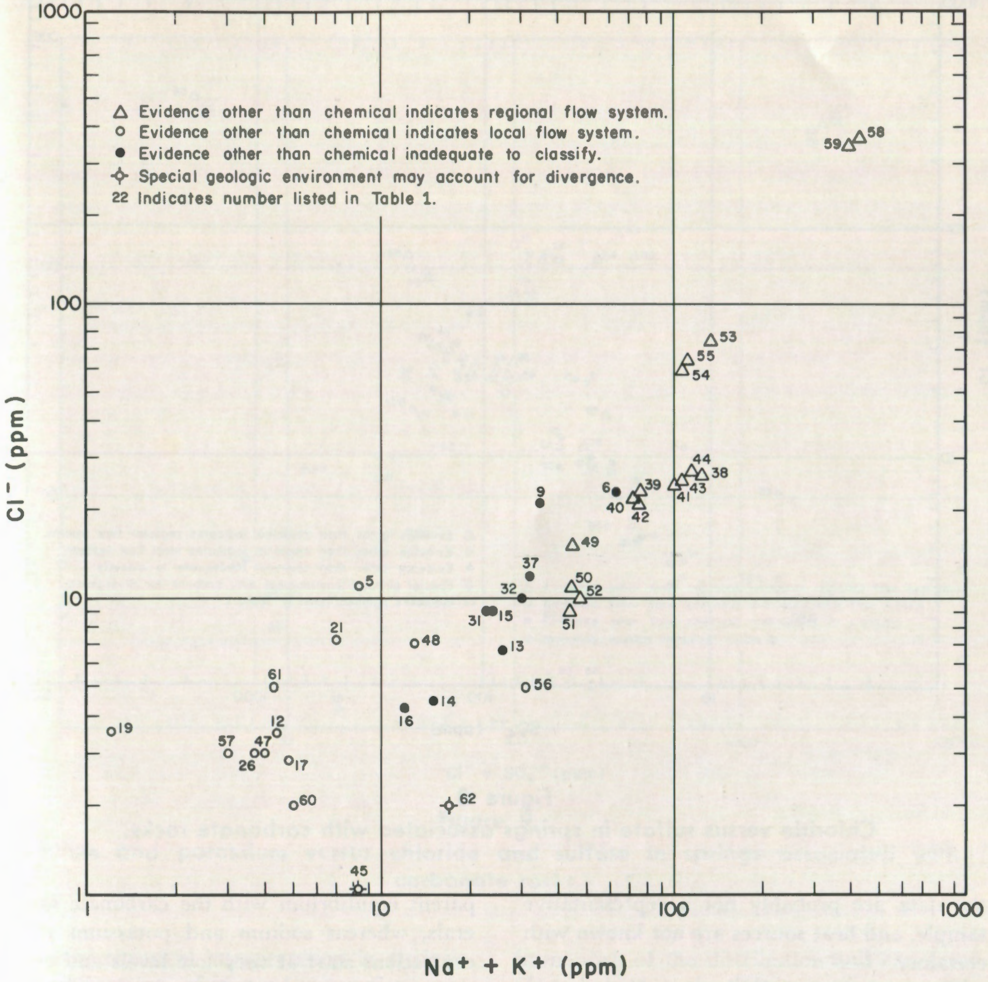

Figure 7.

Chloride versus sodium and potassium in springs associated with carbonate rocks.

detailed study and interpretation for other objectives. However, concentrations of several major constituents are commonly shown and of these sodium and potassium, calcium, magnesium, chloride, and sulfate ions have proved useful in this study. When more complete and reliable analyses are available, other constituents may also prove to be useful in the study of the flow systems.

In spite of their limitations, the analyses do show that nearly all of the waters are of the calcium-magnesium bicarbonate type. Further, high concentrations (in excess of about 30 parts per million) of sodium and potassium and chloride seem to indicate special hydrogeologic environments. This is also probably true of sulfate.

Comparison of concentration of chemical constituents with temperature and discharge of spring waters yields little correlation, although there is some indication that concentration of silica and total dissolved solids increases with higher temperature. Even this tentative generalization is questionable since

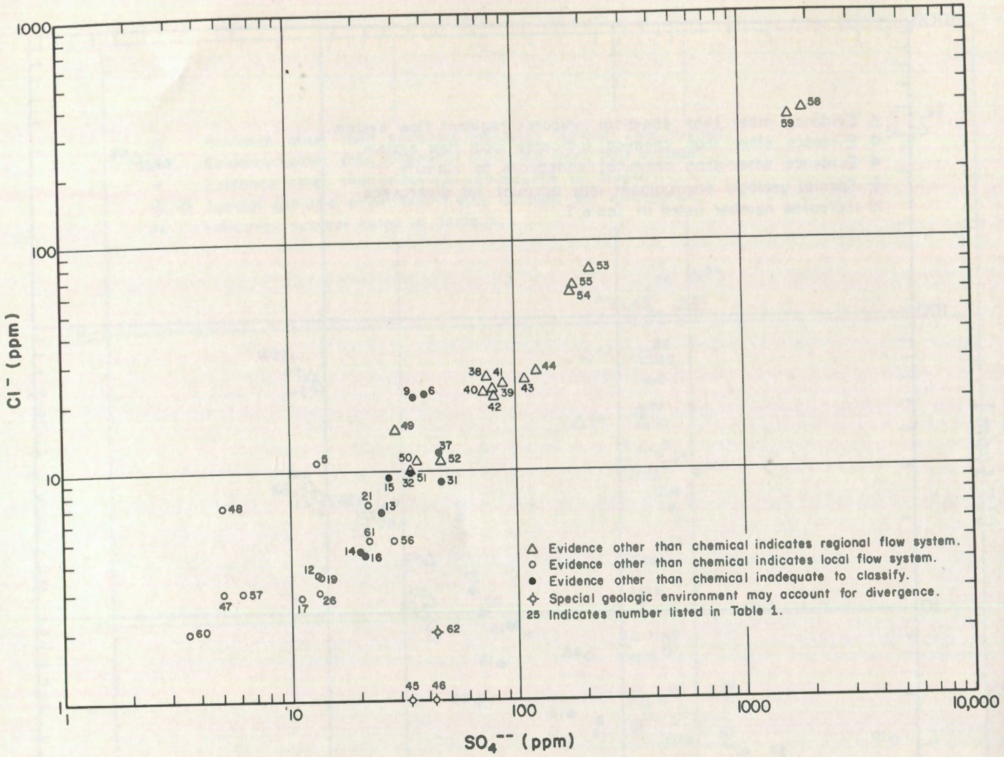

Figure 8.

Chloride versus sulfate in springs associated with carbonate rocks.

the data are probably not a representative sample, and heat sources are not known with certainty.

Figures 6 through 9 illustrate, by logarithmic plots of various chemical constituents in approximately 40 carbonate-rock springs, the more-or-less linear increase of concentration of sodium and potassium, sulfate, and chloride ions with apparent increase in flow path length as determined by evidence other than chemical, that is, by geologic, hydrologic, and physiographic considerations. Thus, these spring waters may be chemically classified as being part of local or regional flow systems as is shown on the graphs and in figure 1.

Figure 6 illustrates how the calcium and magnesium ions rapidly reach a relatively constant concentration, and therefore apparent equilibrium with the carbonate minerals, whereas sodium and potassium concentrations start at very low levels and continuously increase with distance, even though time and various chemical-exchange factors also may be important. Figure 7 illustrates a definite linearity when concentrations of chloride and sodium and potassium are plotted. Further, the proportional increases of these ions with length of flow is clearly shown. Figure 8 shows the same linearity and proportional increase in ion concentration when chloride concentration is plotted with sulfate concentration. Similarly, when the data given in figures 7 and 8 are combined as is shown in figure 9, a comparable increased concentration of the ions with distance is apparent. Thus, it is evident that this method of comparison demonstrates the

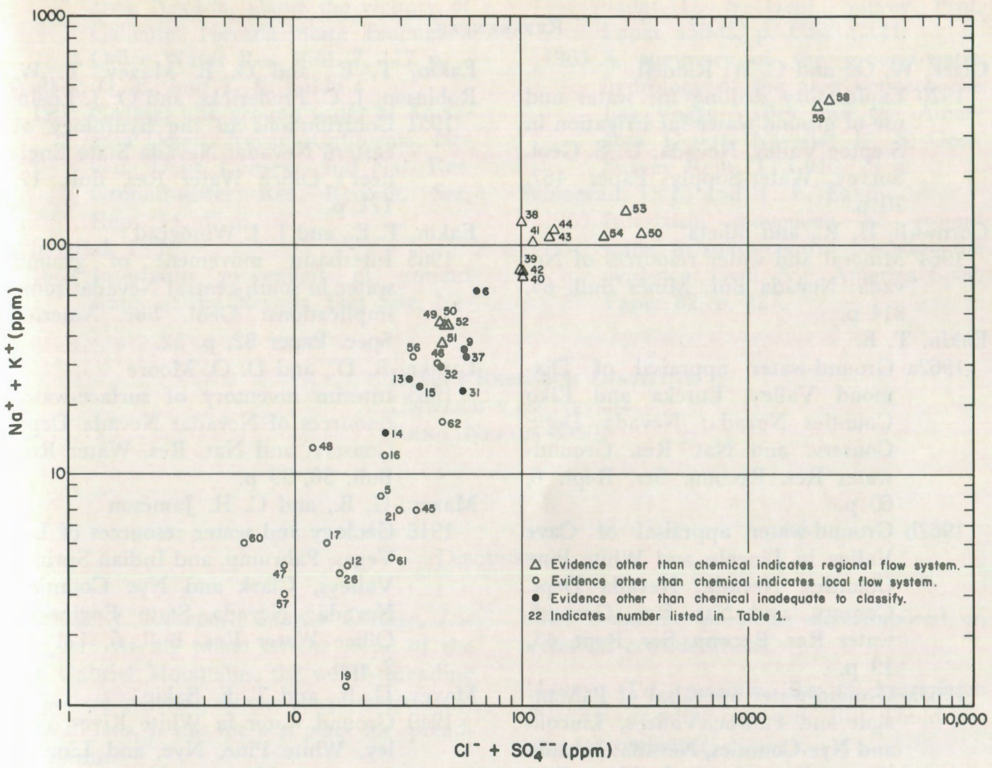

Figure 9.

Sodium and potassium versus chloride and sulfate in springs associated with carbonate rocks.

utility of chemical interpretation in analysis and delineation of ground-water flow systems in carbonate-rock terrains.

With increasing availability of accurate and economically produced analyses, chemical interpretation may be the most valuable qualitative tool in delineation of flow systems. The methods illustrated here represent only an initial phase of a continuing study that, when finished, will likely produce a far more useful and sophisticated methodology.

CONCLUSIONS

When attempts are made to understand the occurrence and movement of ground water in carbonate terrain in areas such as Nevada, direct and classical field observations contributed by the geologist and speleologist can only approximate the physical framework of the distribution and character of carbonate rocks. The actual cause-and-effect relationships the lithologies impart to the hydrology of the terrain must be tested and determined primarily by the various measurable parameters of the water, either at the land surface or in the subsurface. Of these parameters, in our present state of knowledge, the most useful and successful are the energy potentials of the water in accordance with accepted hydrodynamic principles, the water budgets, and the chemical characteristics of the water. Although at present all of these parameters must be somewhat approximated in practical collection and analysis, they establish the theoretical foundation upon which ultimate knowledge of flow systems in carbonate terrains depends.

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Discussion

Richard J. Reardon, *Sawyer College, Los Angeles*: At my home on the slope of the San Gabriel Mountains, the water-spreading basins are generally at the lower ends of the alluvial fans. Is this the best place for spreading basins?

Maxey: I would prefer to see spreading basins high on the alluvial fans where you would expect the highest permeability.

Thomas Aley, *U. S. Forest Service, Winona, Missouri*: In the hope of clarifying the last comment, on the neighboring watershed of the Santa Ana River, the spreading basins are up near the mountains, high on the alluvial fans. Permeabilities there are a number of times greater than down in valleys. In Southern California there are a number of other economic problems superimposed on technical considerations.

Maxey: I'd certainly like to hear from chemists regarding our work. Bill, would you care to comment?

William Back, *U. S. Geological Survey, Washington, D.C.*: Burke, what you've done here is to take a first step in applying the concept of hydrochemical facies that I used a few years ago and which I think we can fairly say has swept the country with a great surge of apathy. The next step would be to show the distribution of these determinations both areally and vertically. This would show the effect of the mineralogy and flow patterns on the ratios that you find. It is certainly what has to be done in many areas before we can get on to the geochemistry.

Central Kentucky Karst Hydrology

By R. A. Watson

ABSTRACT

Three aspects of underground flow in the Central Kentucky Karst are considered. First, the underground drainage pattern is examined as a set of systems tributary to the Green River. The relation of inputs such as vertical shafts to horizontal passages is delineated, and the extent to which structure determines passage alignment is evaluated. Calculations based on rainfall and spring-flow measurements indicate that a substantial portion of the input cannot be accounted for by the discharge of presently known springs and underground streams. Second, the quantity and quality of water flowing underground is examined, and new evaluations are made of the location, character, and extent of limestone solution in the area. Known and extrapolated cave passages cannot account for the amount of limestone removal indicated. Third, the underground drainage is examined as a set of transportation systems for mechanical sediment. Estimates are made of the amount and character of sediment moving from the surface of the region underground to the Green River. This sediment far exceeds the amount brought into the systems by back-up water during floods of the Green River.

It is concluded that the active underground drainage systems are more complex than their known parts indicate, with major water channels probably existing beneath the present-day base level; that a major amount of solutional activity occurs in the mantle and on the bedrock surface below it causing a general lowering of the land surface through removal of limestone which far exceeds that removed during the formation of cave passages; and that most of the sediment in the cave passages, even in back-up areas near the Green River, is provided by the flow which moves from the surface through the cave systems to the Green River.

INTRODUCTION

This paper consists of summary statements on three general aspects of underground flow in the Central Kentucky Karst: (1) the underground drainage pattern as a set of systems tributary to the Green River; (2) the solutional activity of the underground water; and (3) the underground drainage as a set of transportation systems for mechanical sediment. My conclusions are based on 10 years of observation and study, and upon many discussions about their extensive work with R. W. Brucker, G. H. Deike, E. R. Pohl, and W. B. White, all of whom might be listed here as joint-authors except that none of them can be held responsible for my speculative opinions. E. R. Pohl was especially helpful with part 3. Institutional assistance was provided by Mammoth Cave National Park and the Cave Research Foundation.

The Central Kentucky Karst lies on Mississippian rocks dipping 30 feet to the mile northwest in south-central Kentucky, along the southwestern edge of the Cincinnati Arch in the Interior Lowland Province, its center 100 miles south of Louisville (Pohl and others, 1964). It is part of a karsted limestone belt which extends from southern Indiana south through Kentucky into Tennessee along the entire west flank of the Cincinnati Arch. The boundaries of the Central Kentucky Karst are drawn on a hydrologic basis, resulting in an oblong area, at a minimum 20 miles from east to west, and 10 miles from north to south, with Mammoth Cave National Park near its center. The northern boundary is the drainage divide between the Green and Rough Rivers, north of which no recharge into the area is possible. The southern boundary is the drainage divide between the Green and Barron Rivers. To the east, the boundary is near Munfordville beyond which the Green River flows on limestone from which the important caprock of the area has long been removed. The western boundary is near Brownsville where the Green River is located stratigraphically above the caprock so that discharge of underground water into the river is not possible. The Central Kentucky Karst consists of three major physiographic units separated by two major escarpments. From southeast to northwest, these are the Sinkhole Plain, the Dripping Springs Escarpment, the Mammoth Cave Plateau, the Pottsville Escarpment, and the Hilly Country. These have been described by Livesay and McGrain (1962) and by Watson and Smith (1963).

The Underground Drainage Pattern as a Set of Systems Tributary to the Green River

There are both perennial and ephemeral underground streams in the Central Kentucky Karst. We know most about the ephemeral flows. After heavy rains and during wet seasons, the amount of vertical flow in the cave systems increases, water seeping or dripping from joints, and cascading down vertical shafts. The inputs for this water are relatively open to the surface, flow increasing in some places within half an hour after a heavy rainfall starts. Ephemeral sinking creeks also start up during rainy times, and when rain continues for several days, ephemeral ponds form on the Sinkhole Plain and on the floors of the karst valleys of the Mammoth Cave Plateau. The underground streams formed by water flowing through the open inputs and from the slowly draining ponds flow only so long as direct meteoric water is available.

The second sort of underground flow is found in perennial streams such as Hidden River, Cave City Water Supply River, Eyeless Fish River, Candlelight River, Roaring River, and so on. The ephemeral streams are tributary to these perennial streams, and during wet periods increase their flow greatly. However, each of these streams flows at a few cfs (cubic feet per second) year around, obtaining its water from no known reservoir. Presumably they are fed by the constant inseep of subsurface water, though no stream can be followed far enough upstream to gather a direct notion of the total collecting mechanism. However, Brucker (1960; in press) has demonstrated that the vertical shaft complexes beneath the edges of the truncated caprock of the Mammoth Cave Plateau are headwater tributaries of such streams as Eyeless Fish River. Many of these shaft complexes supply water which seeps down their walls even during the driest parts of the year. White and Deike (1963; unpublished) in their studies of the paleohydrology of the region have found evidence to support the notion that some extensive high-level horizontal passages which are now perennially dry were once stream beds related to former vertical shafts which now are collapsed and eroded. It might be concluded that present perennial streams are fed entirely by water which perennially seeps and drips down vertical shaft complexes. That known complexes do not provide enough water is not conclusive evidence against this hypothesis, since it is difficult to get into some complexes, and many must exist unaccessible to man. Further, it might be concluded that the extant stream systems are models upon which to explain the genesis of high-level dry passages. Brucker (1960; in preparation) gives good evidence for these conclusions.

However, there are numerous discordant elements which do not fit this hypothesis in any simple way. One of the most obvious is that much more water falls onto the Central Kentucky Karst, considered at the very minimum as a 200 square mile drainage area with 48 inches of annual rainfall, than discharges from the known surface and spring tributaries to the Green River in the area. Estimating an annual total discharge from all these sources of 100 cfs (which is probably high) 7 times more water falls on the area than discharges from it through these sources. If one considers the 444 square miles of area undrained by surface streams (Cushman and others, 1965), then more than 17 times more water falls on it than is known to discharge from it. Of course these figures may mean little; an immense amount of work remains to be done on the water regime of the region. During heavy rains much

water flows directly into the Green River; there probably are unknown rises in the Green River bed; and some ground water probably seeps out of the region. I present the calculations primarily as basis for the following speculation: The underground streams now known in the Central Kentucky Karst are probably tributary to much larger underground master trunk streams which flow perenially at or just below present regional base level. Only one such master trunk stream can be explored for any distance under the Mammoth Cave Plateau. This river, in a cave which opens out of Cedar Sink on Joppa Ridge, has a flow of 20 cfs, and can be followed for about 300 feet. That these inferred master streams are not now open to exploration because they are below present base level can be explained if two things are taken into consideration. First, the bed of the Green River has been filled with Pleistocene gravel to a depth in places of 30 feet, and the underground trunk drainage systems were probably established prior to this filling. Second, the Brownsville Dam has raised the level of Green River at least six feet through the Central Kentucky Karst. (If it were not for this artificial raising of the water level, there would be no boat trips inside Mammoth Cave. The large body of water with sections known as Styx River, Echo River, and Roaring River is not a master stream, but is backponded water from the Green River. Upstream underground above the backponding, flow is only a few tenths of a cubic foot per second.) The mouths of the big springs discharging into the Green River are filled with silt, and scuba diving has shown that the floor of the underground stream discharging at Pike Spring is about 30 feet below the present bed of the Green River at that point (Smith, 1957). If Pike Spring is considered as an outlet of a major underground trunk stream, further evidence can be given. The only known underground stream in the area is Eyeless Fish River, which can be followed underground to within 300 feet of the Pike Spring outlet, and is undoubtedly tributary to it (Smith, 1964, fig. 1). However, Eyeless Fish River flows at only 1.5 cfs, which is only a tenth of the flow out of Pike Spring. The master stream must have many other sources of water than Eyeless Fish River, and it must flow below present base level.

Recent exploration under the Mammoth Cave Plateau suggests a further possibility. Extensive work in the Candlelight River complex of small canyons under Houchins Valley, the large karst valley separating Flint Ridge from Mammoth Cave Ridge, indicates that these passages are drains for water from the ephemeral sinking creeks of Houchins Valley. They probably also drain vertical-shaft complexes in Mammoth Cave Ridge. No large passages have been found below the valley, but of course they may be below base level. In general in this area as much as 1 cfs of water perennially flows down-dip to the northwest toward Flint Ridge. The total year-around sources of this water are unknown. Other exploration from Mammoth Cave Ridge up Roaring River has penetrated beneath Doyel Valley toward Joppa Ridge, with water flowing from that area northwest toward Mammoth Cave Ridge. It appears that the floors of the solutional karst valleys separating the major ridges of the Mammoth Cave Plateau represent headwater highlands in the underground systems. A karst-valley floor with its sinking creeks, and with vertical-shaft complexes along each of its walls, is thus a highland up dip from the major ridge north of it. The impermeable caprock of the ridges keeps water from flowing vertically into the passages under each ridge, but water flows in general down dip from the uncapped shaft complexes and karst-valley headlands to master streams below the capped ridges. I would suggest that there is at least one master underground stream below each of the three major ridges of the Mammoth Cave Plateau. This conclusion comes from the observation that Doyel Valley south of Mammoth Cave Ridge and Houchins Valley south of Flint Ridge both are sites of small perennial underground streams, and that these karst valleys are obviously headland collection basins for water which flows into them from the surface of the ridges and below ground from the surface of the impermeable caprock of the ridges.

There is further evidence that these inferred major underground trunk streams extend headward 10 miles or more beneath the

Mammoth Cave Plateau out under the Sinkhole Plain. Such a larger drainage area would better account for the suspected large perennial flow. Very extensive drainage systems are indicated by some of the present high-level passages which are several miles long and perhaps once extended farther out into the Sinkhole Plain. The configuration of some of these passages also suggests that they were formed in circumstances where the stream flowing through them filled them completely rather than simply flowing along their floors. It is probable that they formed just below a formerly higher base level. Further details are given by White and Deike (1963; unpublished) and Brucker (in press). Deike is completing a doctorial dissertation in which he considers the relations of linear features such as joint patterns to cave passages in the Central Kentucky Karst, and when this is available more can be said about the drainage nets.

The Solutional Activity of Underground Water

The second major problem I want to consider very briefly in this paper is the matter of underground solution in the Central Kentucky Karst. Many estimates have been made of how much rock is removed in limestone regions by solution. Let me quote one of the more conservative:

> If each year 14,000,000 cubic feet of water [15 percent of 40 inches annual rainfall] are available for solution beneath each square mile of the Kentucky blue-grass region, and if each million parts dissolves 155 parts (volume) of rock materials, then in a single year 2,170 cubic feet of rock are taken into solution. If all of the rock excavation of a single year were concentrated in one tunnel 3 by 6 feet in cross-section, a cavern over 120 feet long would be formed under each square mile of surface. (Swinnerton, 1932, p. 679).

Swinnerton goes on to say that the Bluegrass region has been stable for several hundred thousand years. The obvious thing I want to point out is that neither in the Bluegrass region nor in the Central Kentucky Karst are there cave passages or other solutional voids underground (as estimated from known and extrapolated cave passages and solutional voids observed in quarry excavations) which even begin to represent the volume indicated by Swinnerton's conservative estimate. Yet, I think it obvious that such solution takes place. The voids exist in the Central Kentucky Karst in the form of the major karst valleys on the Mammoth Cave Plateau and the areas where an immense amount of rock has been removed above the present surface of the Sinkhole Plain. Solution of limestone to form sinkholes accounts for some of the removal, but even more must be accounted for by solution in the mantle and on the bedrock surface. As I say, this is usually pointed out as an obvious fact. It has not yet, however, been considered in all its implications by those who theorize about the genesis of karst forms. The major amount of solutional work in the Central Kentucky Karst must be extensive subsurface solutional activity in the mantle and on the limestone bedrock surface. Incidentally, water seeping slowly through the soil and mantle is undoubtedly the chief source of water for those underground rivers with perennial flow. Once this view is taken of the karst region, one can see the dry cave systems as merely the remnant roots of a once dynamic organism, the vast upper bulk of which has been reduced by solution and removed by transportation through the lower drainage system. Again, the known and extrapolated drainage nets active today are neither the largest parts of the dynamic system, nor the sites of the greatest solutional activity. The active cave systems today in the Central Kentucky Karst are primarily the drainage lines for water which does its major work in the massive karst processes of solution in the mantle and on the bedrock surfaces above the cave passages.

The Underground Drainage as a Set of Transportation Systems for Mechanical Sediment

The picture I have drawn so far is of several drainage nets which extend from the mouths of perennially flowing big springs at the Green River, back beneath the Mammoth Cave Plateau as major subsurface streams fed by tributaries from karst valleys and vertical-shaft complexes, and on out under the Sinkhole Plain where water enters through sinking creeks and sinkholes, and where water seeps through the mantle at the outermost limit of the drainage nets. With this picture in mind, I turn to the third problem to be considered in this paper, the underground transportation of mechanical sediment in the Central Kentucky Karst.

Collier and Flint (1964) conclude on the basis of measurements between October 1959 and June 1962

> that sedimentation in [Mammoth] cave is closely related to flooding of nearby Green River. The Green River, which is hydraulically connected to Mammoth Cave by Echo River spring and River Styx spring, is the chief source of sediment and floodwater to the cave.

I shall argue that these conclusions are incorrect in their general implications and in detail.

The Green River is a master through-flowing surface karst stream. It flows 55 miles from Munfordville to Brownsville through the Central Kentucky Karst with no perennial surface tributary. Mammoth Cave lies about two-thirds of the distance downstream on this course. The Green River has a mean annual discharge of 3300 cfs at its junction with the Nolin River 15 miles below the spring outlets of Mammoth Cave (U. S. Geol. Survey, Lexington, Kentucky open files). In the Mammoth Cave region it drains a total area of approximately 300 square miles, almost completely by way of underground drainage systems. The mean annual discharge of the Green River at Mammoth Cave is 2000 cfs (Cushman, and others, 1965). The most distant extension of this drainage system at right angles from the Green River is approximately 12 miles, with underground headwater drainage lines there lying at a level as much as 200 feet above the pond stage (425 feet at the Brownsville dam) of the Green River.

Drainage into this underground network comes from five sources: (1) Water moves underground on the Sinkhole Plain through sinking creeks. (2) It similarly enters on the plain through sinkholes. The water from these first two sources never travels less than five miles from the Sinkhole Plain beneath the Mammoth Cave Plateau to the river. (3) Some of the precipitation on the ridge tops of the Mammoth Cave Plateau goes underground to flow along the top of the impermeable caprock of sandstone and shale. Where this caprock is truncated by the large solutional valleys between the ridges, and by the valley of the Green River, water descends rapidly to base level through vertical features of solution in the underlying limestone; it then discharges horizontally into the Green River. (4) Between the ridges of the Mammoth Cave Plateau all precipitation in the solutional karst valleys, which are closed depressions, and all water which flows into them, enters the underground drainage network through swallow holes and sinks on the valley floors. (5) Water draining from the Hilly Country on the north side of the Green River generally goes underground through vertical courses.

I am primarily concerned, however, only with the first four sources which provide water from the south side of the Green River, where lies the Mammoth Cave System.

The outlets of this water are five known big springs — Blue, Hix, Pike, Styx River, Echo River — flowing into the Green River at river level; and two known big springs— Turnhole and one unnamed — which rise from the river bottom. These big springs have an estimated mean flow of approximately 100 cfs. As already remarked, it appears probable that there are unknown rises in the bed of the Green River.

The Mammoth Cave Region provides no exception to the observation that the lowering of a karst surface is relatively rapid, and often spectacular. Swinnerton's estimate of the extensive amount of limestone probably removed in solution has already been quoted. Processes of mechanical erosion are no less apparent. From the Mammoth Cave Plateau, debris of sizes from clay to five-foot boulders of sandstone are washed into the drainage systems through vertical openings which may be only joints widened a fraction of an inch, or vertical shafts 50 feet in diameter. On the Sinkhole Plain, the ravages of soil erosion are often apparent. Here, with sporadic forest cover and intensive agriculture, great quantities of silt and clay are washed into the underground drainage systems each year.

Besides mechanical removal of this material, chemical analyses (U. S. Geol. Survey, Lexington, Ky., open files; Cushman and others, 1965) show spring discharges with over 800 ppm dissolved solids, indicating extensive breakdown and removal of material.

Boulders, cobbles, pebbles, sand, silt, and clay are found throughout the vast underground drainage systems. Indeed, many of the headwater channels are nearly closed by the influx of silt. Many sinkholes on the Sinkhole Plain (where there are as many as a thousand sinkholes to the square mile) support ephemeral ponds when their outlets become plugged with clay. Clay and silt pan fills are known up to 30 feet deep, and some are no doubt deeper. When a plug is breached, the movement of clay and silt through the new opening is often phenomenal. Soil erosion is the greatest problem of any karst region. In Eastern Europe and the Near East there are now barren regions of exposed limestone which have been within historic times green and forest laden. Soil erosion is a problem in the Mammoth Cave Region, and it should be a warning sign that recently in some isolated areas soil has been removed to bedrock and karren has begun to appear.

Therefore, it must be concluded that the chief source of sediment in Mammoth Cave, which is known to extend at least four miles back from the Green River and more than 100 feet above its highest flood level, is not the Green River, but is the drainage network extending underground into the Mammoth Cave Plateau and the Sinkhole Plain. This conclusion is based on extensive observations over 30 years throughout more than 100 miles of underground passages in the Mammoth Cave region. The distribution of fluvial sediment in the cave systems is consistent only with the conclusion that these underground passages are part of an immense drainage net as described in the first two parts of this paper.

Thus, though I agree that "sedimentation in the cave is closely related to flooding of nearby Green River", I would argue *not* that "the Green River . . . is the chief source of sediment and floodwater to the cave", but, on the contrary, that *the cave systems are the chief* LOCAL *sources of sediment and floodwater to the Green River.*

The flooding of any river is resultant upon the inflow of water from its tributaries. Over most of the Central Kentucky Karst, water from rainfall and snow-melt during the usual flood times can move only underground. At this time, water levels throughout the cave systems rise rapidly. Where only trickles of water fall from vertical shaft ceilings or flow along horizontal passages during most of the year, streams of water flowing as much as 5 cfs appear. Four miles from the Green River on the Sinkhole Plain, Hidden River in Hidden River Cave, 40 feet above the Green River pond stage, which normally flows at 4 to 5 cfs, has been observed to flood to a maximum of 60 feet above its usual level. This is more than 50 feet higher than the highest recorded floods of the Green River and the areas of Mammoth Cave adjacent to it.

The conclusion must be that the rise of water level in the caves is general to the region, subsequent upon the general rise of the underground water level in wet seasons. The rise of water level in Mammoth Cave during flood time is as much a result of the rise of water in the total set of underground drainage systems tributary to the Green River in the Central Kentucky Karst, as it is the result of the rise of the Green River fed by sources outside the local area. The floodwater of the Green itself must come in part from the high underground water of the Mammoth Cave Plateau and Sinkhole Plain drainage basin. This conclusion is at least in part implied by Davies and Chao (1959), Hendrickson (1961), and Cushing and others (1965).

In Hidden River Cave, floodwaters some years leave a deposit of as much as 10 inches of silt. Yet, the cave is not plugged with silt, nor is Mammoth Cave plugged with silt. The total year-around result is removal of sediment from the cave systems. The essential process is one of transportation, not deposition. The caves are simply the ever-evolving underground passages through which water and debris are moved from the drainage basin to the Green River. Thus, though I do not wish to deny the fact that near the Green River during time of flood some backwater enters the cave systems and deposits some sediment, I do wish to argue that such flooding and deposition is not of major significance, even for the large quantities of water and silt found in the caves near the Green River during the flood. The major influx of water and sediment in the caves is from the Mammoth Cave Plateau and the Sinkhole Plain, not from the Green River.

Two final items will conclude this discussion. First, since "the sediments in the lower levels of the cave are similar in physical character and mineralogy to the sediments on the flood plain of the Green River" (Collier and Flint, 1964, p. D141), they obviously come from the same source. It is my conclusion that they do not come from outside the region to be deposited on the extended underground floodplain of the Green River in the lower levels of Mammoth Cave. Rather, the sediments in the cave and on the flood plain are similar because they both derive from the immediately adjacent Mammoth Cave Plateau and Sinkhole Plain.

Second, it seems enigmatic that water enters the cave systems rapidly while the Green River rises rapidly, but that it does not leave the systems as rapidly as the river lowers. Rather than postulating some intricate valve-system of passages, I conclude that this circumstance is another confirmation of the conclusion that the major source of underground water in the caves during flood is not the Green River. The water lowers less rapidly than it rises in the cave near the Green River because inflow from the extensive underground drainage net continues to be high for several days after the Green River begins to lower. This is because of the more restricted underground flow. Thus, the total rise of the Green River, which is only partially dependent upon local sources, causes some backflooding into the cave systems. But the high level of the underground water which lasts for a few days after the Green itself begins to lower is not the result of capture of backwater, but of the continued high flow of water into the system along underground courses.

Conclusion

It is concluded that the active underground drainage systems in the Central Kentucky Karst are more complex than their known parts indicate with major water channels probably existing beneath the present-day base level; that a major amount of solutional activity occurs in the mantle and on the bedrock surface below it causing a general lowering of the land surface through removal of limestone far exceeding that removed during the formation of cave passages; and that most of the sediment in the cave passages, even in back-up areas near the Green River, is provided by the flow which moves from the surface through the cave systems to the Green River.

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DISCUSSION

GEORGE W. MOORE, *U. S. Geological Survey, Menlo Park, California*: You say that all the solution from infiltrating water occurs at the bedrock surface, and that solution by water flooding seasonally from the river into the cave system and out again, as Thrailkill has suggested, is unimportant. I presume then that you favor Bogli's idea of solution where ground water and infiltrating water, both saturated but with different carbon dioxide partial pressures, mix at depth to form an undersaturated solution?

WATSON: No, actually I favor Swinnerton's hypothesis, and Thrailkill's position is similar. The fluctuation that takes place during the wet periods is related to the rising of the river, but it isn't necessarily backflooding that causes the fluctuation. The fluctuation also comes from water flowing in from the ground-water reservoir.

MOORE: And the infiltrating water still maintains undersaturation by the time it gets down into the cave passages?

WATSON: Very probably. You said I said all the solution takes place in the mantle and on the bedrock surface. Of course some of the solution does form cave passages. It is clear that solution is increased in the areas of fluctuating water level. Whether this is explained by Bogli's hypothesis of mixture corrosion is not clear. Certainly in the cave passages quite near the Green River there is extensive mixing of Green River backup water and water flowing toward the Green River in underground streams, and this might lead to increased solution potential. But such massive mixing would not occur in areas far from the Green River, where water levels also fluctuate. There, mixing resulting in greater solution potential would have to be of two batches of water which have infiltrated from the surface. Perhaps the agitation of the water during the fluctuation has as much to do with the solution as does the mixing.

WILLIAM I. GARDNER, *U. S. Bureau of Reclamation, Denver, Colorado*: One of the problems in limestone terrain is getting an estimate of ground-storage capacity. I wonder if you have been able to arrive at an estimate for this basin, or to estimate the average effective porosity on a basin-wide basis?

WATSON: No, I have not. We can give rough estimates, taking how much water falls and how much goes out. One of our problems is that there must be other outlets in the bed of the Green River which we do not know of. We have a program of scuba divers going along with thermometers to check for such rises.

I want to comment on the hydrostatic flow nets which have been illustrated in several papers today. I am sure that the flow-net surface is quite flat in the Central Kentucky Karst. The Green River valley is always shown as a great opening, which flow arrows originating in highlands come up to from great depth. That opening is probably no more significant to the flow net as a point of release than openings which extend as much as ten miles back under the Mammoth Cave Plateau and Sinkhole Plain. The whole region is underlain by underground streams that are within 100 feet of the Green River level. No high head exists in the Central Kentucky Karst from which deep-lying flow lines can derive. Therefore, no arrows should be shown rising from great depth to the Green River valley.