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NATIONAL SPELEOLOGICAL SOCIETY

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The Society serves as a central agency for the collection, preservation, and dissemination of information in fields related to speleology. It also seeks the preservation of the unique faunas, geological and mineralogical features, and natural beauty of caverns through an active conservation program.

The affairs of the Society are controlled by an elected Board of Governors, which appoints national Officers. Technical affairs of the Society are administered by specialists in fields related to speleology through the Society's Biology Section, Section on Cave Geology and Geography, Social Science Section, and Research Advisory Committee.

Publications of the Society include the quarterly *Bulletin*, the monthly *News*, an annual *Speleo Digest*, *Special Publications*, and the *Caving Information Series*. Members in all categories except Family receive the *Bulletin* and *News*.

A library on speleological subjects is maintained by the Society at its office. Material is available to Society members at a nominal fee to defray the cost of handling and to others through inter-library loan. An extensive file of information on caves of the United States is maintained by the Society and currently is housed in Bloomington, Indiana.

Geomorphology and Hydrology of the Indiana and Kentucky Karst: A Symposium

Arthur N. Palmer* and Michael C. Moore**

The limestone regions of Indiana and Kentucky are, in many respects, the birthplace of American karst science. Numerous studies of this area by prominent geologists date back to the late 19th and early 20th centuries, culminating in the work of Clyde Malott, whose thorough and painstaking investigations of the Indiana Karst span the period from 1920 to 1950. Although, in subsequent years, interest in karst geomorphology has spread to limestone regions throughout North America, the karst of Indiana and Kentucky has remained the subject of research on an ever-increasing scale.

The papers included in this *Bulletin* issue were originally presented as a symposium at the 1973 annual convention of the National Speleological Society in Bloomington, Indiana and provide a representative cross section of the field work currently being done in the karst areas of Indiana and Kentucky. The papers are arranged in geographic sequence, from Indiana southward to northern and central Kentucky. The location of each study area is shown in Figure 1.

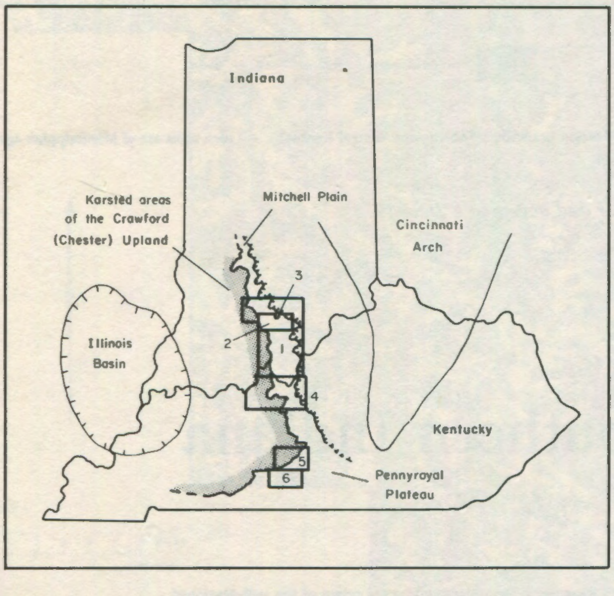

Figure 1. Location of study areas within the karst of Indiana and Kentucky: 1 = Mitchell Plain (M. Palmer): 2 = Lost River watershed (J. Bassett); 3 = sinkholes in Lost River watershed (R. Hall); 4 = karst of north-central Kentucky (A. George); 5 = Central Kentucky Karst (J. Hess); 6 = area of karst drainage to Barren River, Pennyroyal Plateau (S. Wells). The Illinois basin and Cincinnati arch are the dominant negative and positive features of the geologic structure, respectively, which exert the overall control over the pattern of rock exposures and geomorphic regions.

Each paper concerns a different aspect of the extensive karst landscape developed on the carbonate rocks of Mississippian age that are exposed in a continuous band from southern Indiana into much of central, western, and eastern Kentucky. Limestones and dolomites of older Paleozoic ages are found in southeastern Indiana and in the Bluegrass region of Kentucky, but karst features are comparatively sparse in these formations and, therefore, they have not been the subject of extensive field work by speleologists.

Similar geologic conditions exist throughout the karst belt on the Mississippian rocks. Carbonate rocks are underlain by siltstones and shales of late Devonian and early Mississippian ages and are overlain by predominantly clastic rocks of Late Mississippian and Pennsylvanian age. The gentle regional dip is nearly everywhere imperceptible to the eye. This sequence of rocks was deposited in a shallow epicontinental sea whose northern shoreline extended roughly through central Indiana. Throughout the Mississippian Period, the continental portions of the sea gradually were filled with marine limestones, as well as by detrital sediment from the land to the north, so that terrestrial conditions alternated with marine conditions during much of the Pennsylvanian.

The karst-forming limestones belong mainly to the Meramecian and Chesterian series. Their thickness increases southward, away from the sources of detrital sediment. Northward, toward the Mississippian shoreline, the lower Chesterian rocks become progressively more clastic in nature, eventually isolating the limestones as thin units sandwiched between sandstone and shale formations. The base of the carbonate sequence rises stratigraphically toward the north as the Osagian series, grading from the cherty limestone of the Ft. Payne formation to siltstone of the Borden group. Most of the carbonate units thin considerably toward eastern Kentucky, although they extend high into the Chesterian Series as a continuous limestone sequence interrupted only by thin shale or sand beds. Regional stratigraphic correlations are shown in Figure 2.

Gentle dips have been imparted to the beds mainly as the result of crustal subsidence and arching contemporaneous with deposition. The rocks of Indiana and of northern, central, and western Kentucky dip toward the center of the Illinois basin (Fig. 2).

The similarity of geologic settings among the various study areas is reflected in similar landscape types. Where the Mississippian limestones are exposed at the surface over large areas, karst topography is generally well developed as broad, low, sparsely dissected plateaus of sinkhole plain from which most drainage takes place underground through cave systems. In central Kentucky and southern Indiana, where the sinkhole plain is most prominent, it is named, respectively, the Pennyroyal Plateau and the Mitchell Plain. The karst topography is bordered on the down-dip edge by hilly uplands capped by relatively resistant clastic rocks of Late Mississippian and Pennsylvanian ages, known as the Crawford (or Chester) Upland. In those parts of the upland where limestone is exposed above base level, caves are numerous and commonly extensive, although surface karst features are frequent only in karst valleys entrenched below the base of the clastic cap-rock.

Although local variations in geology and geomorphology are described in each of the following papers, the value of compiling them in a single *Bulletin* issue is to emphasize the regional significance of each study. The conclusions of each paper should, therefore, shed light on the karst geomorphology of the entire Mississippian limestone belt in the east-central United States.

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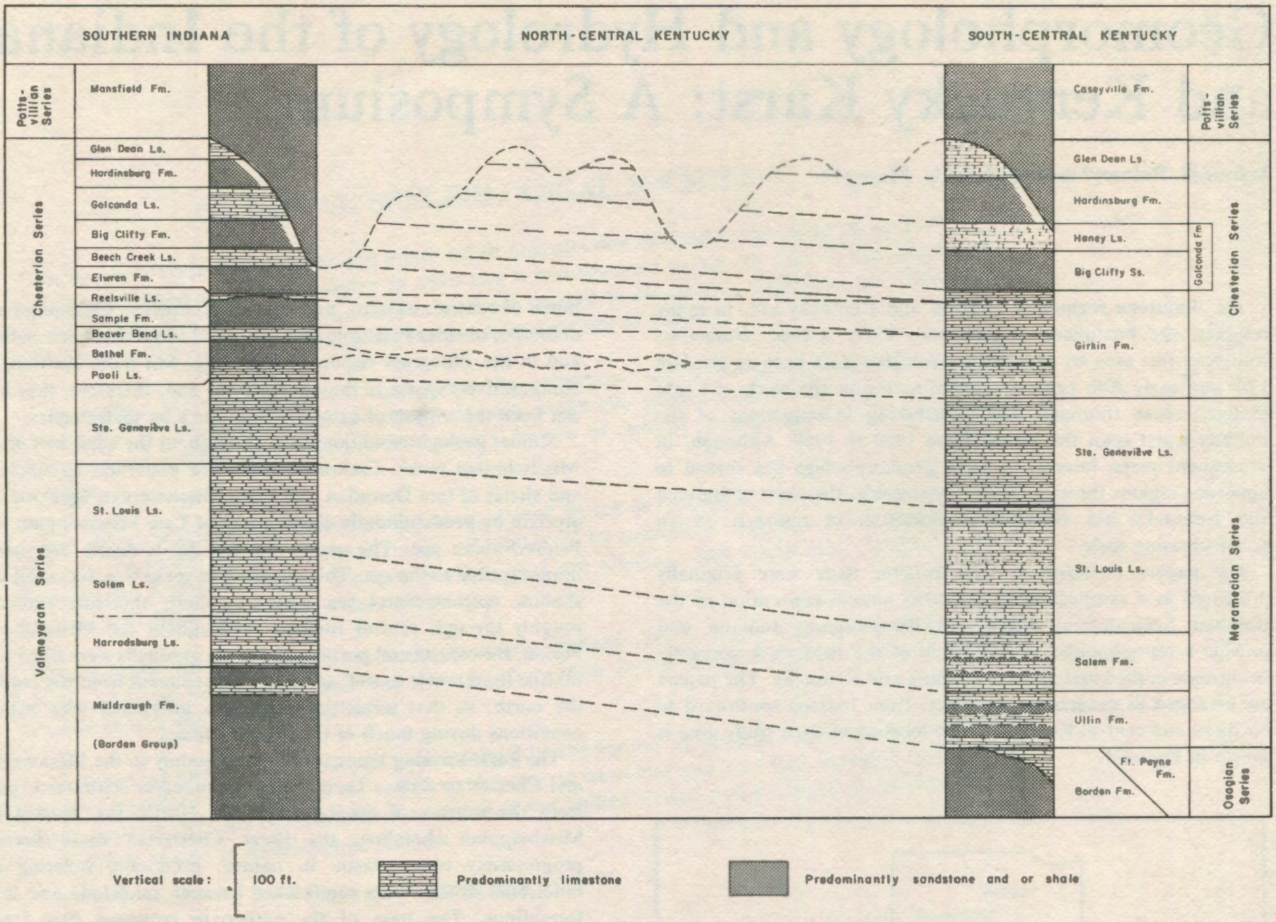

Figure 2. Generalized stratigraphic correlation chart of the Mississippian limestones and associated strata in southern Indiana and central Kentucky. All rock units are of Mississippian age, except for the Pottsvillian series, which is Pennsylvanian.

Manuscript received by the Editor and accepted 8 June 1975.

The Mitchell Plain of Southern Indiana

Margaret V. Palmer*

ABSTRACT

A topographic study of the Mitchell Plain of southern Indiana offers a framework for understanding the origin of the well-developed karst features located in its western portions. The Mitchell Plain is composed of three landform types and has been modified by at least three base-leveled surfaces, which cut across the different landform types. The landform types are (1) areas of sinkhole plain, (2) areas of unconsolidated surficial cover, and (3) areas of dissected upland. The base-level-controlled surfaces include: (1) the upper Mitchell Plain surface, (2) the lower Mitchell Plain surface, and (3) the Blue River Strath.

The Mitchell Plain of southern Indiana is a sparsely dissected, low plateau in which much of the surface has been subjected to karst development. It is bounded on the west by the Crawford Upland, an area of ridges capped by the predominantly clastic Chesterian series, and on the east by the Scottsburg Lowland, a broad plain formed on shale. Sub-accordant summit elevations of the Crawford Upland at altitudes of 900-1000 feet are believed to represent the hypothetical Lexington-Highland Rim peneplain (Thornbury, 1965, p. 191). The Mitchell Plain, which lies at altitudes of 700 to 1000 feet, is a composite, low-relief plateau formed by a combination of erosional and depositional events. Analysis of the geomorphic history of the area is complicated by the low gradient of the plateau surface, which nearly coincides with the regional dip of the rocks (about 30 ft/mi) into the Illinois structural basin. Despite local evidence for structural control, however, the

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entire surface is subtly discordant to the regional dip, in that the land surface possesses a gentler slope.

The Mitchell Plain is made up of three landform types and contains evidence for at least three different erosional-depositional surfaces transecting the various landforms. The three landform types, shown on Figure 3, include: (1) areas of sinkhole plain developed along the down-dip edge of the Mitchell Plain in the vicinity of major entrenched rivers, at relatively low elevations; (2)

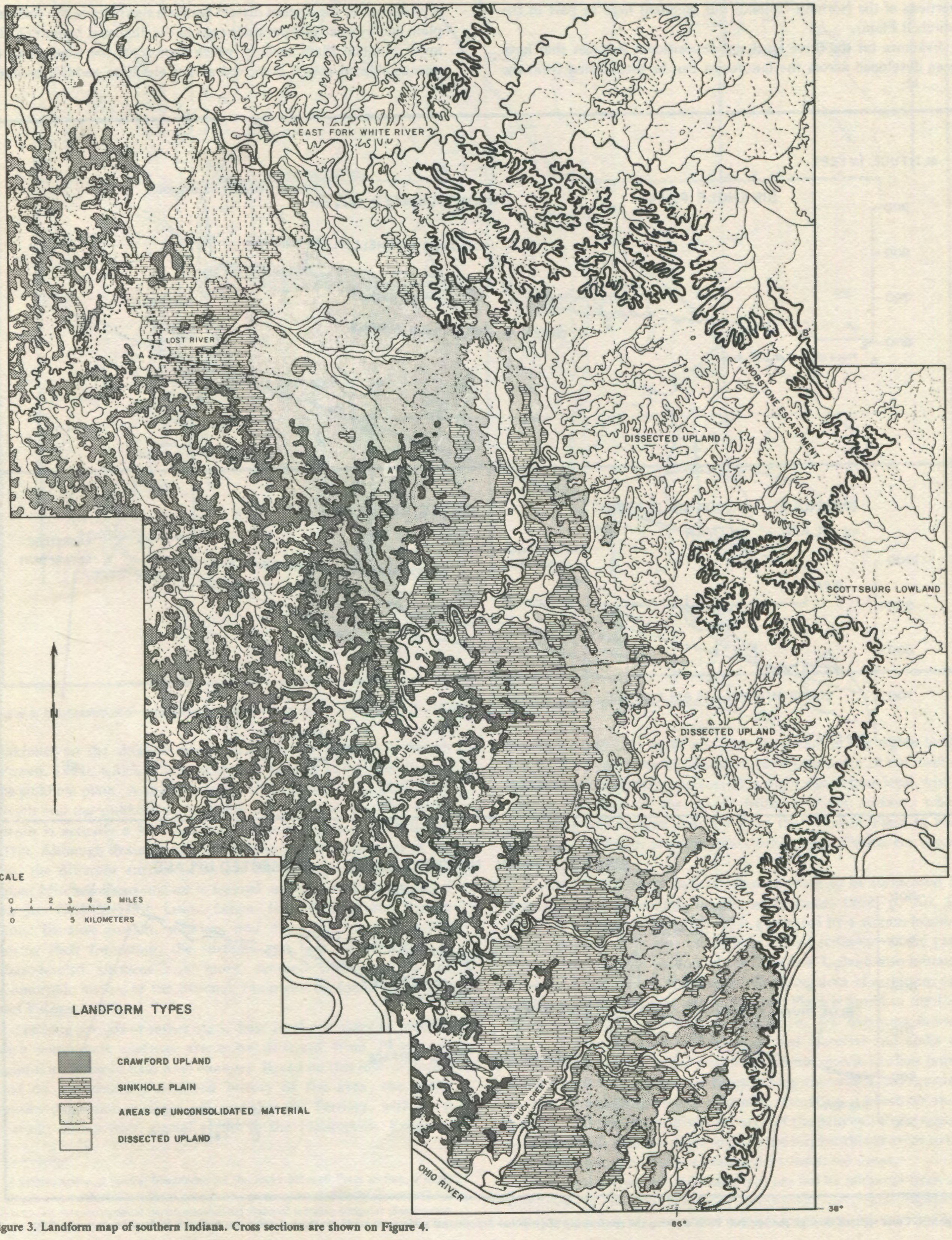

Figure 3. Landform map of southern Indiana. Cross sections are shown on Figure 4.

areas mantled with residuum and unconsolidated sediment to the east of the sinkhole plain, farther removed from the entrenched rivers; and (3) finely dissected areas of non-karsted limestone and shale ridges, at relatively high elevations in the furthest up-dip parts of the Mitchell Plain (commonly considered to be the western portions of the Norman Upland, but included here as part of the Mitchell Plain).

Evidence for the three erosional-depositional surfaces that have been developed across the landforms can be established from the cross sections (Fig. 4). Tracing the surfaces is difficult, as later dissection has unevenly lowered much of the landscape. The different erosion-depositional surfaces include: (1) the upper Mitchell Plain surface, best preserved on the gently sloping upper surfaces of the residuum and unconsolidated sediment but, also, represented by the highest divides between sinkholes in the sinkhole plain and by relatively flat ridge tops in the dissected region; (2) the lower Mitchell Plain surface, which represents the lower limit of unconsolidated cover, preserved as low-elevation saddles between

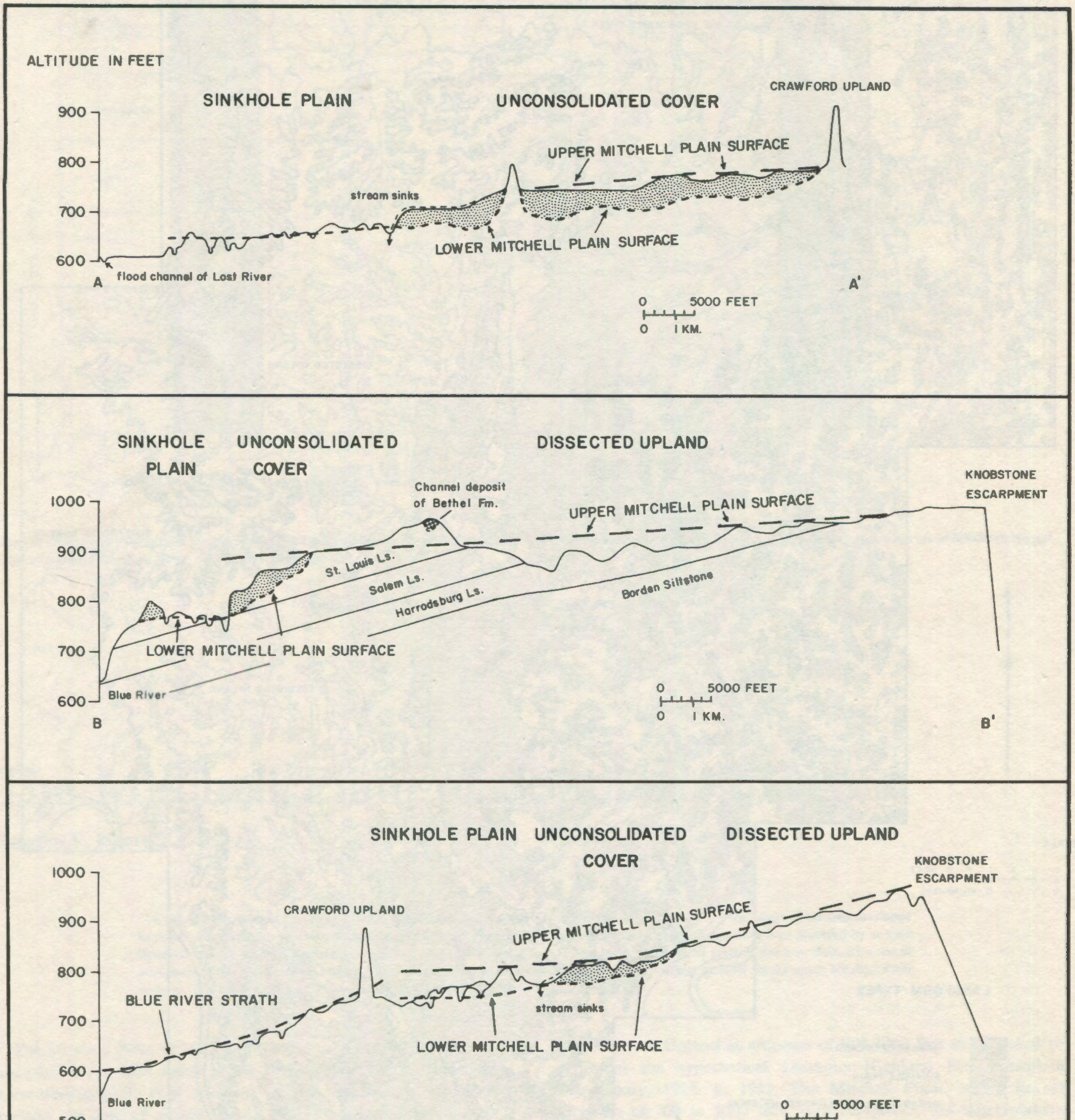

Figure 4. Cross sections through the Mitchell Plain showing the relationship of erosional-depositional surfaces to landform types. Geology on section B-B' from Sunderman (1968).

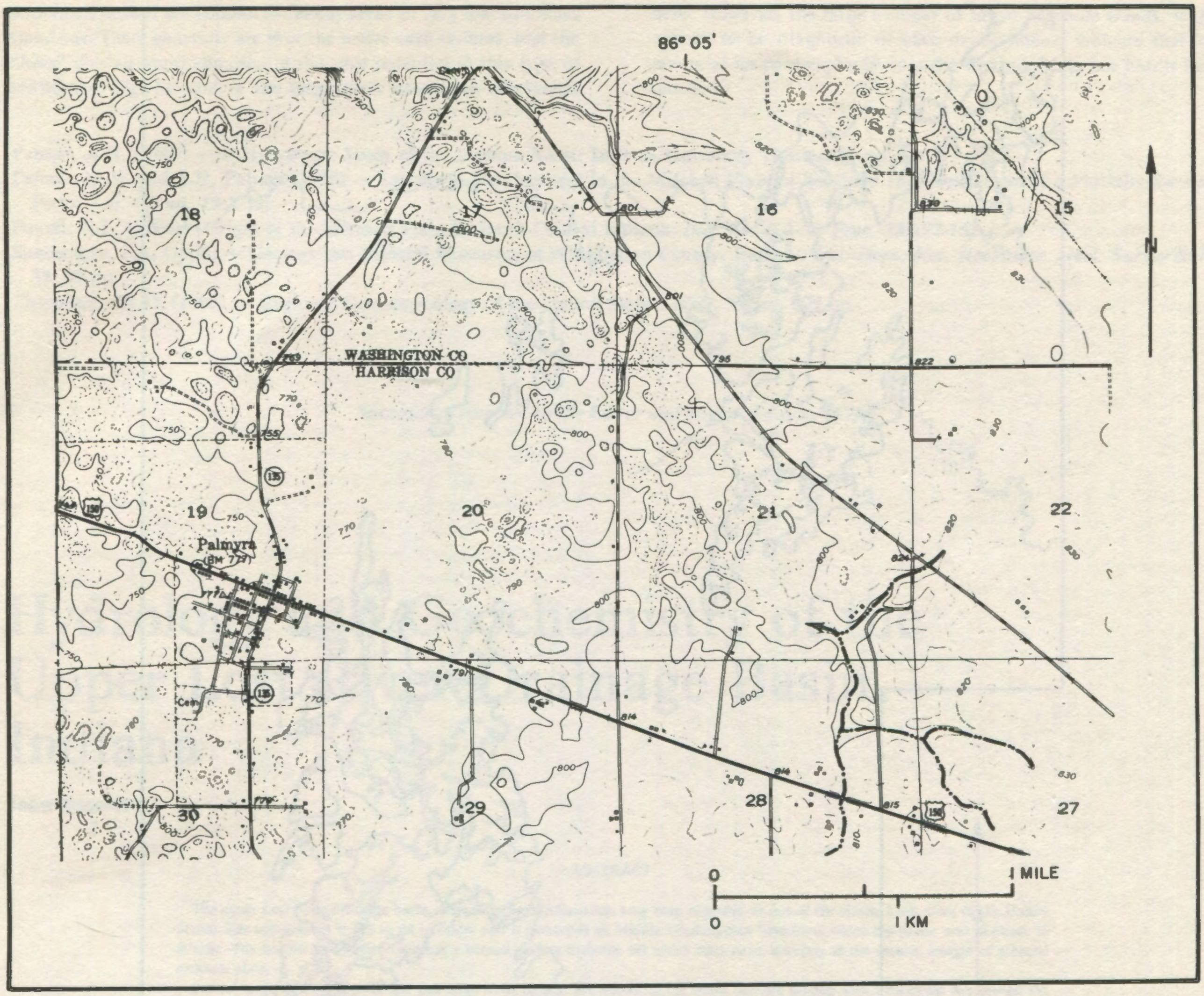

Figure 5. Relationship of sinking stream to sinkhole development. From U.S.G.S. 7½-minute topographic map of the Palmyra Quadrangle.

sinkholes on the sinkhole plain[1], and (3) the Blue River Strath (Powell, 1964), which is represented by karsted valleys heading in the sinkhole plain. A slight break in slope between the Blue River Strath and the sinkhole plain proper suggests that the Blue River Strath is actually a later surface entrenched below the Mitchell Plain. Although drainage is predominantly in the direction of the dip, the different surfaces transect the geologic structure. The upper Mitchell Plain surface is formed on the varied lithologies of the Ste. Geneviève, St. Louis, Salem, Harrodsburg, and Borden units. Because erosion, solution, and deposition were all active during their formation, the surfaces can best be described as "base-leveled surfaces". A more detailed discussion of the geomorphic history of the Mitchell Plain can be found in Palmer and Palmer (1975).

The long periods of rather static base level necessary to produce such continuous surfaces are quite different from Pleistocene conditions of rapid base level changes. Based on this interpretation and on the generally accepted history of the area, the major erosion-deposition surfaces are probably all Tertiary, with karst features dating from glacial events in the Pleistocene. Erosional history includes: (1) uplift and slow dissection of the region in the late Tertiary, with the formation of residuum down to the level of the lower Mitchell Plain surface; (2) a slight rise in base level, with a halt at the level of the upper Mitchell Plain surface, where weathering and erosion reduced most the Mitchell Plain to base level; and, (3) finally renewed dissection prior to Pleistocene glacial events.

A study of the karst features showed them to be influenced by drainage from nonsoluble landforms (Palmer, 1969, p. 30). On Figure 5, a line of sinkholes is seen, generated by a stream heading on the impermeable area of unconsolidated sediment to the east. Water draining from outliers of the Crawford Upland also initiated sinkhole development. The largest remaining area of residuum and unconsolidated sediment on the Mitchell Plain is found in the Lost River area (Fig. 1). Lost River, one of the few east-west flowing rivers, drains this area of unconsolidated material and sinks on contact with bedrock to the west in a series of classic swallow holes. Karst streams are continually transporting unconsolidated material so that the nonkarst area is gradually retreating. Limestone areas lacking sinkholes in the dissected area of the Mitchell Plain appear to be identical, in terms of geology, to the large sinkhole areas to the west; however, they are not bounded by insoluble areas.

Caves underlying the sinkhole plain are fed by recharge from the sinkholes. A few sinkholes are abnormally deep and represent collapse sinkholes over active stream passages. If the sinkhole

[1] Determination of vertical boundaries for the lower Mitchell Plain surface is highly subjective, as much of the surface is obscured by unconsolidated material. From well data it is known that the depth of the unconsolidated material is highly irregular (Sunderman, 1968, p. 32-33).

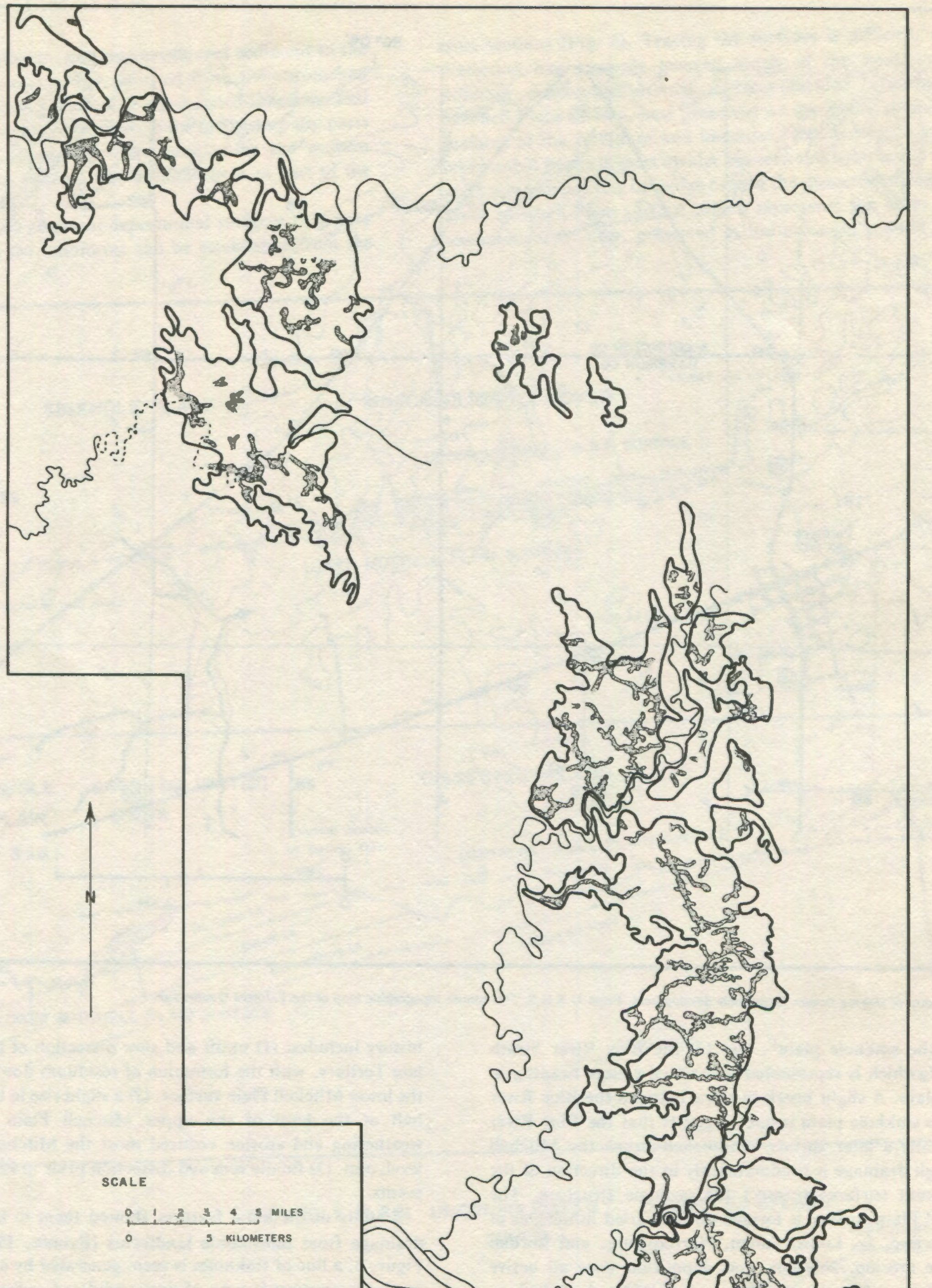

Figure 6. Map of linear zones (stippled) with low sinkhole bottom elevations. Dark lines match boundaries of sinkhole plain development shown on Figure 3.

bottom elevations are contoured, linear areas of very low sinkholes stand out. These generally are over the active cave systems, and the "lows" give an approximation of the cave potential in this type of setting (see Fig. 6). Only a few large caves have been discovered here. However, the large number of linear sinkhole trends, which appear to be diagnostic of cave development, indicate that the potential for cave exploration in the sinkhole plain has hardly been touched.

Palmer, A.N. (1969) — A Hydrologic Study of the Indiana Karst: Indiana University Thesis, 181 pp.

Palmer, M.V. and A.N. Palmer (1975) — Landform Development in the Mitchell Plain of Southern Indiana: Origin of a Partially Karsted Plain: *Zeit. Geom.* 19:1-39.

Powell, R.L. (1964)—Origin of the Mitchell Plain in South-Central Indiana: *Ind. Acad. Sci., Proc.* 73:177-182.

Sunderman, J.A. (1968) — Geology and Mineral Resources of Washington County, Indiana: *Ind. Dept. Nat. Resources, Geol. Survey Bull.* 39, 90 pp.

Thornbury, W.D. (1965) — Regional Geomorphology of the United States: NYC, Wiley, 609 pp.

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Hydrology and Geochemistry of the Upper Lost River Drainage Basin, Indiana

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ABSTRACT

The upper Lost River drainage basin of south-central Indiana has long been regarded as one of the classic karst areas of the United States. The upper basin is 163 sq mi in extent and is developed on Middle Mississippian limestones which dip to the west at about 30 ft/mile. The eastern part of the basin has a normal surface drainage net which terminates abruptly at the eastern margin of a broad sinkhole plain.

Several large sinking streams and two large karst springs are known in the basin. Stream tracing with fluorescent dye proved the existence of two major, independent karst drainage systems. One has a drainage area of at least 46 sq mi in the northwestern portion of the basin and discharges at the Orangeville Rise. The other has a drainage area of at least 110 sq mi, principally in the eastern part of the basin and discharges at the Rise of Lost River. The karst springs respond very rapidly to rainfall, and quantitative dye tracing has shown that flow velocities as great as 5.5 mi/day exist in subsurface drainage conduits.

Based on minimum monthly discharge per unit area calculations, base flow in the portion of the basin discharging at the Orangeville Rise is significantly higher than the base flow from the upper Lost River basin as a whole, indicating greater infiltration and storage in the karsted portion of the basin. The calculated water balance for the Orangeville Rise drainage basin substantiates the previous estimate of a 46 sq mi drainage area, based on dye-tracing experiments, and documents a roughly 10-fold increase in ET between summer and winter months.

Waters sampled from the Orangeville Rise have a dominantly $CaCO_3$ composition at moderate to high flow rates, but, at low flow rates, contain appreciable Mg and SO_4, are higher in CO_2, and are closer to (or are at) saturation with respect to calcite. Concentrations of major dissolved chemical species are inversely proportional to discharge, as defined by regression equations of the form $Y = a - b \log X$, or $Y = aX^b$. The molar ratios, Ca/Mg and HCO_3/SO_4, relate directly to discharge, but the Ca/Mg molar ratio is inversely related to SO_4 concentration. Saturation with respect to calcite in the spring waters is controlled both by discharge and by CO_2 partial pressure. These have a seasonal trend, with high CO_2 pressures in the summer and late fall and low pressures in the spring and early winter.

Waters entering swallow holes along the eastern margin of the sinkhole plain are consistently saturated or supersaturated with respect to calcite and have equilibrium CO_2 pressures about an order of magnitude lower than those at the Orangeville Rise. Supersaturation in sinking streams and downstream from springs is attributed to degassing of CO_2.

Two principal forms of recharge occur in the basin: :1) direct and rapid recharge from open swallow holes and (2) diffuse infiltration from the sinkhole plain. At the Orangeville Rise, the observed changes in chemistry with decreasing flow rate may be attributed to either 1) a mixing mechanism, where a Ca-Mg-CO_3-SO_4 groundwater becomes dominant over $CaCO_3$ water derived from sinking streams, or 2) a calcite saturation mechanism, where gradual saturation with respect to calcite and undersaturation with high-Mg-calcite, dolomite, and SO_4 is the principal control.

Introduction

The upper Lost River drainage basin of south-central Indiana has long been regarded as one of the classic karst areas of the United States and has been the basis for numerous descriptive geologic studies (Malott, 1932, 1952; Childs, 1940). These early works did not discuss the importance of hydrologic and geochemical factors on the development of the karst, although modern study is so oriented (Holland *et al.*, 1964; Thrailkill, 1968). Because of the ability of groundwater in karst terranes to modify its flow pattern by solution of the bedrock, is is important to understand both the chemical evolution of the water as it moves through the drainage system and the significance of variations in water chemistry relative to the hydrologic setting.

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Some modern studies of karst drainage systems have related geomorphology to hydrology (Palmer, 1969; Brown, 1969); others have documented what are apparently seasonal variations in water chemistry in carbonate terranes (Shuster and White, 1971). The variability in geochemistry relative to hydrology has, however, received little attention until quite recently (Jacobson and Langmuir, 1974). Some very basic information, such as the variation of chemical parameters with discharge is lacking, particularly for the Indiana karst.

The objective of this study is to relate the chemical variability in the surface and groundwaters of a typical karst terrain in Indiana to hydrologic parameters and the general hydrologic regimen of the drainage system. This research was supported by the U. S. Department of Interior, Office of Water Resources Research, Grant No. 14-31-0001-3689 and by Indiana University.

Description of the Study Area

The upper Lost River drainage basin is located in portions of Lawrence, Orange, and Washington Counties, Indiana, and includes about 163 sq mi above the confluence of Lost River with Lick Creek (Fig. 7). This part of the basin sets astride the Mitchell (sinkhole) Plain (Malott, 1922) and has undergone intense karst development. The Mitchell Plain is underlain by Middle Mississippian limestones at least 275 ft thick (Sunderman, 1968), and dip to the west at about 30 ft/mi.

The upper Lost River drainage basin contains three distinct physiographic divisions. The eastern part, at elevations of 660 to 900 ft, has normal surface drainage through the two forks of Lost River, Carter Creek, Stampers Creek, and a few other, smaller, streams (Fig. 7). The middle part of the basin is a broad, rolling, sinkhole plain, where surface drainage is essentially absent. The surface drainage from the eastern part of the basin terminates rather abruptly in a series of major sinks and swallow holes at elevations of 660 to 680 ft along the eastern edge of the sinkhole plain. In all, drainage from about 80 sq mi in the eastern part of the basin is diverted underground. The entire low flow of Lost River, as well as that of Stampers Creek to the south, sinks into subterranean solution conduits in the limestone and flows westward beneath the sinkhole plain straight-line distances of 7 and 10 mi, respectively, to the Rise of Lost River (Figure 7, E).

Between the sinks and the rise, is a 22-mi meandering surface channel known as the "dry bed." Several swallow holes and stormwater rises are located along the dry bed, many of which apparently serve both as sinks and as resurgences. Perennial flow in the downstream section of Lost River, west of the sinkhole plain, comes from two large karst springs, the Orangeville Rise and the Rise of Lost River (Fig. 7 B, E). The Orangeville Rise is the second-largest karst spring in the state, with a minimum measured discharge of 8.9 cfs. The Rise of Lost River is somewhat smaller.

The western part of the basin is in the Crawford Upland, a rugged, hilly area developed on clastic rocks of the Upper Mississippian Chester Series (Malott, 1922). Local relief in the Crawford Upland is more than 200 ft, with sandstone-capped ridge tops extending to elevations greater than 800 ft asl. Many streams in the Crawford Upland are incised into the underlying limestone and sink at several places along their channels. The floors of abandoned surface valleys have a karst character and have been described as karst valleys (Malott, 1939).

Hydrology

Subsurface flow paths from swallets to springs in the Lost River area have been defined by fluorescent dye-tracing techniques.

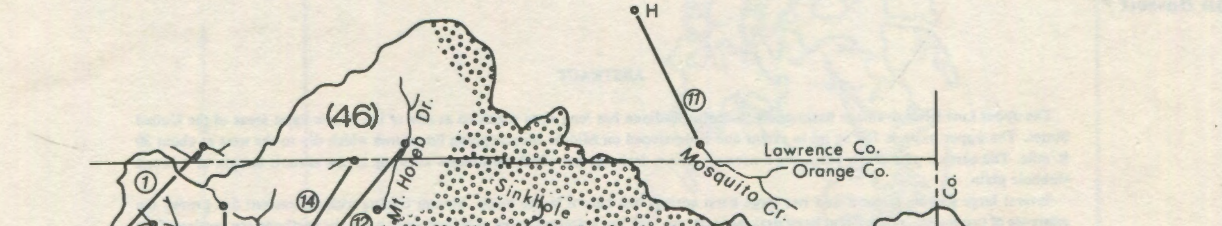


Fig. 7. Map of the upper Lost River drainage basin, showing location of stream gages, sampling sites, and dye tracing results: A, downstream gage; B, Orangeville Rise; C, upstream gage; D, Stampers Creek gage; E, Rise of Lost River; F, Mill Spring; G, Sulphur Spring; H, Donaldson Cave Spring; I, Orleans sewage treatment plant; J, Mather storm water rises.

During the period 1966 to 1973, 14 successful dye tests were made in the upper Lost River basin. Murdock and Powell (1966), as part of a Soil Conservation Service study, defined the general extent of the basin and the approximate drainage areas of the Orangeville Rise and of the Rise of Lost River (Fig. 7, Nos. 1-10). Additional tracing by the author was used to define in more detail the subsurface drainage basin of the Orangeville Rise and to delineate the northeastern boundary of the basin (Fig. 7, Nos. 11-14) (Bassett and Ruhe, 1973; Bassett, 1974).

These dye tests indicate that surface and subsurface drainage divides in the basin generally coincide; the only major exception is the South Fork of Stampers Creek. Although it is part of the topographic drainage basin of Lost River, dye tracings indicate that this part of the basin drains underground to Mill Spring and Lick Creek to the southwest (Murdock and Powell, 1966; Fig. 7, Nos. 7-9). Sinking streams located outside the topographic drainage basin of Lost River drain to springs outside the basin. A small sinking stream just north of the basin divide in the Crawford Upland drains southwest through a large, known cavern to Sulphur Spring along Sulphur Creek (Fig. 7, No. 1). Mosquito Creek, which is located just north of the low basin divide across the Mitchell Plain, drains north to Donaldson Cave Spring (Fig. 7, No. 11).

Quantitative dye tracing, using Rhodamine WT dye and a filter fluorometer, showed that flow velocities in excess of 5.5 miles per day exist in the major karst drainage systems under moderate to high flow rates, and that higher discharges are accompanied by substantially shorter travel times (Bassett and Ruhe, 1973). Dye injected into the effluent stream of the Orleans sewage treatment plant (Fig. 7, I), as it enters an open sinkhole on the plant grounds, appeared 25 hours later at the Orangeville Rise, 5.9 mi to the west. The Orangeville Rise normally responds to rainfall within 8 hours. This rapid response is related to its direct connection with several sinking streams in the Crawford Upland and adjacent parts of the Mitchell Plain and to rapid conduit flow velocities.

No sinking stream was found to resurge at both the Rise of Lost River and the Orangeville Rise, indicating that at least two independent karst drainage systems exist in the area; each has a definite drainage basin and discharge point. The first and largest (by area) system drains the eastern part of the basin, mainly through the swallow holes of Lost River and of Stampers Creek, and discharges at the Rise of Lost River. The other major system drains the northwestern part of the basin and discharges at the Orangeville Rise.

The flow regimen of the Orangeville Rise is modified by the presence of ephemeral stormwater rises, such as the Mater Rises (Fig. 7, J), which divert water from the subsurface conduit systems supplying the Rise back to the dry bed. As determined by actual stage and discharge measurements at Orangeville, these outlets effectively limit the maximum discharge of the Orangeville Rise by limiting the hydraulic head that may be imposed on the system.

Drainage areas of the two principal rises were calculated from the results of dye-tracing experiments. The Orangeville Rise drains about 46 sq mi, whereas the Rise of Lost River drains about 110 sq mi; much of the latter is in the eastern, surface-drained portion of the basin (Fig. 7).

Discharge in the karst-drainage systems was monitored from May, 1972 to June, 1973 at three recording-gage stations established in the basin (Figure 7, A, B, and C). Rating curves and hydrographs were constituted for each of the sites. Gage A, located 5 mi downstream from the Orangeville Rise and 4 mi below the Rise of Lost River, monitored discharge from the entire 161 square mile basin. Gage B measured discharge from the Orangeville Rise portion of the basin (46 sq mi), and gage C measured discharge from 51 sq mi of surface drainage eventually entering the Sink of Lost River. Periodic discharge measurements were also made near the Sink of Stampers Creek (Site D).

Analysis of base-flow regimen was made using minimum-monthly discharges. Base flow is highly dependent on basin geology and evapotranspiration. Minimum-monthly discharge per square mile of basin area is a useful index of groundwater storage in and the hydrologic homogeneity of a drainage basin. Minimum-monthly discharge (Q_{min}) and basin area were used to compare the low-flow regimen at gages A and B. During the period June, 1972 to June, 1973, minimum-monthly discharge at gage A ranged from 0.10 to 1.43 cfs/sq mi and at gage B varied from 0.19 to 2.74 cfs/sq mi. The ratio $(Q_{min}/area)_A$ ranged from 1.93 in April, 1973 to 2.98 in August, 1972; which implies differences in the base-flow regimen of the Orangeville Rise compared to that of the total basin. Values greater than 2 are associated with the dry summer months.

It is apparent that the drainage basin of the Orangeville Rise, which contains a much larger percentage of sinkhole plain, has a higher base flow per unit area than the basin as a whole. This must be the result of greater infiltration and groundwater storage in the karsted (sinkhole plain) portion of the basin. This tendency has also been noted by Palmer (1969), who found the surface streams in the Indiana karst to have a higher base flow per unit area than have streams in adjacent areas of clastic rocks. This, again, is, apparently, due to higher infiltration and groundwater storage in karsted areas.

Water Balance

Water balances for selected storms in the upstream part of the basin (Gage C) were calculated and presented by Ruhe (1975). Actual evapotranspiration was determined by first calculating potential evapotranspiration, using the Blaney-Criddle method (1950), and then comparing these values with soil moisture changes on level upland sites, as determined from neutron-probe moisture-gage measurements. Relations between actual evapotranspiration (ET) and empirically determined potential evapotranspiration (PET) were derived by regression analysis for both increasing and decreasing soil moisture conditions (Ruhe, 1975).

These results were used in the present study to calculate a water balance for the Orangeville Rise drainage basin. Rainfall, discharge, and soil moisture data for the period June, 1972 to July, 1973 were used in the calculations. The simple mathematical model used in the balance may be expressed as:

$$I = O + ET + \Delta S$$

where I = rainfall, O = spring discharge, ET = evapotranspiration, and ΔS = change in groundwater storage.

The study period was divided into seven time intervals, ranging in duration from 37 to 81 days and including any number of hydrologic events. Intervals were chosen to begin and end at major hydrologic events. Rainfall during any time interval was converted from inches of precipitation to acre-feet (Table 1, col. 3). Spring discharge was similarly converted to acre-feet and represents the total measured flow during the time interval (Table 1, col. 4). Discharge ranged from 14% of rainfall during intervals 1 and 2 (summer months) to about 100% during intervals 4 and 5 (winter months) (Table 1, col. 5). Similar I/O values were derived by Ruhe (1975) for single hydrologic events. Clearly, evapotranspiration is a major factor in the balance, especially during the summer months.

In instances where Blaney-Criddle PET exceeded I (intervals 1 and 2) ET was set equal to I. Otherwise, ET for each interval was calculated from the empirical Blaney-Criddle (1950) values, using the regression equations modified from Ruhe (1975) (Table 1, col. 6).

$$ET = 0.05 + 0.59\ PET \text{ (increasing soil moisture)}$$
$$ET = 0.26 + 0.24\ PET \text{ (decreasing soil moisture)}$$

Changes in storage were estimated, using a generalized spring-flow recession curve constructed by graphically superimposing individual winter and spring recession curves on the long summer base flow recession, as described. The discharges at the beginning and end of each time interval were then determined, and the total area under the generalized recession curve between the two discharges was regarded as an estimate of the change in storage. A similar technique has been used by Atkinson (1975) in the (English) Mendip Hills karst. In this manner, an estimate of the total change in storage is obtained, be it in the form of soil moisture, intergranular water, or conduit storage. Positive ΔS value represents an increase in storage over the time period, and negative value a decrease (Table 1, col. 7).

TABLE I. Water balance for the Orangeville Rise drainage basin.

(1)	(2)	(3)	(4)	(5)	(6)	(7)	(8)
Interval	Duration (days)	I*	O*	I/O (%)	ET*	ΔS*	Bal. (%)
1 June 1 August 20, 1972	81	26.4	3.8	14	26.4	-2.8	- 3.8
2 August 21 October 29	70	20.7	2.9	14	20.7	+1.4	-21
3 October 30 December 5	37	15.3	5.5	36	3.2	+2.3	28
4 December 6 January 18	44	11.4	10.8	95	2.8	+0.1	-18
5 January 19 March 2	43	6.6	7.9	120	0.8	-0.4	-26
6 March 3 May 6	65	32.4	19.7	61	3.7	+1.3	24
7 May 7 July 5, 1973	61	20.1	10.0	50	7.2	-1.6	6

* Units are acre-feet x 10^3.

The balance (Table 1, col. 8) is obtained by adding *O*, *ET*, and ΔS, and subtracting the resulting value from *I*. The results are here expressed as the percentage of rainfall (*I*) unaccounted for, or in excess of the balance. Positive values indicate precipitation unaccounted for, and negative values indicate insufficient rainfall to make the balance.

The balances range from 28 percent to -26 percent of the total rainfall input. The mean value of the 7 balances is -1.5%, indicating a tendency for a slight precipitation deficiency. The results of the balance generally support Ruhe's (1975) observations, based on single storm events, that 10 to 17% of measured precipitation is discharged at the Orangeville Rise during the late summer months, 79 to 86% during the winter months (January to April), and 21 to 46% during the intervening, transitional, months (November to December and May to July). All of these figures are 4 to 13% higher than the corresponding runoff/precipitation ratios for the upstream, non-karst, part of the basin discharging through gage C (Ruhe, 1975). These figures support the earlier contention, that infiltration and groundwater storage are greater in karsted portions of the basin.

The balance tends to substantiate the previous estimate of a 46 sq mi drainage area for the Rise, based on dye-tracing experiments. The figures indicate that, in the warm, dry summer months, ET is a much more important factor in the balances than are the changes in storage. These parameters become of nearly equal importance in the winter months, primarily due to a roughly 10-fold decrease in ET.

Discharge-Water Chemistry Relations

During the period June, 1972 to June, 1973, water samples were collected from each of the recording gage stations at regular intervals as well as at the non-recording station located near the Stampers Creek sink (Figure 7, D). The principal purpose of the sampling program was to relate the chemistry of the water to the discharge of springs and surface streams and to the general hydrology of the basin.

A successful geochemical analysis of the karst drainage systems of Lost River, required that representative parts of the entire drainage system be monitored over the entire range of flow rates. As the relationships between discharge and chemical parameters were easily determined at each of the recording gage stations (Fig. 7, A, B, and C), these sites were used as a framework for sampling.

Calcium and magnesium content was determined on all samples by EDTA titration, or by atomic absorption spectroscopy. Bicarbonate content was determined from standard acid titration. To facilitate statistical comparison, all of the above concentrations are expressed as ppm $CaCO_3$. Sodium was determined by flame emission spectroscopy, chloride by $Hg(NO_3)_2$ titration, and sulfate turbidimetrically as $BaSO_4$. A Sargent Welch Model LS meter and combination glass electrode were used to determine sample pH in both the field and laboratory.

A total of 36 samples, representing essentially the entire range of measured flow rate, were collected from the Orangeville Rise (Site B). Total hardness (ppm$CaCO_3$) was found to be inversely related to discharge, as expressed by the regression equation $Y = 516.2 - 154.9 \log X$. Mean total hardness of the 36 samples was 258 ppm, but total hardness ranged from 400 ppm at a discharge of only 8.9 cfs to 120 ppm at a discharge of 148 cfs.

Calcium hardness of the same 36 samples was also inversely related to discharge, as expressed by the relation $Y = 402.7 - 113.2 \log X$ (Fig. 8). Mean calcium hardness of the 36 samples was 214 ppm. Total alkalinity varied inversely with discharge, as defined by $Y = 259.9 - 67.2 \log X$, but did not show nearly as much variation with discharge as did total hardness or calcium hardness, particularly at low flow rates. Total hardness at discharges less than 30 cfs varied between 270 and 400 ppm, but alkalinity varied only about 40 ppm.

While total alkalinity showed little variation with discharge at low flow rates, sulfate concentration greatly increased as discharge decreased, as expressed by $Y = 542 X^{-0.636}$ (Fig. 9). The determination of total sulfate includes not only the ion $SO_4^{=}$, but included SO_4 as the uncharged ion pairs $CaSO_4^\circ$ and $MgSO_4^\circ$ (Wigley, 1971). The sample with a total hardness of 400 ppm had a correspondingly high SO_4 concentration of 171 ppm. Three samples collected at a discharge of about 20 cfs showed SO_4 concentrations of only 65 to 70 ppm. At still higher discharges, the SO_4 content showed only slight variation with discharge and tended to level off at about 20 to 25 ppm. It is apparent that SO_4 becomes an increasingly dominant chemical species with decreasing discharge at the Orangeville Rise.

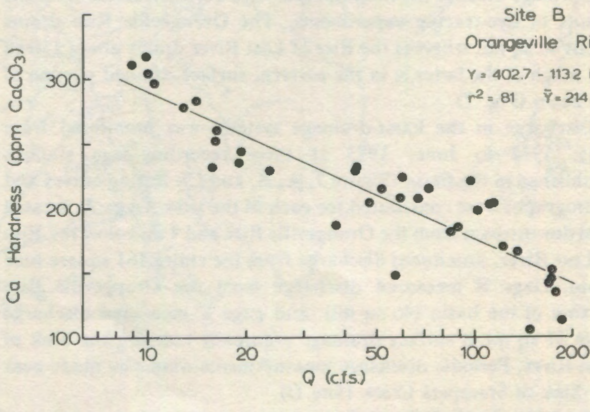


Fig. 8. Graph of calcium hardness vs. discharge at the Orangeville Rise.

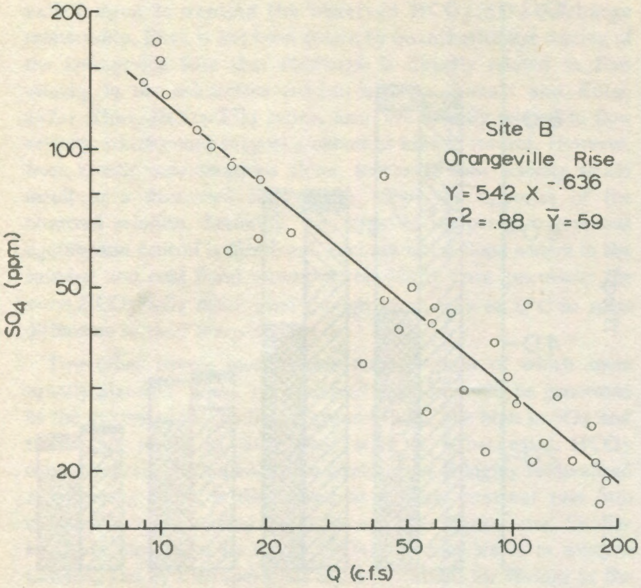


Fig. 9. Graph of SO_4 vs. discharge at the Orangeville Rise.

Sodium and chlorine are present in waters sampled from the Orangeville Rise, in approximately molar equivalent concentrations, which also vary inversely with discharge. The concentrations of both ions were generally low, with the mean Na concentration being 6.5 ppm and the mean Cl concentration being 10.5 ppm.

The chemistry of surface waters in streams which subsequently sink (sites C and D) is quite different from that of resurging water at the Orangeville Rise. The surface waters are much lower in Ca, Mg, and SO_4. The mean total hardness of 23 samples collected at site C was 190 ppm, much less than the value of 258 ppm for waters from the Orangeville Rise. Samples were not taken at site C during periods of extremely high flow.

The chemical content of waters at site C shows little correlation with discharge, up to at least 135 cfs. Similarly, total alkalinity, calcium hardness, and Na and Cl show very little correlation with discharge. The variations in total hardness, calcium hardness, and total alkalinity are much less than those observed at the Orangeville Rise. Sulfate concentrations at sites C and D generally increased with discharge, up to at least 100 to 150 cfs. This unusual relation is not easily explained and represents the only observed case where the concentration of a chemical species varied directly with discharge. At discharges greater than 100 cfs, SO_4 concentrations were greater than 20 ppm, but at discharges less than 10 cfs, SO_4 concentrations were less than 15 ppm.

Solution Chemistry

The thermodynamic state of saturation of water with respect to calcite is the most important chemical parameter expressing the ability of groundwater in karst terranes to modify its flow path by solution. The calcite saturation coefficient, S_c, has been used as an index of saturation and is defined as the activity product (Ca^{++}) ($CO_3^{=}$) divided by the solubility product of calcite. Using the FORTRAN computer program developed by Thrailkill (1969), saturation coefficients and equilibrium CO_2 partial pressures (P CO_2) were calculated for 70 spring and sinking stream samples from the basin. The solubility product of calcite was taken to be $10^{-8.40}$ at 25 C° (Langmuir, 1971). Waters sampled from the Orangeville Rise were found to have saturation coefficients ranging from 0.15 to 1.22, and were, thus, highly undersaturated to slightly supersaturated with respect to calcite (Fig. 10). In general, the most highly undersaturated waters were associated with high discharges in late winter and early spring. The correlation between saturation and discharge is not very good. In particular, four samples collected at discharges greater than 60 cfs during the winter months were found to have a higher degree of saturation than samples collected at less than 20 cfs discharge during the summer months.

What are apparently seasonal variations in equilibrium CO_2 pressures are seen at all 3 recording gage stations (Fig. 11). Plotted are the calculated CO_2 pressures for each sample and the hydrograph of the Orangeville Rise from August, 1972 to June, 1973. The highest CO_2 pressures ($10^{-1.7}$ to $10^{-2.3}$) are associated with samples from the Orangeville Rise, which are about an order of magnitude higher than CO_2 pressures associated with samples from the sinking stream site C ($10^{-2.6}$ to $10^{-3.6}$). Waters sampled at each site tended to have their highest equilibrium CO_2 pressures in late September and October, and their lowest values in late March and early April.

There is a fairly strong correlation between equilibrium CO_2 pressure and discharge at the Orangeville Rise, but significantly higher CO_2 pressures are found in waters collected in the summer months (Fig. 12), open circles) over those collected in the winter months (Fig. 12, darkened circles) at roughly equivalent flow rates. It appears that the reason for the comparatively high degree of saturation found in waters collected at moderate discharge rates in the winter is the much lower CO_2 pressure with which the waters are in equilibrium. Lower equilibrium CO_2 pressures would result in a higher degree of saturation for a given flow rate. Clearly both discharge and P CO_2 exhibit significant control on the state of saturation of the spring waters.

Although the waters at sites A and B, on the western side of the basin, have basically the same chemical character, the waters at site A, 5 mi downstream from the Orangeville Rise and 4 mi

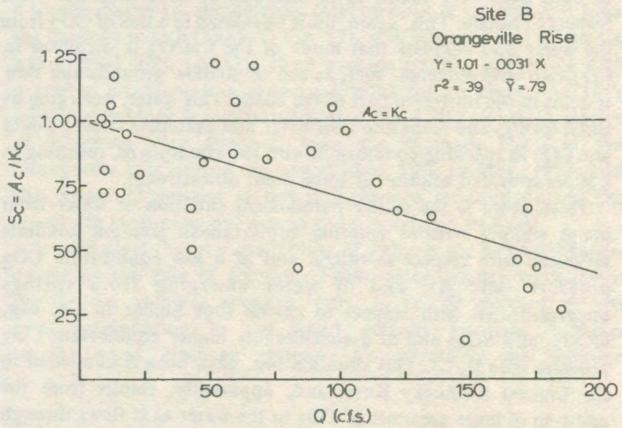


Fig. 10. Graph of S_c vs. discharge at the Orangeville Rise.

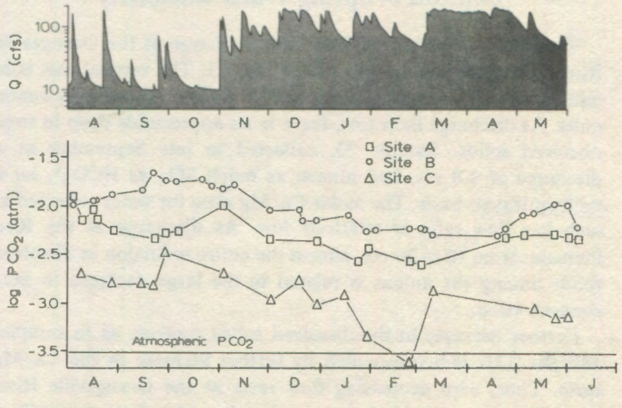


Fig. 11. Diagram illustrating variation in P CO_2 at each of the recording gage stations, plotted with discharge at the Orangeville Rise.

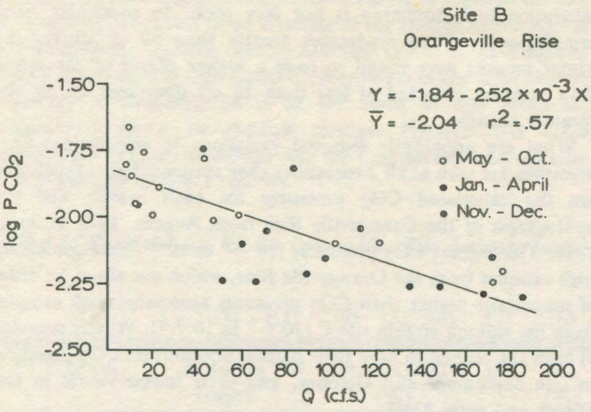


Fig. 12. Graph of P CO_2 vs. discharge at the Orangeville Rise.

downstream from the Rise of Lost River, are found to be consistently supersaturated with respect to calcite. Only under very high flow rates are waters at site A found to be undersaturated with respect to calcite. Waters at site A are in equilibrium with a lower CO_2 pressure than are those at site B (Fig. 11). Waters discharging from the Orangeville Rise and the Rise of Lost River apparently evolve CO_2 to equilibrate with the lower CO_2 partial pressure of the atmosphere. The loss of CO_2 is the probable cause of supersaturated surface waters.

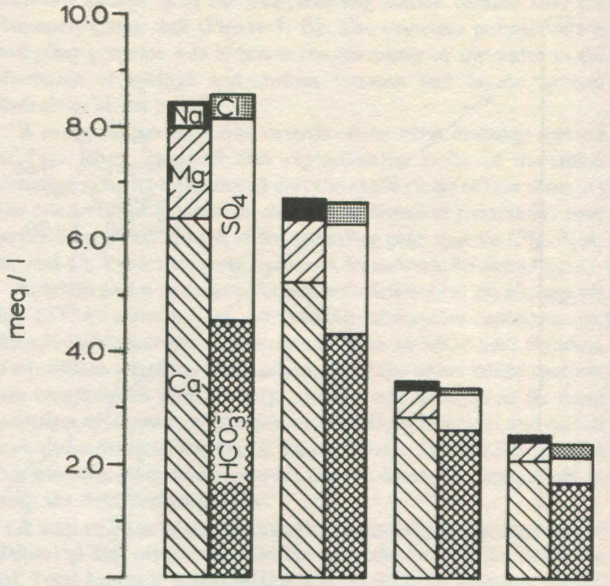
Waters sampled at sinking stream sites have equilibrium CO_2 pressures about an order of magnitude lower than those found at the Orangeville Rise. Because of their moderately high alkalinity and Ca concentration, and relatively low CO_2 pressures, the waters at both sites were found to be consistently supersaturated with respect to calcite. This, again, must be related to a loss of CO_2 from the water and suggests that much of the $CaCO_3$ is dissolved in CO_2-rich environments, such as soil or shallow groundwater-flow systems in the upstream part of the basin. This water, emerging as small springs and seeps into Lost River and Stampers Creek, would lose CO_2 in reaching equilibrium with the atmosphere, resulting in a supersaturated solution at some point downstream.

Thus, there is the rather paradoxical situation of water from major sinking streams entering subterranean solution conduits saturated with respect to calcite and at a low equilibrium CO_2 pressure (site A), and of water emerging from springs undersaturated with respect to calcite (but higher in Ca, Mg, HCO_3, and SO_4) and at a considerably higher equilibrium CO_2 pressure (site B, C). This situation has, also, been documented in the Central Kentucky Karst and, apparently, results from the addition of large amounts of CO_2 to the water as it flows through solution conduits (Thrailkill, 1969, 1972).

Controls of Spring Water Chemistry

Variations of water chemistry with discharge at the Orangeville Rise are summarized graphically in Fig. 13. The vertical axis is in milli-equivalents per liter, to express all concentrations in common units. As discharge increases, there is an appreciable drop in total dissolved solids. Sample 70, collected in late September at a discharge of 9.8 cfs, had almost as much SO_4 as HCO_3^-, on a milliequivalent basis. The molar Ca/Mg ratio for water collected at such low flow rates is relatively low. As discharge at the Rise increases from 10 to 20 cfs, almost the entire reduction in dissolved solids among the anions is related to the large decrease in SO_4 concentration.

Further decrease in the dissolved solids content, as in samples 140 and 131, is accompanied by further increase in the Ca/Mg ratio. Thus, with decreasing flow rates at the Orangeville Rise, there is a pronounced change in the chemistry from basically a Ca-CO_3 water to a high Mg-Ca-CO_3-SO_4 water.



Sample	70	78	140	131
Ca/Mg	3.96	4.64	6.17	7.00
Q (c.f.s.)	9.8	16.2	168	148

Fig. 13. Diagram illustrating variation in water chemistry with respect to discharge at the Orangeville Rise.

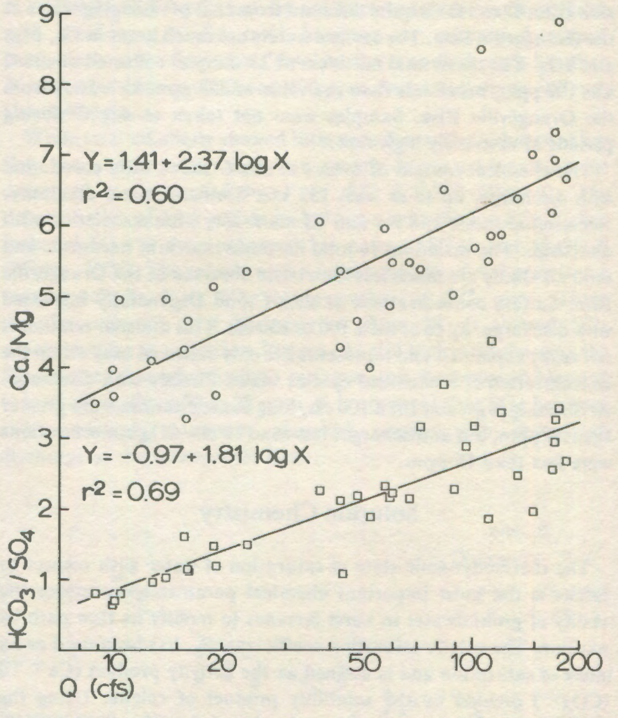


Fig. 14. Graph of Ca/Mg and HCO_3/SO_4 ratios vs. discharge at the Orangeville Rise.

Such a chemical regimen, where the Ca/Mg and HCO_3/SO_4 molar ratios vary with discharge, cannot be attributed solely to mere dilution of groundwater with relatively dilute recharge waters entering the subsurface through sinking streams or soil infiltration. Samples collected from the Orangeville Rise have Ca/Mg ratios ranging from 3.1 to 9.1 and HCO_3/SO_4 ratios ranging from 0.64 to 4.32; low ratios, in both cases, are associated with low flow rates (Fig. 14).

As discussed by Drake (1974), several mechanisms could be called upon to explain the observed HCO_3/SO_4-discharge relationship. First, it has been shown by quantitative dye tracing of the Orangeville Rise that discharge is directly related to flow velocity in the subsurface conduit systems (Bassett and Ruhe, 1973). Thus, HCO_3/SO_4 ratios, also, are directly related to flow velocity, which could suggest a chemical kinetic control. However, from kinetic considerations alone, increased flow velocity would result in a decreased HCO_3/SO_4 ratio—the opposite of the observed relation. Secondly, any type of temperature-governed equilibrium control is dismissed, because warm flood waters in the summer and cold flood waters in the winter have essentially the same HCO_3/SO_4 ratio, even though there may be 6°C or more difference in their temperatures.

Two other simple mechanisms remain, both of which seem equally plausible. First, the observed relation could be generated by the mixing of two bodies of groundwater, one high in SO_4 and the other lower in SO_4 (but with a substantial HCO_3 concentration). Such a condition could occur if highly mineralized SO_4-bearing water is discharged at a fairly constant rate into conduits carrying varying quantities of more dilute water. Similar ideas are postulated by Drake (1974) for river water in western Canada, and by Fish (personal communication) for springs in the Sierra de El Abra region of Mexico.

The second mechanism does not necessarily involve a mixing of groundwater bodies, but assumes that the conduit water becomes saturated with calcite at some low flow rate but remains undersaturated with more readily soluble SO_4-bearing minerals, such as gypsum. The solution of gypsum would add Ca to the water and further induce saturation of the water with respect to calcite. Although, as discussed previously, saturation of the Orangeville Rise with respect to calcite is controlled by both discharge and by PCO_2, there is a strong tendency toward saturation at low flow rates.

As thin gypsum beds are present in the lower St. Louis limestone beneath the western portion of the basin, the solution of gypsum could have a pronounced effect on the Ca/Mg ratio of waters at the Orangeville Rise. Waters high in SO_4 would, presumably, also have a high Ca/Mg ratio. The observed relation between Ca/Mg ratio and SO_4 concentration is the exact reverse of what would be expected if the ratio were controlled directly by the solution of varying amounts of gypsum (Fig. 15). Waters with low HCO_3/SO_4 ratios have a low Ca/Mg ratio. If the mixing mechanism described above is indeed real, then the high SO_4 groundwater body must also be very high in Mg relative to Ca, high enough to more than compensate for the Ca added to the solution through the solution of soluble gypsum.

Alternatively, by the calcite saturation mechanism, Mg could be concentrated in a water saturated with calcite by the solution of dolomite or of high Mg calcite, both of which are more soluble than pure calcite.

Corrosion Rate

By combining the hydrologic and geochemical data, it is possible to determine the effectiveness of the overall solution process operating in the basin (Bassett and Ruhe, 1973). Specifically, how much rock is being removed from the basin by solution, and, under wht hydrologic conditions is the solution process optimized? Calculations are made for the drainage basin of the Orangeville Rise.

By measuring both the discharge and Ca concentration for each sample from the Orangeville Rise (Fig. 8), it is possible to calculate an instantaneous Ca mass flow rate (gCa/minute). The Ca mass flow rate may be used as an index of the rate of solution and is directly proportional to discharge (Fig. 16).

These data point out the interesting and not at all obvious fact that, as discharge increases, solution rate increases. As discharge at the Orangeville Rise increases from 10 to 100 cfs, the total amount of Ca removed from the basin per unit time is increased almost 6-fold. The calculated regression equation for mass flow rate verses discharge (graph, Fig. 16) is used to estimate the amount of Ca and, hence, limestone removed per unit time as a function of discharge at the rise (Fig. 16).

Mean-monthly discharges at the Orangeville Rise were calculated for each month from June, 1972 through May, 1973. Given the regression equation (Fig. 16) and the mean-monthly discharge, it is a simple matter to calculate the total amount of Ca removed from the basin by month (Table 2, Fig. 17). It is assumed that all the Ca was derived from $CaCO_3$. If an average $CaCO_3$ content of 2.45 g/cm^3 of limestone is assumed (allowing for an average of 10% insoluble residue), it may be calculated that 4.90×10^3 cubic meters of limestone were dissolved from the drainage basin of the Orangeville Rise during the period June, 1972 to May, 1973, and that about 73 percent of this amount was removed during the six-month (non-growing) season from November to April.

It is obvious that the availability of water is the principal factor governing the rate of solution. From water-balance calculations during the summer months, only 10 to 20 percent of the total precipitation in the basin actually becomes recharge to the karst drainage systems. Comparatively low solution

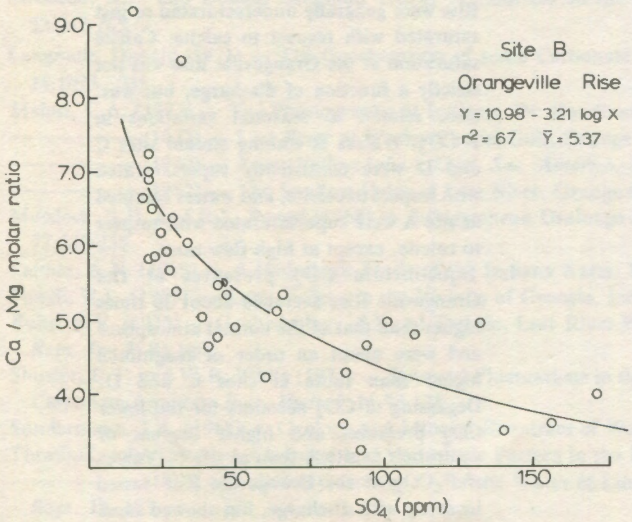


Fig. 15. Graph of Ca/Mg molar ratio vs. SO_4 concentration at the Orangeville Rise.

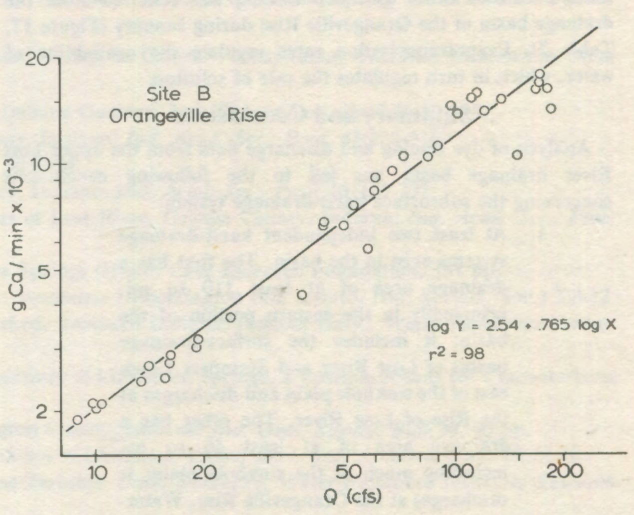


Fig. 16. Graph of Ca mass flow rate vs. discharge at the Orangeville Rise.

TABLE II. Mean-monthly discharge, Ca mass flow rate, and monthly Ca mass discharge from the Orangeville Rise.

Month	Mean discharge (cfs)	Mean Ca flow rate (g Ca/min. x 10^3)	Total Ca dissolved (g x 10^8)
June, 1972	22.8	3.81	1.65
July	24.8	4.05	1.81
August	32.4	4.97	2.22
September	18.1	3.18	1.38
October	15.0	2.76	1.23
November	82.1	10.1	4.37
December	130.0	14.4	6.42
January, 1973	108.5	12.5	5.59
February	86.8	10.6	4.26
March	147.1	15.9	7.06
April	163.9	17.2	7.41
May	87.5	10.6	4.74
TOTAL			48.10

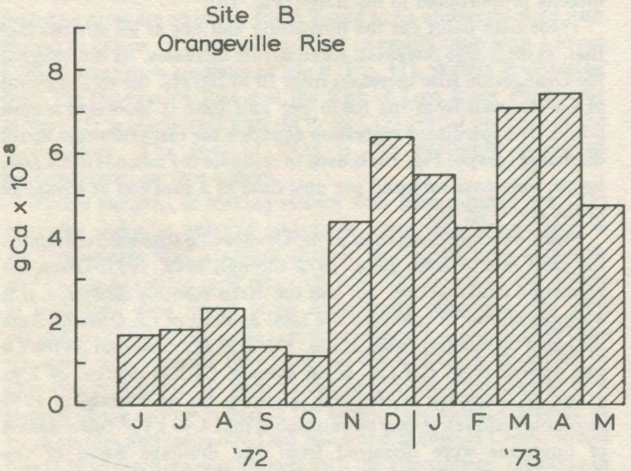


Fig. 17. Graph of total monthly Ca mass discharge at the Orangeville Rise, June, 1972 to May, 1973.

rates for the summer months are not merely a function of lack of precipitation, however, for rainfall during the months of July and August was actually greater than the mean-monthly value.

Although roughly equivalent amounts of water in the form of precipitation were available for solution in September and January, more than four times as much $CaCO_3$ was removed from the drainage basin of the Orangeville Rise during January (Figure 17, Table 2). Evapotranspiration rates regulate the availability of water, which in turn regulates the rate of solution.

Summary and Conclusions

Analysis of dye tracing and discharge data from the upper Lost River drainage basin has led to the following conclusions concerning the subsurface karst-drainage system:

1. At least two independent karst-drainage systems exist in the basin. The first has a drainage area of at least 110 sq mi, principally in the eastern portion of the basin; it includes the surface-drainage basins of Lost River and Stampers Creek east of the sinkhole plain and discharges at the Rise of Lost River. The other has a drainage area of at least 46 sq mi, including much of the sinkhole plain; it discharges at the Orangeville Rise. Water-balance calculations are in good agreement with this figure. Subsurface drainage divides are coincident with surface drainage divides, with minor exceptions.
2. Flow velocities through the major conduit systems of the basin may exceed 5.5 mi/day. These high velocities explain, in part, the rapid response of the Orangeville Rise to rainfall. Dye tracing has shown that the Orangeville Rise is directly connected to several large sinking streams in the Crawford Upland and on adjacent parts of the sinkhole plain.
3. Values for minimum-monthly discharge per unit area show that the drainage basin of the Orangeville Rise supports a substantially higher base flow than does the basin as a whole. This is believed due to the higher infiltration and greater degree of groundwater storage in the karst portion of the basin.

Discharge-water-chemistry relations, as well as ion ratios, equilibrium CO_2 partial pressures, and calcite saturation data, were used to define the geochemical regimen of waters in the basin. Major features of the geochemical regimen are:

1. Waters sampled at sites A and B (downstream gage and the Orangeville Rise) showed chemical variability that was correlated with discharge, as expressed by equations of the general form, $Y = a - b \log X$ or $Y = a X^b$. In general, the chemistry of the waters at the Orangeville Rise changes from basically a $CaCO_3$ composition at high flow rates to a high Mg-Ca-CO_3-SO_4 composition at low flow rates. Ca/Mg and HCO_3/SO_4 molar ratios vary directly with discharge, but Ca/Mg ratios vary inversely with SO_4 concentration.
2. Concentrations of major dissolved chemical species at sites C and D (Sink of Lost River and Stampers Creek sink) show little correlation with discharge. Variations in alkalinity, total hardness, and calcium hardness were much less at site C than at either site A or B.
3. Waters discharging from the Orangeville Rise were generally undersaturated or just saturated with respect to calcite. Calcite saturation at the Orangeville Rise was not strictly a function of discharge, but was, also, related to seasonal variations in P CO_2. Waters at sinking stream sites C and D were consistently supersaturated with respect to calcite, and waters sampled at site A were supersaturated with respect to calcite, except at high flow rates.
4. Equilibrium CO_2 pressures at the Orangeville Rise averaged about 30 times higher than that of the normal atmosphere and were about an order of magnitude higher than those at sites C and D. Degassing of CO_2 accounts for the lower CO_2 pressures and higher degrees of saturation at site A than at site B. Values of P CO_2 at the Orangeville Rise varied inversely with discharge, but showed some degree of seasonal dependence.

The geochemical regimen of the basin is highly influenced by the hydrologic regimen. During the winter months, when recharge from sinking streams is generally high, the chemical character of the water at site B is similar to that at sites C and D. During the summer months, water stored in the karst becomes a dominant control of the base flow chemistry.

The changes in chemistry observed with changes in discharge at the Orangeville Rise may be attributed to one or both of two mechanisms: First, a mixing mechanism is possible, where two geochemically and hydrologically dissimilar waters are mixed in varying proportions, depending on discharge through the major conduit systems. One water is a saturated, low P CO_2, low SO_4, $CaCO_3$ type, derived from sinking streams, and draining rapidly through large conduits; the other is a Ca-Mg-CO_3-SO_4 water, saturated with respect to calcite, and at a higher P CO_2 ($10^{-1.7}$ to $10^{-1.8}$ atm.), perhaps representing diffuse infiltration from the sinkhole plain.

By the second mechanism, the same chemical changes could be brought about by the gradual saturation of conduit water as it moves through the system more slowly at lower discharges. The water becomes saturated with respect to calcite but remains undersaturated with respect to dolomite, high-Mg calcite and gypsum, hence concentrating Mg and SO_4 relative to Ca.

A few observations concerning groundwater flow and solution mechanisms in karst terrain may be made: First, it can be seen from the dye-tracing data that many square miles of a karst-drainage system may drain to a single well-defined spring or rise. Water may move through the subsurface at very rapid rates (several miles per day). A pollutant introduced into the groundwater system anywhere in the basin may be expected to move rapidly through the subsurface drainage conduits. Sewage effluent pumped into the ground at Orleans reaches the Orangeville Rise within 25 hours.

Secondly, karst springs, as exemplified by the Orangeville Rise, show pronounced variations in chemistry that are related to discharge. As discharge decreases, the chemical character of the water approaches that of the major body of groundwater in the area. It is, therefore, thought that the sampling of karst springs at base flow stage might be a useful technique for evaluating groundwater quality in a karst basin. High concentrations of a chemical species in spring waters might be expected to be associated with similarly high concentrations in water wells in the area.

With regard to solution mechanisms, the fact that large sinking streams in the Lost River area are saturated with respect to calcite seems to contradict some of the older thoughts on the development of subsurface drainage conduits in karst. Surface water saturated with respect to calcite cannot be expected to enter the subsurface and dissolve $CaCO_3$; yet, the presence of large, open solution conduits associated with the sinking streams indicates that such solution does occur. For the surface waters to become solutionally aggressive, CO_2 must be introduced into the system. The fact that spring waters have a higher CO_2 content than the surface waters indicates that this is probably the case, and that the addition of CO_2 to the subsurface water is a major factor in the development of underground solution conduits.

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Investigations of Sinkhole Stratigraphy and Hydrogeology, South-Central Indiana

Robert D. Hall *

ABSTRACT

Bedrock basins dissolved in the St. Louis and Ste. Geneviève limestones of Middle Mississippian age in the Mitchell Plain of south-central Indiana are partly filled with surficial deposits whose thicknesses may total several tens of feet. These deposits consist of a variety of silty and clayey sediments, including, from oldest to youngest: 1) calcareous, laminated muds in the lower part of the basin; 2) red clay in the upper part of the basin, grading laterally to red silt above the calcareous mud in the lower part of the basin; 3) chert gravel over the red clays and silts; 4) loess in the upper part of the basin; and 5) yellowish-brown silts above the red silt in the lower part of the basin. All the deposits appear to be of sedimentary origin, including the red clays and silts often called *terra rossa*.

The water table in Sinkhole A does not respond with a change in elevation until the materials above the clay are saturated, under which conditions the water table normally responds with a lag time of about 2 weeks. In summer, heavy cloudbursts probably supply some intermittent runoff to the basin center which, in turn, results in some temporary downward flow through the basin silts to the water table. This type of flow can also occur in winter, but saturation of the loess during that season also favors flow at the top of the clay toward the basin center and probably downward flow through the clay along fractures.

Introduction

The surficial deposits, geomorphology, and hydrology of the Lost River Watershed in south-central Indiana have been the subject of a study by the Water Resources Research Center at Indiana University since July 1971, under a research grant by the Office of Water Resources Research. As part of the project, a detailed investigation of sinkhole stratigraphy and hydrogeology was undertaken by the writer in the summer of 1972. Results of the study of stratigraphy and origin of surficial deposits in the karst area have been presented previously (Hall, 1973, 1976) and will be summarized briefly here. Preliminary results of the hydrogeology study are also presented here.

Much of the Lost River Watershed lies in the Mitchell Plain, where sinkholes occur in great numbers (Fig. 18). Bedrock basins dissolved in the St. Louis and Ste. Genevieve Limestones of Middle Mississippian (Meramecian) age are partly filled with surficial deposits, whose thicknesses may total several tens of feet. The stratigraphy of these deposits was studied in three sinkholes (Fig. 19). Deposits and soils penetrated by coring were described, and the particle-size distribution of selected samples was determined. Wells and neutron probe access tubes were installed in Sinkhole A.

Geomorphology

Sinkhole A is shaped like a broad bowl and is slightly elongated in th east-west direction (Fig. 19). It includes an area of about 6.5 a and has a maximum relief of 27 ft. Sinkhole B (8.0 a) and Sinkhole C (2.0 a) are irregularly shaped. Coring in Sinkhole A has revealed that the bedrock is much more irregular and of much greater relief than the overlying land surface (Fig. 20).

Stratigraphy of Surficial Deposits

From oldest to youngest, the stratigraphic units of the sinkhole are (Fig. 20): 1) sandy, calcareous muds covering the limestone bedrock in the lower part of the basin; 2) red clay in the upper part of the basin, grading laterally to red silt above the calcareous mud in the lower part of the basin, (both facies commonly known as *terra rossa*); 3) chert gravel over the red clays and silts; 4) loess in the upper part of the basin; and 5) yellowish-brown silts in the lower part of the basin that are derived from loess stripped from the slopes above.

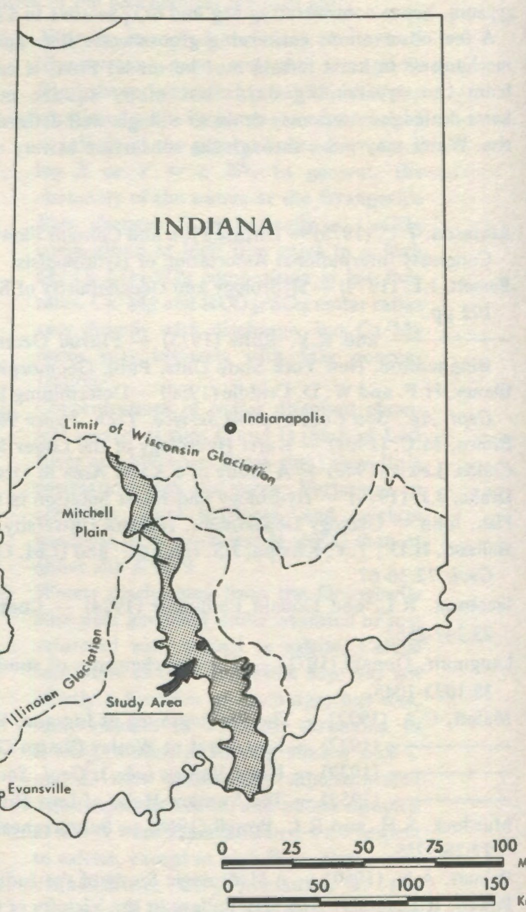

Fig. 18. Index Map.

The nature and distribution of the surficial deposits suggest the filling of bedrock basins by a variety of sediments, including the red clay and red silts that have previously been called *terra rossa* and implied to be of residual origin. In this case, however, all of the deposits appear to be of sedimentary, not residual, origin.

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Hydrology of Sinkhole A

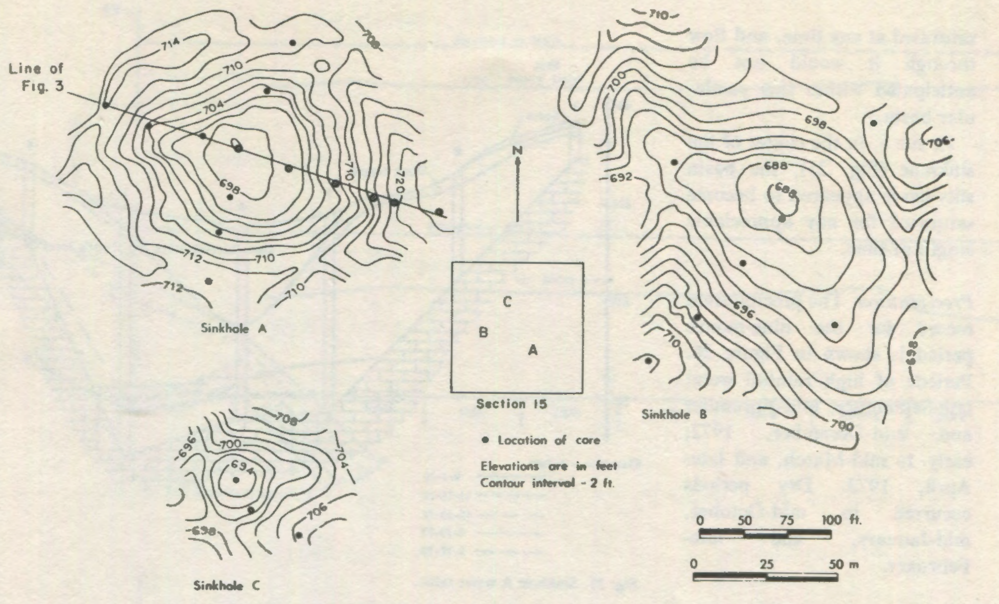

Fig. 19. Topographic maps of Sinkholes A, B, and C.

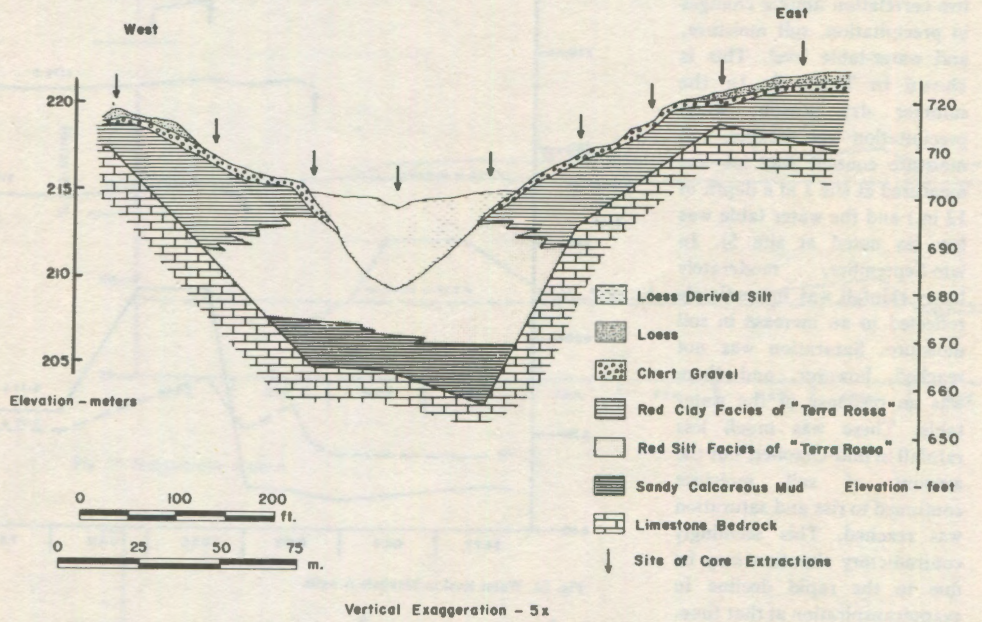

Fig. 20. East-west cross section of Sinkhole A.

This preliminary report on the hydrogeology of Sinkhole A covers the nine-month period from September 1972 to June 1973, after the installation of wells and neutron-probe access tubes. During this time, water levels in the wells were periodically checked, and soil moisture was monitored with the neutron probe. Precipitation also was recorded. These data are presented graphically in the figures and the table, and some tentative conclusions are drawn about the relationship of precipitation, soil moisture, evapotranspiration, and changes in the water table. A tentative interpretation of the flow of water through the surficial materials of the sinkhole is also included.

It should be noted that the data upon which these interpretations are based were collected during an interval that included an unusually wet fall and an unusually dry midwinter.

The Sinkhole Water Table. Figure 21 shows the position of the water table in the sinkhole for five dates, beginning with the end of the 1972 dry season and extending to the beginning of the 1973 dry season. In early September, 1972, the water table was observable only in wells 5 and 6 and the gradient of the water table was from 6 to 5. In the middle of November, during an unusually wet fall, the water table was elevated in well 6, and water was also present in well 3 (but not in well 4). It is not known whether the water in wells 3 and 5 was connected through the bedrock or if, in both cases, the water was perched on the bedrock and unconnected. Indeed, it is not known if any of the water in the wells has a relationship to a water table in the bedrock. By the middle of December, the water table in the sinkhole was at a much higher level, especially in well 5, where there had been a rise of 20 feet in one month. Water was also present in wells 2 and 4 for the first time. The gradient of the water table had become reversed and now sloped toward well 6. The water table maintained the same overall position through the rest of the winter and spring. A maximum height of the water table in the basin was reached in late April 1973; at that time, water was present in well 7 for the first and only time during the period of measurement. This high level was followed by an abrupt decline, during which the water table fell to approximately the same level as it had had in early September, and the original gradient was re-established.

A record of water levels in the sinkhole is given in Figure 22. The fluctuations in wells 5 and 6 are similar, although the changes in 5 are more dramatic. As noted previously, water was present in well 3 before it appeared in well 4, although water remained longer in well 4. A water level was present only very briefly in well 2.

Soil Moisture. Changes in soil moisture are noted first at site 1 on the eastern perimeter of the sinkhole (Fig. 23). The volumetric percentage of moisture is given at three depths: at 12 in. in the loess, at 42 in. in the red clay facies of the terra rossa, and at 84 in. in the red clay. The horizontal dashed lines indicate the value of saturation at particular depths with respect to hygroscopic and capillary water. (The term "saturation" is used throughout the following discussion only in this sense.) The loess became saturated in mid-October 1972 and remained so until early June 1973. During this time, water flow could occur through the loess toward the underlying red clay. However, the terra rossa did not become saturated at any time, and flow through it would not be anticipated within this particular basin.

At site 6, in the center of the sinkhole (Fig. 25), the basin silts never appeared to become saturated for any appreciable length of time.

Precipitation. The precipitation record for the nine-month period is shown in Figure 26. Periods of high rainfall were: late-September, mid-November and mid-December, 1972; early- to mid-March, and late-April, 1973. Dry periods occurred in mid-October, mid-January, and late-February.

Hydrologic Events. The writer has made a tentative, qualitative correlation among changes in precipitation, soil moisture, and water-table level. This is shown in Table 3. In the summer dry season, when precipitation was meager, soil moisture content was low (as measured at site 1 at a depth of 12 in.) and the water table was low (as noted at site 5). In late-September, moderately heavy rainfall was immediately reflected in an increase in soil moisture. Saturation was not reached, however, and there was no response of the water table. There was much less rainfall in mid-October, but the amount of soil moisture continued to rise and saturation was reached. This seemingly contradictory situation may be due to the rapid decline in evapotranspiration at that time.

Heavy rainfall again occurred in mid-November and soil moisture continued to increase, but, at that time, there was no rise in the water table until about two weeks later, during a period when there actually was less rainfall. This observation suggests that, after saturation has occurred, there is about a two-week lag between changes in precipitation and response of the water table.

A similar response is seen when a period of low precipitation in mid-January is reflected in a drop in the water table in late-January. However, the dry period of late-February does not

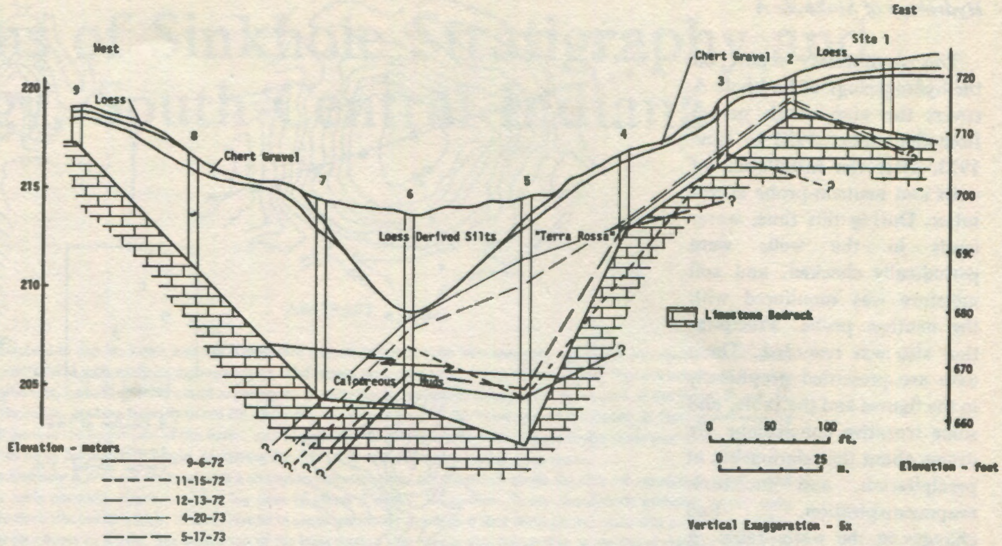

Fig. 21. Sinkhole A water table.

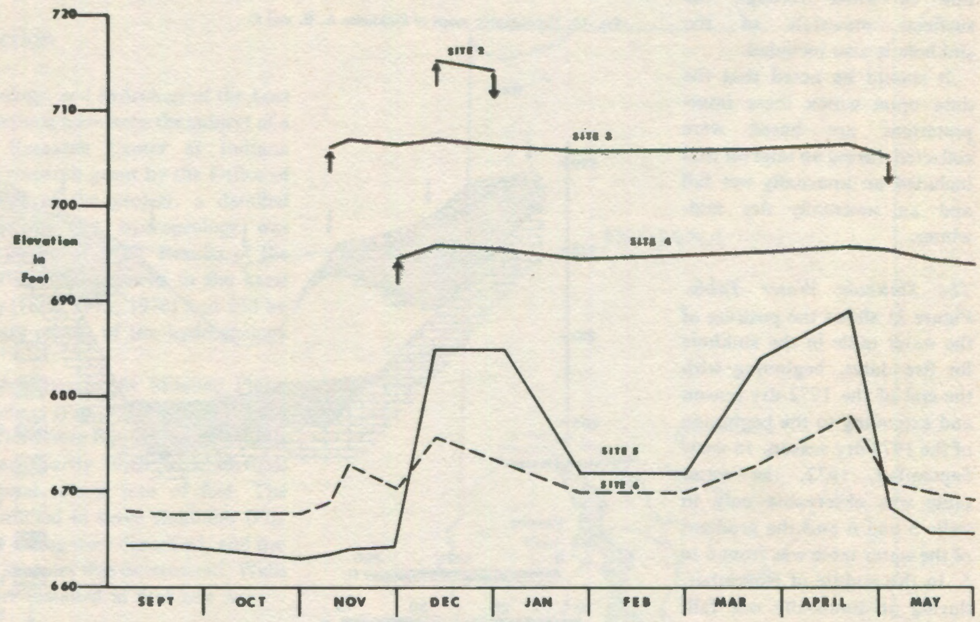

Fig. 22. Water level in Sinkhole A wells.

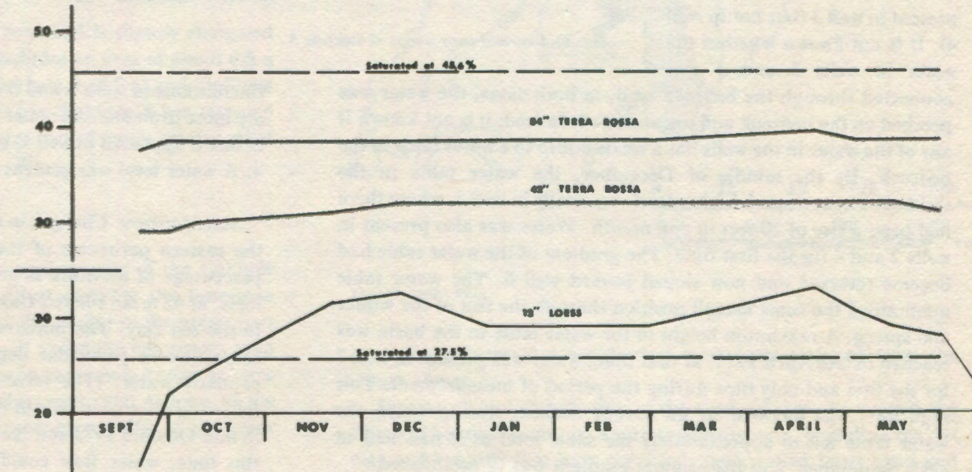

Fig. 23. Soil moisture at site 1.

seem ever to have been reflected in a drop in the water table. Perhaps the drop was missed in the sampling, or it might have been obscured by a freezing of the ground which occurred at that time.

An increase in rainfall during the early spring was reflected in a greater amount of soil moisture and a higher water table. During the time of heaviest precipitation in late April, however, there was a *decrease* in soil moisture, probably owing to a sharp increase in evapotranspiration. This trend continued through early June as soil moisture continued to decline, and the water table fell to about the same level as that of early September.

In summary, changes in precipitation appear to be reflected almost immediately in changes in soil moisture, except when there are over-riding effects of evapotranspiration. There was no response of the water table until saturation of the loess occurred with respect to hygroscopic and capillary moisture, and, after this, there was usually about a two-week lag between changes in precipitation and fluctuation of the water table.

Flow. The writer's interpretation of groundwater flow conditions during the summer is shown in Figure 27. The only flow that occurs is runoff toward the basin center, which then collects and funnels into animal burrows, primarily those of ground hogs, which are numerous in the silty materials of many sinkholes. Although the basin silts were always unsaturated when moisture readings are taken, it seems likely that there must be a route for this water through the silts, possibly accompanied by an almost instantaneous and very temporary saturation. Water is probably conducted to the water table at the points where the gradient indicates flow out of the sinkhole toward sites 5 and 7.

In winter (Fig. 28), the same type of flow may occur, but, at that time, the loess is saturated, and subsurface flow occurs

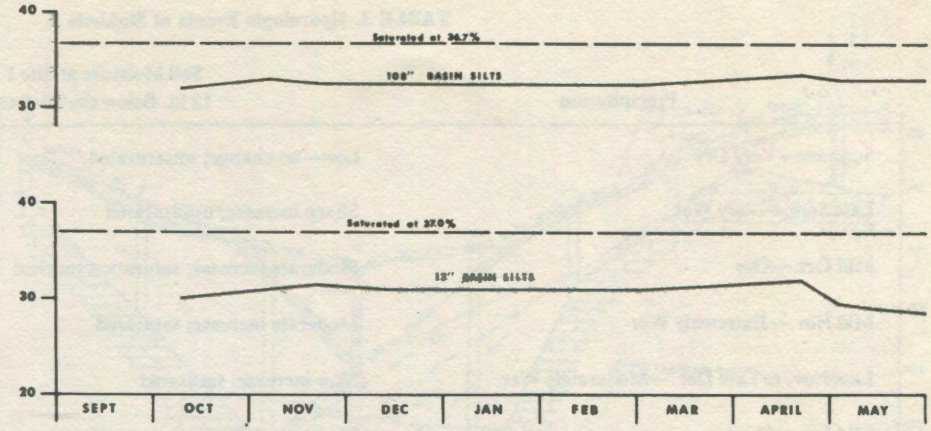

Fig. 24. Soil moisture at site 5.

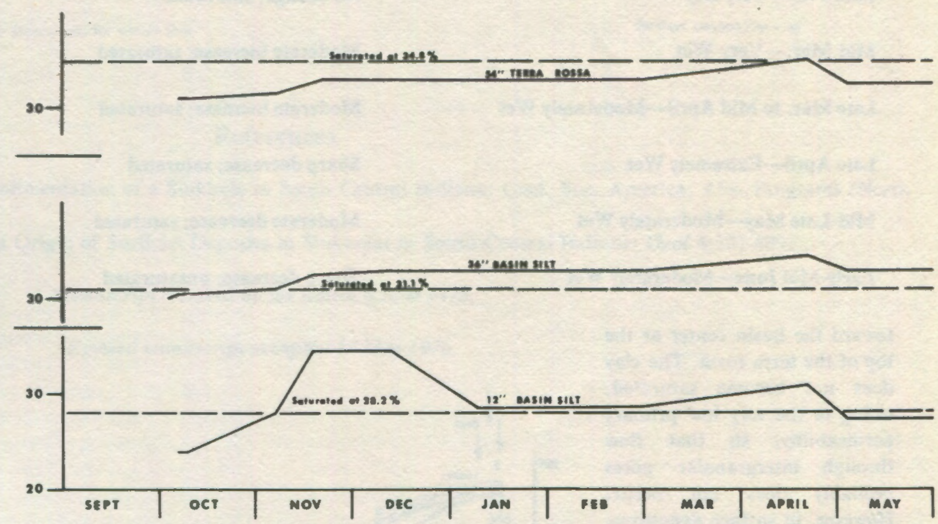

Fig. 25. Soil moisture at site 6

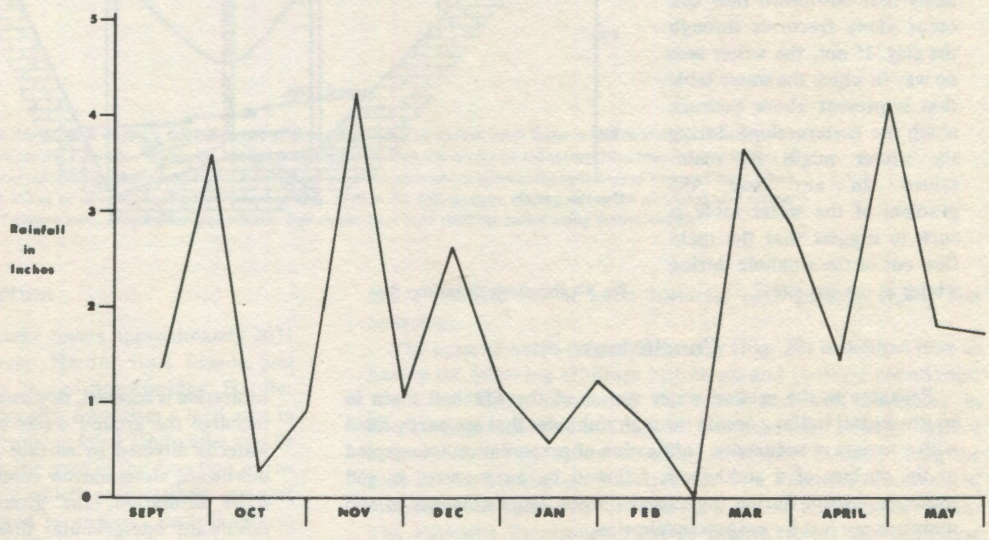

Fig. 26. Rainfall in sinkhole area.

TABLE 3. Hydrologic Events at Sinkhole A

Precipitation	Soil Moisture at Site 1 12 in. Below the Surface	Water Table Elevation (ft) at Site 5
Summer—Very Dry	Low—no change; unsaturated	± 665
Late Sept.—Very Wet	Sharp increase; unsaturated	± 665
Mid Oct.—Dry	Moderate increase; saturation reached	< 665
Mid Nov.—Extremely Wet	Moderate increase; saturated	665
Late Nov. to Late Dec.—Moderately Wet	Slow increase; saturated	685
Mid Jan.—Dry	Moderate decrease; saturated	685
Late Jan. to Mid Feb.—Moderately Wet	No change; saturated	673
Late Feb.—Very Dry	No change; saturated	673
Mid Mar.—Very Wet	Moderate increase; saturated	685
Late Mar. to Mid April—Moderately Wet	Moderate increase; saturated	690
Late April—Extremely Wet	Sharp decrease; saturated	670
Mid-Late May—Moderately Wet	Moderate decrease; saturated	± 665
Early-Mid June—Moderately Wet	Sharp decrease; unsaturated	± 665

toward the basin center at the top of the terra rossa. The clay does not become saturated, owing to the very low primary permeability, so that flow through intergranular pores probably does not occur. However, in surface exposures, the clay contains many fractures, and, because those are also seen in cores, it would seem likely that downward flow can occur along fractures through the clay. If not, the writer sees no way in which the water table that is present above bedrock along the eastern slope during the winter might be maintained. In any case, the gradient of the water table is such to suggest that the main flow out of the sinkhole during winter is toward site 7.

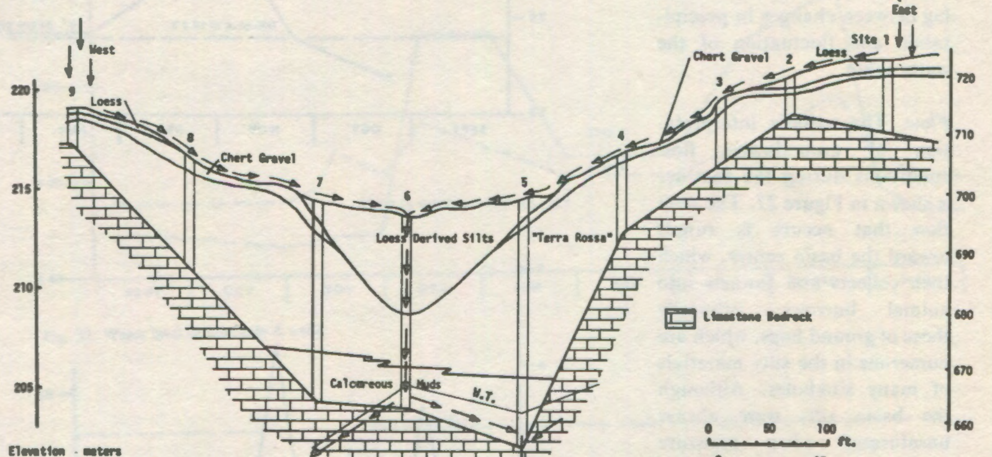

Fig. 27. Hypothesis for summer flow.

Conclusions

Recharge to the ground water system of the Mitchell Plain in south-central Indiana occurs through sinkholes that are partly filled with a variety of sediments. Infiltration of precipitation intercepted at the surface of a sinkhole is followed by an increase in soil moisture, except during the summer when large amounts of soil moisture are lost by evapotranspiration.

The key factor in directing subsurface flow is the saturation of the loess with respect to hygroscopic and capillary moisture. Until this saturation is reached, downward flow through the loess sufficient to recharge the ground water system cannot occur, in which case, water is directed by surface runoff to the basin center and then downward along narrow conduits which begin as animal burrows. After saturation, the ground water system is recharged by downward intergranular flow through the loess and flow along fractures in the "terra rossa." Only in this way can a water table be maintained above bedrock along the slopes of the sinkhole.

The study described above has contributed to our knowledge of sinkhole stratigraphy and hydrogeology, but it should be considered as only a preliminary step. Additional data should be collected, over a longer period of time. At some future date, it should be possible to express the hydrogeologic relationships of sinkholes quantitatively and as part of a water-balance relationship for the entire Lost River Watershed.

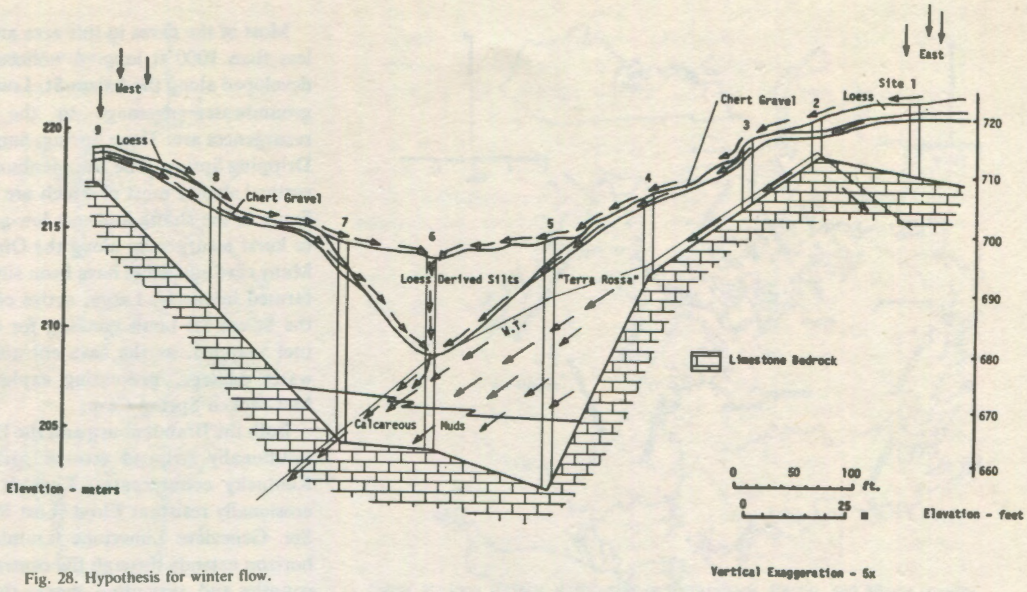

Fig. 28. Hypothesis for winter flow.

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Karst and Cave Distribution in North-central Kentucky

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ABSTRACT

The main object of this paper is to give a brief summary of prominent cave and surface karst features in Breckinridge, Hardin, and Meade counties, north-central Kentucky. Within the tri-county locality, more than 620 caves and 110 springs have been indexed as part of a long-term regional cartographic and hydrological study. Particular reference and space is given to the hydrology of Sinking Creek, which drains approximately 252 sq mi of surface and subsurface land area. Within this hydrosystem, Boiling spring (an alluviated cave spring) drains more than 147 square miles. Subsurface groundwater flow routes have been partially traced using fluorescein dye.

Introduction

The karst of north-central Kentucky covers approximately 2031 sq mi of land in Breckinridge, Grayson, Hardin, Hart, Meade, and Larue counties. Only the tri-county area of Breckinridge, Hardin, and Meade will be discussed. This locality comprises a land area in excess of 1475 sq mi. Sequentially evolving karst landscapes can be observed extending from the east to the west in the direction of the regional dip, representing areas in which progressively younger sedimentary rocks have been removed by erosion. Each landform has a discrete set of karst features, cave passage types, and hydrology.

The karst of north-central Kentucky (Fig. 29) is defined here as having the following arbitrary hydrologic and geologic boundaries: The Ohio River determines the north and west boundary in Breckinridge and Meade Counties. Rolling Fork, a major surface stream, forms the eastern boundary in Hardin County, along the base of the Muldraugh Escarpment. The Pottsville Channel, of northern Hart County, delineates the southern limits of the region. The Pottsville Escarpment, low in profile, forms the southwest boundary, which encloses portions of northern Grayson and western Breckinridge counties. Specific features described in the text are located in figures 2 and 3.

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Figure 29. Map of the Mississippian plateau region in northern Kentucky, showing the regional relationship of prominent escarpments and solutional surfaces.

Geomorphic Setting

The Mississippian plateau region is a subdivision of the Interior Low Plateaus. From east to west in Hardin and Meade counties and through the interior of Breckinridge County, there are developed two major geomorphic units separated by three prominent east-facing escarpments (Fig. 29).

The Muldraugh Escarpment delineates the eastern boundary of the Mississippian plateau region along the Ohio River and Rolling Fork. In the vicinity of the Fort Knox Military Reservation, this cuesta rises 280 ft above the floodplain of the Ohio River. This elevation corresponds roughly to the level of the Lexington Peneplain. On the back slope of the cuesta is a sinkhole plain called the Pennyroyal Plateau. This is the geomorphic and stratigraphic equivalent of the Mitchell Plain of southern Indiana.

The Pennyroyal Plateau is locally divided into two sections, called the Brandenburg and Elizabethtown plains. The boundary between the two sinkhole plains is an inverted clastic topographic feature, called the Bethel (Brownsville) Channel. This channel diagonally crosses the sinkhole plain from south-central Indiana, along the Muldraugh Escarpment, into Kentucky, through Tip Top and Flaherty, to Losenville Valley east of Big Spring, where the channel intercepts and merges with the Chester Escarpment.

Southeast of the channel, is the Elizabethtown Plain (Sauer, 1927), a solutional surface developed on the updip slope of the Muldraugh Escarpment. This plain spans the length of eastern Hardin County. It contains a distinct zone of ephemeral surface streams draining the upland sinkhole plain northward to Mill Creek and, also, southward toward Nolin River. These stream valeys are dendritic. Most have steep profiles. There are few sinking streams within the interfluvial areas.

Northwest of the Bethel Channel, is developed a dissimilar karst landscape called the Brandenburg Plain. This territory lacks fluvial texture, its doline fields cover a greater surface area, there are re-entrant stream valleys tributary to Ohio River, the karst base-level drainage is 120 to 150 ft below the surface, and subsurface drainage trends radially to exsurgences flanking the Brandenburg Plain. In both landscapes, karst is developed on the Salem, St. Louis, and, to some extent, the Ste. Geneviève limestones.

Most of the caves in this area are small in lateral extent, usually less than 1000 ft long. A noticeable perennial spring horizon is developed along the Salem-St. Louis contact and accounts for most groundwater drainage to the surface. Examples of these resurgences are: Tioga Spring, Sanders Spring, Falling Spring, and Dripping Spring. The Brandenburg Plain is characterized by many vertical shafts, most of which are between 60 and 150 ft in depth. Some of the shafts intersect low-gradient trunk passages that lead to karst resurgences along the Ohio River bluffs (George, 1972b). Many cave entrances have been silted up because the area has been farmed intensely. Large, active cave systems are developed along the Salem-St. Louis contact; for example, Doe Run Cave System (not mapped, as the cave entrance has collapsed into the main water passage, preventing exploration), Grahamton Cave, and McCracken Spring Cave.

Both the Brandenburg and the Elizabethtown sinkhole plains are solutionally stripped erosion surfaces exactly like their central Kentucky counterparts. Each is structurally controlled by the erosionally resistant Elrod (Lost River) Chert horizon in the basal Ste. Geneviève Limestone (Quinlan and Pohl, 1968). The chert horizon extends through the central portions of Hardin and Meade counties and, very often, marks the stratigraphic junction of input for ephemeral sinking streams. The sinkhole plain has low relief, usually less than 50 ft, which causes a characteristic rolling topography of intense sinkhole development. Outliers occur on the sinkhole plain and, generally, support ephemeral sinking streams that drain the upland slopes and sink near the bases of the outliers. Shaft development is common within outliers capped by the Mooretown (Bethel) and Sample sandstones. Many large karst springs border master, base-level streams that have cut across the sinkhole plain; for example, Nolin River, Rough River, Otter Creek, Doe Run Creek, and Sinking Creek. Examples of major karst resurgences of second magnitude (10-100 cfs) are: Head of Rough Spring, White Mill Spring, and McCracken Spring. Cave passages are commonly large in volume, and many carry active, low-gradient streams through trunk passages ending at resurgences. Examples of cave springs that drain into the Ohio River are: Mint Spring, Head of Doe Run, Blue Spring, Morgan Cave Spring, and Daniel Boone Spring Cave. All of these are developed along the Salem-St. Louis contact. Sandwiched between Otter Creek and Mill Creek is an elevated tract of land on the sinkhole plain that contains a distinct zone of sinking streams. They flank and, in places, interfinger with zones of through-flowing surface streams. Caves are sparse and are generally small in cross section and length, although Grahamton Cave, which is also developed along the Salem-St. Louis contact, is an exception.

The Chester Escarpment is a prominent east-facing topographic feature, long recognized as a compound solutional cuesta determined by differential weathering of the Ste. Geneviève and Paoli limestones, the Mooretown Sandstone, Beaver Bend Limestone, and the Sample Sandstone (Dicken, 1935a). This cuesta rises 180 ft above the level of the sinkhole plain. Ephemeral streams drain from distal ridges that extend outward from the escarpment and sink in swallow holes near the base of the escarpment, or on the sinkhole plain. The ephemeral sinking streams do not occur in linear zones, as in the Central Kentucky Karst. The face of the escarpment is dissected into re-entrant uvalas that act as intake points for surface upland drainage into the karst groundwater system. Some of the largest uvalas occur in eastern Breckinridge County; for instance, Corners Uvala (4 mi west of Big Spring), Big Spring Uvala (near Big Spring, at the tri-county intersection) and the impressive Nortons Valley Uvala (4.5 mi northeast of Hardinsburg). The highest density of cave and spring entrances is located along the Chester Escarpment. There is a decrease in density east and west of the escarpment (Fig. 31).

The Chester Upland (equivalent to the Crawford Upland of southern Indiana) is situated on the back slope of the Chester

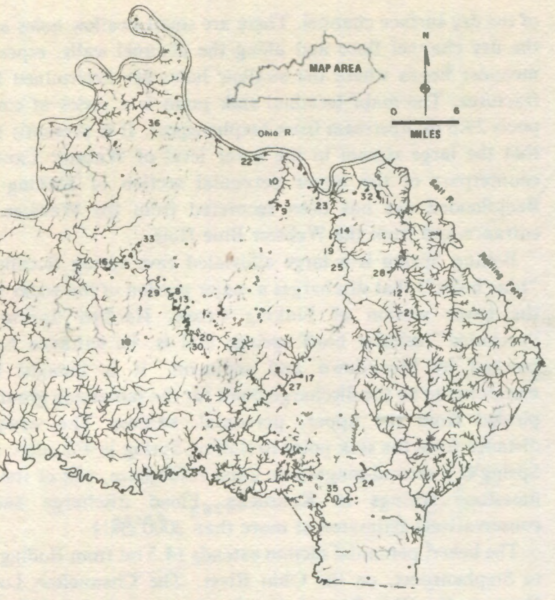

Figure 30. Stream and spring density map of Breckinridge, Hardin, and Meade Counties, Kentucky. About 110 springs are indicated on the map. 2 = Blue Fork; 3 = Blue Spring; 4 = Boiling Spring; 5 = Clifton Church karst head; 7 = Daniel Boone Spring Cave; 8 = Dents Bridge; 9 = Doe Run Cave; 10 = Doe Run; 11 = Dripping Spring; 12 = Falling Spring; 13 = Flat Rock Spring; 14 = Grahamton Cave; 16 = Hardins Spring; 17 = Head of Rough Spring; 18 = Lost Run; 19 = McCracken Spring Cave; 20 = Milburns Blue Hole; 21 = Mill Creek; 22 = Mint Spring; 23 = Morgan Cave Spring; 24 = Nolin River; 25 = Otter Creek; 27 = Rough River; 28 = Sanders Spring; 29 = Sinking Creek; 30 = Stony (Rocky) Fork and Stony Fork Spring; 32 = Tioga Spring; 33 = Webster Blue Hole; 35 = White Mill Spring; 36 = Wolf Creek. The location of other caves is shown on Figure 31.

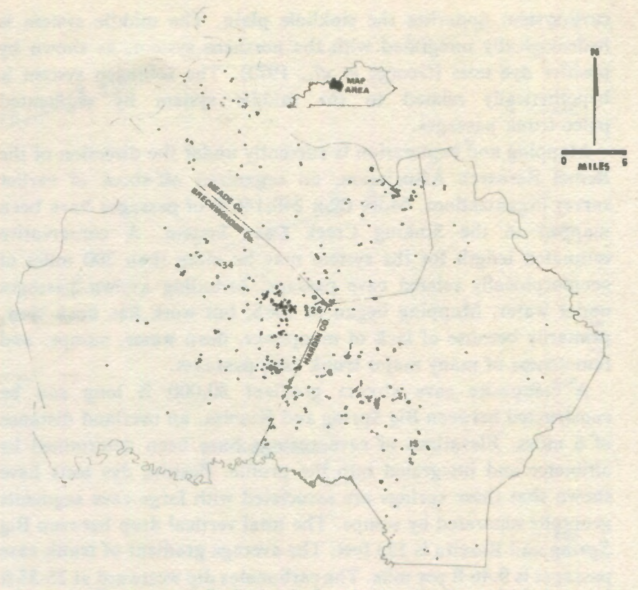

Figure 31. Cave entrance distribution in Breckinridge, Hardin, and Meade counties, Kentucky. The highest density of cave entrances marks the position of the Chester Escarpment. About 450 cave entrances are indicated on this map. 1,1 = Belt-Whitehead Cave system; 6 = Constantine Saltpeter Cave; 15 = Great Wonderland Cavern; 26 = Roaring River Cave; 31 = Sweet Potato Cave; 34 = Webster Cave.

Escarpment. This upland, like the sinkhole plain, is a stripped erosion surface. It is developed on the Sample Sandstone. Most hills in the upland are capped by the Big Clifty Sandstone, although the area is also studded by outliers of Pennsylvanian rocks. There is a distinct peripheral area of sinking streams, such as Ivory Hollow, Duncan Valley (both located 2.5 miles south of Custer, Kentucky), and Lost Run, all of which drain into deep uvalas. Karst drainage takes place through master cave passages that intersect surface base-level streams (Sinking Creek, Rough River, and Wolf Creek), thus establishing a hydrological communication between recharge points on the Chester Upland and discharge points on the sinkhole plain. An exceptionally high density of cave entrances is found in this section.

The western halves of Breckinridge and Meade counties are marked by a return to through-flowing surface drainage. Thick carbonate rocks containing large cave passages with active streams are not found in this district. The reason is that the St. Louis, Ste. Geneviève, Paoli and Beaver Bend limestones have dipped below the surface, leaving thinner and younger rock units (Haney, Reelsville, and Glen Dean limestones) on the surface. Most of the caves in the Haney and Reelsville limestones are small in cross section and in length. Few caves are known from this district because of its remoteness.

Occurence of Cave Entrances

In the three-county area, there is a northwest-trending linear cluster of cave entrances developed along the Chester Escarpment (Fig. 31). Here, the highest known density of known caves is lowest in eastern Hardin County, where the Muldraugh, Harrodsburg, and Salem Limestone units crop out, because most of this area is contained in the Fort Knox Military Reservation and is, therefore, inaccessible to explorers. There are also few caves known in western Breckinridge and Meade counties. Much additional field work is needed in these two areas.

Table 4 indicates the distribution of caves and entrances that were known as of May, 1973.

TABLE 4. Distribution of known caves in the karst of north-central Kentucky.

County	Number of caves	Number of entrances	Number of mapped caves
Breckinridge	296	312	57
Hardin	156	168	21
Meade	169	171	35
Total:	621	651	113

Of all the known cave entrances in this district, 47.22% are in the Ste. Geneviève Limestone. The Paoli Limestone contains 21.72% of the total number. This is a greater number than would be expected from the slight thickness of the Paoli, primarily because of the high number of vertical shafts associated with the Mooretown Sandstone. Only 11.02% of the caves are developed in the St. Louis Limestone. Cave development above the Paoli Limestone and below the St. Louis Limestone accounts for 20.04% of the caves. These data were obtained from a total of 411 cave entrances whose stratigraphic positions could be located precisely.

The Sinking Creek Cave System

Jillson (1927, p. 121) has said that extensive caves could not occur between Munfordsville and Brandenburg, because the Ste. Geneviève and Paoli limestones thin northward, become more siliceous, less pure, and do not contain streams sufficiently entrenched to initiate large caves. However, large cave systems are found in the Ste. Geneviève, Paoli, and Beaver Bend Limestones in the Big Spring-Rosetta-Dyer-Webster area and comprise a hydrologically integrated network of cave passages. The Sinking Creek Cave System is segmented into three mapped sections. The Big Bat Cave section is developed beneath the Chester Upland, a few miles south of Big Spring; the middle system is located along the dissected edge of the Chester Escarpment; and the northern cave system underlies the sinkhole plain. The middle system is hydrologically integrated with the northern system, as shown by positive dye tests (George *et al.*, 1973). The southern system is hypothetically related to the middle system by segmented paleo-trunk passages.

Mapping and exploration is currently under the direction of the Bethel Research Association, an organized off-shoot of earlier survey organizations. More than 240,146 ft of passages have been mapped in the Sinking Creek Cave System. A conservative estimated length for the system may be more than 200 miles of geomorphically related cave passage, including known passages under water. Mapping began in 1965, but work has been slow, primarily because of lack of manpower, deep water, sumps, and remoteness of many major trunk cave passages.

A composite cave stream gradient 50,000 ft long can be constructed between Big Spring and Rosetta, an overland distance of 8 miles. Elevations of cave springs have been determined by altimeter and integrated into the profile. Positive dye tests have shown that these springs are associated with large cave segments generally separated by sumps. The total vertical drop between Big Spring and Rosetta is 133 feet. The average gradient of trunk cave passages is 9.46 ft per mile. The carbonates dip westward at 25-35 ft per mile. Cave passages generally trend northwest, with a few trending toward the southwest. Underground stream flow in the system is mostly northwest. Dip and strike of the rocks exert little control over major paths of flow in the system. The hydrologic gradient is more important to the genesis, distribution, and enlargement of these caves than is the influence of rock dip. Cave passages oriented along the strike or dip are a rarity.

Hydrogeology of Sinking Creek

The Sinking Creek hydrosystem drains approximately 252 sq mi of surface and subsurface land area. Ground water is collected mostly from eastern Breckinridge and west-central Meade counties. The boundaries of drainage area were estimated from the position of known karst recharge points and topographic divides, and from cave-passage trends, dye tests, and the distribution and magnitude of springs.

Sinking Creek is developed within carbonate rocks, heads within the back slope of the Chester Escarpment, and flows in a 240 ft-deep karst gorge south of Dents Bridge that is possibly the result of stream piracy coupled with the Locust Hill Fault system. The channel of Sinking Creek is 36.5 mi long, measured from the confluence of Blue Fork and Rocky (Stony) Fork to its mouth at Stephensport, on the Ohio River. In its upper reaches, the stream loses and gains noticeable amounts of surfacewater flow.

The creek can be divided into three main sections. The upper, perennial region is recharged by a suite of perennial springs located on either side of the creek. These springs discharge water from cave passages and account for all base flow to Sinking Creek. Ephemeral recharge is contributeo by runoff from the adjacent valley walls and by springs draining sinkholes, uvalas, and sheet-like solution cavities along the stream bank. There are five second-magnitude (10-100 cfs) resurgences: Stony Fork Spring, Milburns Blue Hole, Conners #1 Spring, Flat Rock Spring, and Fiddle Spring. These springs reach first-magnitude status (more than 100 cfs) during periods of heavy rain. There are 18 additional springs known along the creek bank that contribute water, although their flow has not been measured. They are of third (1-10 cfs) and fourth (<1 cfs) magnitude discharge. The upper section of Sinking Creek is 10 mi long and extends from the main swallow holes downstream from Dents Bridge to the headwater confluence of Blue Fork and Rocky Fork.

The ephemeral section of Sinking Creek is 12.2 mi long and is located in a partly blind valley that transmits water only when the flow exceeds the capacity of the caves, causing sequential flooding of the dry surface channel. There are small swallow holes all along the dry channel floor and along the channel walls, especially at meander bends where the swallow holes are determined by rock fractures. The main terminal sink point is a series of connected pools 29.5 mi upstream from Stephensport. It is presently thought that the large stream in the lower level of Webster Cave is the counterpart of the upper perennial section of Sinking Creek. Backflooded dye has been recovered from the Webster Spring entrance and from the Webster Blue Hole.

Boiling Spring is a large alluviated cave spring (locally called "blue holes") that discharges a major portion of the water forming the lower section of Sinking Creek. Hardins Spring is an "occluded" tubular bluff spring; that is, its entrance has been blocked by breakdown and sediment. It is thought by this researcher to be the discharge point for the remaining groundwater pirated from the upper, perennial, section. The straight-line distance from the sink point to Boiling Spring is 4.3 miles. Boiling Spring is a second-magnitude karst resurgence, one of the largest limestone springs in Kentucky. Flood discharge has been conservatively estimated at more than 2000 cfs.

The lower, perennial section extends 14.5 mi from Boiling Spring to Stephensport, on the Ohio River. The Channelton Lock and Dam, on the Ohio River below Stephensport, now ponds Sinking Creek to a point 2 mi upstream from Sample, Kentucky.

The Boiling Spring karst watershed drains about 147 sq mi; it includes the 69 sq mi of the upper, perennial, section of Sinking Creek and about 17 sq mi in the Rocky Fork-Big Bat area. The middle portion of the Sinking Creek Cave System, from Rosetta east to Big Spring, drains about 29 sq mi. In the Big Spring-Rosetta area, cave flood crests exceed 50 ft. This researcher realizes that the concept of a watershed in any karst area is tenuous, and that water from one area can cross drain to many different localities, depending upon the seasonal character of the water flow within each karst hydrosystem.

Extended Profile of Sinking Creek

The overland profile of Sinking Creek (Fig. 32), from the headwater junction of Blue Fork and Rocky Fork (A) to the Ohio River at Stephensport (L), is 36.5 stream miles long. The solid line represents the stream bed. The segments between points A and B has a relative steep gradient, typical of a youthful stream. There are small waterfalls and rapids. Between points B and J, the profile begins to flatten to a more mature gradient. The profile changes significantly between points J and G, where an extremely youthful channel is developed. Flow in the section between E and G is ephemeral and occurs only during periods of exceedingly high precipitation. The profile between H and L is almost flat. If subsurface karst piracy had not occurred in Sinking Creek, the profile between E and H would probably be similar to that between H and L.

In general, Sinking Creek has a gradient that in concave upward. However, the section between points G and J is convex, which is atypical of a normal stream profile. In karsted carbonate terrains, such a profile could result from rapid diversion of surface flow to underground routes. Stream load and discharge would be rapidly reduced. The entire cut-off section (E-G) is out of equilibrium, as a result of stream piracy from the surface route to the low-gradient trunk cave passages.

In essence, this evidence all points to rejuvenation of Sinking Creek (the convex profile is, I believe, typical of a rejuvenated stream). Sometime after rejuvenation of the surface stream, however, water was pirated out into existing cave passages and followed a more direct course toward the mouth of Sinking Creek. Piracy appears to be migrating upstream, and, at the same time, the large karst resurgences are migrating downstream. An example is the Clifton Church karst head, which was abandoned in favor of a

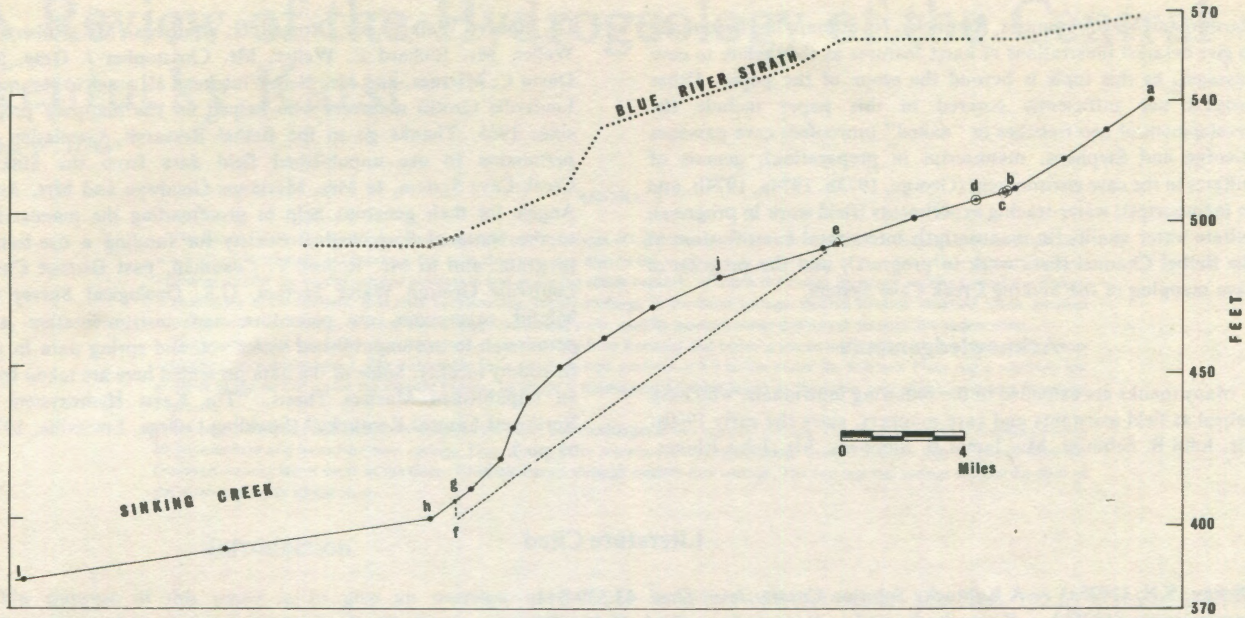

Figure 32. Extended profile of Sinking Creek from the perennial headwater region (A) to the Ohio River (L). The lengths of the subsurface-stream flow routes (E and F) are exaggerated to allow presentation of the surface stream profile. Symbols: (A) Head of Sinking Creek at junction of Blue Fork and Rocky Fork; (B) Flat Rock Spring; (C) Conners Spring No. 1; (D) Fiddle Spring; (E) Major sinks of Sinking Creek; (F) Projected base of rise pit to Boiling Spring; (G) Boiling Spring; (H) Start of normal surface-stream gradient; (L) Stephensport, on the Ohio River.

lower discharge point at Hardins Spring. The Webster karst head very likely was once the major resurgence for the entire area, but the resurgence has since migrated to where Boiling Spring is today. With time, the major resurgence will migrate downstream from Boiling Spring to the ephemeral Clifton Church karst head, forming a rejuvenated karst resurgence. During flood periods, numerous springs resurge from the walls of the Clifton Church karst head. This suggest that a large, mostly abandoned paleo-trunk cave passage is situated behind this ephemeral terminal discharge point. Migration of resurgences and the presence of a convex stream profile are good indicators for subsurface piracy.

In the upper perennial reaches of Sinking Creek, two major stream terraces are found between B and J (Fig. 32). The elevation of the long stream profile between these two points is 490 to 520 ft. The lower stream terrace is between 530 and 540 ft, and the other, much the higher of the two, is situated between 550 and 570 ft elevation. These terraces are reminents of much higher local base levels. The higher terrace is tentatively correlated with the Blue River Strath of south-central Indiana (Powell, 1964). The Blue River Strath in Indiana is developed from 60 to 125 ft above local drainage systems; the upper terrace along Sinking Creek is 60-70 ft above the thalweg.

Anastomosing Feeders to Karst Springs

Where a master cave stream discharges at a valley wall, it appears to form a system of anastomosing, interconnecting, eliptical tubes with satellite spring outlets. This mode of spring discharge is typical of large karst resurgences in north-central Kentucky. It is best illustrated by the Head of Doe Run Spring, an "occluded", tubular bluff spring (George, 1972a) and, during floods, by Boiling Spring. Two karst springs that can be entered by man are McCracken Spring and Webster Cave.

Cave Passage Geometry as Related to Mooretown Lithology

Bisecting the tri-county area is a clastic-paleo channel deposit (Bethel Channel). It consists of upper, distributary, channels of Mooretown sandstone to the northwest and of Mooretown shale to the southeast. The presence or absence of the Mooretown Sandstone Member effectively divides area caves into two genetically different classes.

Caves southeast of the channel consist of high, narrow, canyon passages, some of which display more than 150 ft of vertical relief. Passages may extend all the way from the St. Louis Limestone to the top of the Beaver Bend Limestone. Examples of this kind of cave passage are found in Constantine Saltpeter Cave (George, 1973a), Great Wonderland Cavern (George, 1972c), Sweet Potato Cave, and the Belt-Whitehead System. These canyon passages extend through the Mooretown Shale Member. Large, rectangular trunk passages are rare, although one of the largest in Kentucky is found in Constantine Saltpeter Cave (George, 1973a).

Three cave-passage types occur northwest of the channel. These are: (1) rectangular trunk passages, (2) eliptical tube passages, and (3) high, narrow, canyon passages. Rectangular trunk passages are generally found under upland areas. Some of these passages follow fracture traces that, also, determine the positions of ridges and valleys. Most trunk passages are developed in the Ste. Geneviève and Paoli limestones. Examples of abandoned trunk passages are G-17 in Thornhill Cave and Lead IV in Big Bat Cave. Some contain cave streams that fill the passage about half way to the ceiling (during flood periods); for example, Mountain Rooms in Big Bat Cave, Griffith Avenue in Constantine Saltpeter Cave, and Roaring River Cave. Eliptical sewer tube passages (trunk drains which are very similar to shaft drains) act as connectors between trunk passages located on either side of large uvalas (George and Stephens, manuscript in preparation). The Thornhill-Lockard connection, by way of the Sinkhole Series in the Corners Uvala, is an example. In places where the Mooretown Sandstone is missing, high, narrow canyons are encountered, some of which extend up to the top of the Beaver Bend Limestone. On the other hand, where the Mooretown Sandstone overlies the Paoli Limestone, locally produced recessional canyons and shafts are encountered. Several hundred shafts are known in this area.

Conclusion

The main object of this paper has been to give an overview of the major cave systems and prominent karst features in Breckinridge, Hardin, and Meade counties, Kentucky. No attempt has been made to give detailed illustrations of karst features as they relate to cave passages, as this topic is beyond the scope of the paper. Other projects not sufficiently covered in this paper include the development of neo-trenches or "naked" (unroofed) cave passages (George and Stephens, manuscript in preparation); genesis of sulfates in the cave environment (George, 1973b, 1974a, 1974b, and in manuscript); water-tracing experiments (field work in progress); sulfate water quality (in manuscript); interstratal karstification of the Bethel Channel (field work in progress); and the progress of cave mapping of the Sinking Creek Cave System.

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A Review of the Hydrogeology of the Central Kentucky Karst

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ABSTRACT

The Central Kentucky Karst represents one of the principal types of karst terrain in North America. This mature karst is developed in a sequence of Mississippian limestones 160 m in thickness that dips to the northwest at an average of 6 m/km. Above the major basal limestone aquifer are several perched aquifers in and above the clastic caprock, of which the Haney limestone aquifer is most significant.

Four types of catchments, plus river backflooding, supply recharge to the basal springs: sinking streams, sinkhole areas, perched aquifers, and karst valleys. Vertical shafts, large cylindrical voids, rapidly transmit water downward through the vadose zone.

The aquifer has a large secondary porosity and permeability from fractures and bedding planes and a large tertiary permeability from extensive cavern development. The water table has a relatively high gradient of 9.9 m/km under the Sinkhole Plain and a relatively low gradient of 0.5 m/km under the Chester Cuesta. Its shape is determined by shale layers in the lower part of the cavernous limestone sequence.

The large Sinkhole Plain catchment drains to the north and discharges at two regional springs. The Chester Cuesta is drained by some 80 intermediate and local base-level springs. From Green River records and spring discharge data, it appears that all runoff from the Central Kentucky Karst south of the Green River discharges through conduit-flow springs. The two regional springs account for 80% of the discharge south of the river.

Introduction

The purpose of this paper is to give an overview of the hydrogeology of the Central Kentucky Karst aquifer system, by providing a review of the literature, as well as new data on drainage lines and divides, groundwater discharge, and vertical shafts.

The area of interest (Fig. 33) lies in the Interior Lowlands Province, in the south-central portion of Kentucky, approximately 160 km south of Louisville, Kentucky. The karst is developed in a sequence of Mississippian limestones 160 m in thickness. These limestones have an average dip to the northwest of 6 m/km. They are exposed at the southeastern edge of the Illinois Basin, in a broad syncline superimposed on the general structure. The Central Kentucky Karst is part of a karstic limestone belt that extends from southern Indiana through Kentucky into Tennessee and west to the Cumberland River, along the entire eastern and southern

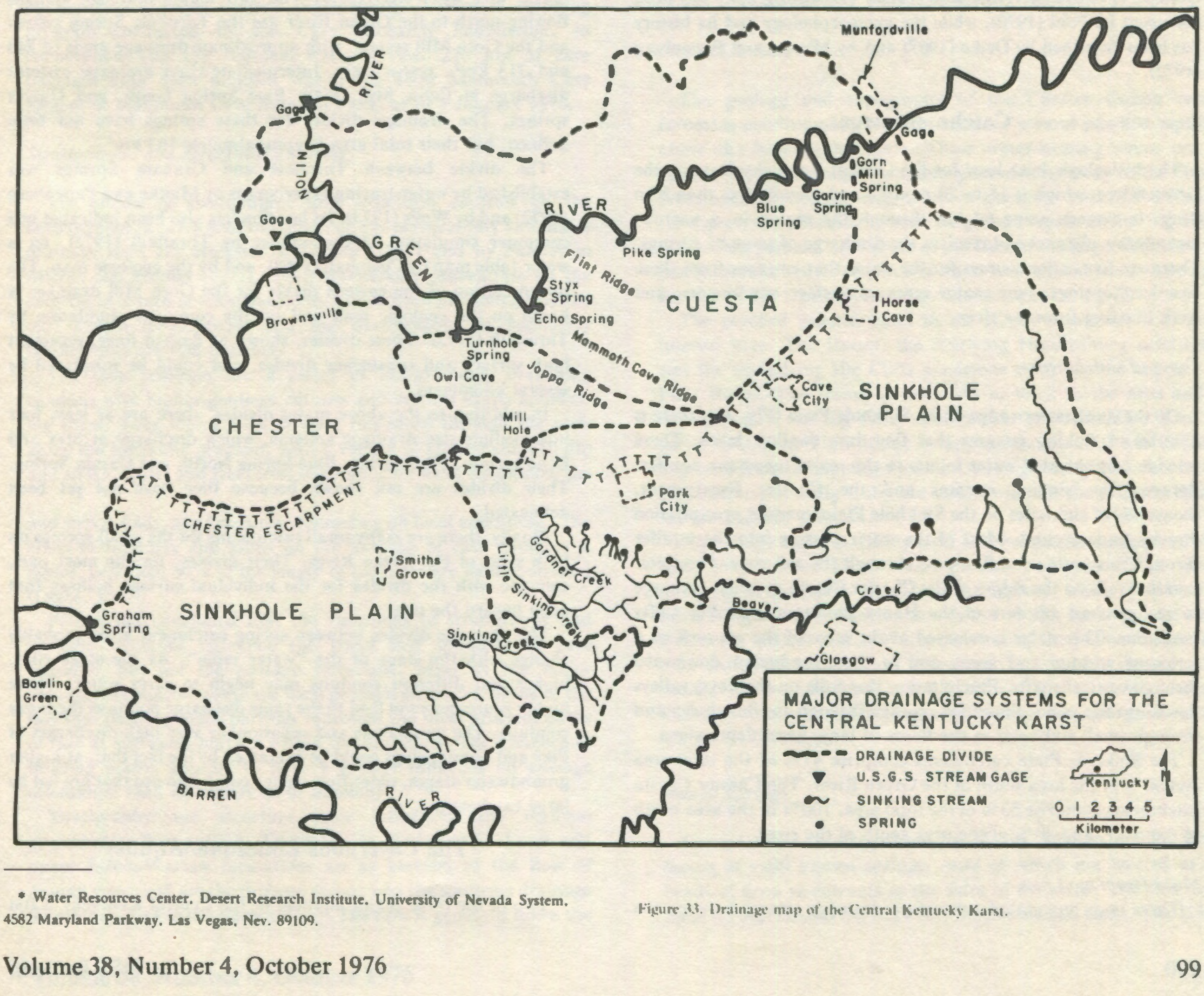

Figure 33. Drainage map of the Central Kentucky Karst.

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perimeters of the Western Kentucky Coal Field. The area is divided into two physiographic units by the Chester Escarpment. The Sinkhole Plain (Pennyroyal Plateau), a limestone plateau of low relief, lies to the southeast. To the northwest are the rugged, dissected, clastic-capped ridges of the Chester Cuesta.

The boundaries of the Central Kentucky Karst are drawn on a hydrologic basis (White *et al*, 1970). The northern boundary is the drainage divide between the Green and Rough Rivers. In the south, the divide is the drainage divide between the Green and Barren Rivers. The rim of the Little Barren River basin near Munfordville is considered the eastern boundary. To the west, the boundary is near Brownsville, where the Green River has not yet dissected the clastic caprock. Recent investigations by Miotke and Papenberg (1972) and by Wells (1973) have indicated that the subterranean drainage to the Barren River should be included. For the purposes of this study, the Central Kentucky Karst will be defined as the 989 km² drainage area of the Green River between the Brownsville and Munfordville U.S.G.S. river gages, minus the drainage of the Nolin River above the Nolin River Reservoir (Fig. 33).

The hydrogrologic setting provides the boundary conditions that determine the flow characteristics of an aquifer. Such geologic parameters as stratigraphy, lithology, and structure are important in determining the type of aquifer, the location of recharge and discharge areas, and the nature of the porosity and permeability of the aquifer.

Various aspects of the hydrogeology of the Central Kentucky Karst have been discussed by many authors. The geology of the area is summarized by Lobeck (1929), Dicken (1935a, 1935b), McFarlan (1943), McGrain and Walker (1954), Deike (1967), Quinlan (1970), and White *et al* (1970). The stratigraphy has been discussed by Pohl (1970), while the geomorphology and its history has been described by Deike (1967) and by Miotke and Papenberg (1972).

Catchment Areas

The hydrologic base level for the Central Kentucky Karst is the Green River, which is 15 to 30 m wide and usually less than 8 m deep. It extends some 64 km through the region in a narrow, meandering channel and receives the discharge of some 85 springs. There are five major sources for the water that emerges from these base-level springs: four major types of surface catchments, and back-flooding from the river.

Types of Catchments

On the southeastern edge of the Sinkhole Plain (Fig. 33), there is a series of sinking streams that flow into swallow holes. These provide concentrated water inputs to the major limestone aquifer. Between the sinking streams and the Chester Escarpment, thousands of sinkholes on the Sinkhole Plain conduct precipitation directly underground. Most of the water passing into the aquifer through this route must first pass through the soil zone. Precipitation that falls on the ridges of the Chester Cuesta provides recharge to the perched aquifers of the Haney limestone and Big Clifty sandstone. This water is released at the edge of the caprock as a series of springs and seeps and is then conducted downward through vertical shafts. Precipitation that falls on the karst valleys flows into the major limestone aquifer through swallow holes and through small sinkholes in the floors of large karst depressions.

The Sinkhole Plain catchments comprise 47% of the total area and 60% of the area south of the Green River. The Chester Cuesta catchments comprise 53% of the total area, 100% of the area north of the river, and 40% of the area south of the river.

Main Drainage Lines

There is an integrated subsurface drainage system between the Sinkhole Plain and the Green River. At the present time, there appear to be two major drainage lines to the north: (1) the Turnhole Spring drainage system, which drains the central part of the Sinkhole Plain through Mill Hole to Cedar Sink (Owl Cave) and then to Turnhole Spring on the river, and (2) the Gorn Mill Spring drainage system, which carries water from the eastern part of the sinkhole plain to the Green River at Gorn Mill Spring (Fig. 33). The evolution of the drainage systems from the Sinkhole Plain is discussed by Deike (1967), Miotke and Papenberg (1972), and Wells (1973). The locations of the major drainage lines are stratigraphically and structurally controlled, in that their spring locations were determined by where and in what sequence the Green River breached the limestone and provided outlets to the river (Deike, 1967).

The drainage systems of the Chester Cuesta are not as well defined. Some of the water enters the major drainage systems described previously, whereas part of it feeds the smaller systems of Echo, Styx, Pike, Blue Spring South, and Garvin Springs. In addition, valley drain systems are developed in all the hollows leading to the river on both sides of the Green River.

Drainage Divides

The dashed lines on Figure 33 show the major drainage divides for the Central Kentucky Karst and for the Graham Spring drainage to the Barren River. The interpretation of these divides is based on dye-tracing results, geologic maps, water-table maps, and computer simulation of the aquifer. The limits of the 363 km² Graham Spring drainage were defined by Miotke and Papenberg (1972) and by Wells (1973). The two major drainage systems flowing north to the Green River are the Turnhole Spring system and the Gorn Mill system, with approximate drainage areas of 246 and 313 km², respectively. Intermediate-sized drainage systems discharge at Echo, Styx, Pike, Blue Spring South, and Garvin springs. The drainage divides for these springs have not been defined, but their total area is approximately 163 km².

The divide between Turnhole and Graham Springs was established by water-tracing experiments by Miotke and Papenberg (1972) and by Wells (1973). Its location has also been indicated in a computer simulation of the aquifer by Thrailkill (1972), on a water-table map by Cushman (1968), and by the geologic map. The interpretation of the eastern divide for the Gorn Mill drainage is based on the geologic map and on the computer simulation by Thrailkill (1972). These divides, shown as dashed lines, represent both surface and subsurface divides, and could be misplaced by several kilometers.

In addition to the above major divides, there are at least four intermediate-size drainage systems, which discharge at Styx and Echo springs, Pike Spring, Blue Spring North, and Garvin Spring. Their divides are not shown because they have not yet been delineated.

Finally, there are many small catchments for the small springs on both sides of the Green River. Their divides, for the most part, coincide with the divides for the individual surface hollows that open toward the river.

The drainage divides between spring catchment areas probably change with the stage of the "water table". As the stage rises, higher and different conduits may begin to carry water. These higher routes may not lead to the same discharge points as the lower conduits. The anomalous and occasionally very high discharges of Pike and Echo springs could be explained by the fact that, at higher groundwater stages, older flow routes come into use that are fed by large catchments.

The Cavernous Limestone Aquifer

Two of the three major aquifers in the Central Kentucky Karst are within carbonate rocks. The major base-level aquifer is developed in the thick St. Louis-Ste. Geneviève-Girkin limestone sequence, which is limited in depth by the impermeable shales and siltstones in the lower St. Louis and Upper Salem-Warsaw formations. Two perched aquifers are developed above the major carbonate aquifer in the Haney limestone and Big Clifty sandstone. In addition, there are several small bodies of perched ground water in the upper part of the Girkin formation.

Porosity and Permeability

Permeability depends on the porosity, size of openings, and fracture pattern. Carbonate rock permeability is of three types: (1) primary porosity and permeability that is due to the presence of the initial communicating pore spaces, (2) permeability that is due to a network of joints, fractures, and bedding planes and (3) permeability due to cavernous openings. A carbonate aquifer can have a very high permeability due to the development of an extensive system of interconnected solution conduits ranging in size from tenths of milimeters to tens of meters in diameter.

The ranges for porosity and permeability in the carbonates of central Kentucky is indicated by Brown and Lambert (1963). The primary porosity of the Ste. Geneviève Limestone is 3.3% and the coefficient of permeability is 0.0016 liters per day per square meter as determined from core samples. Specific capacities of wells drilled in the St. Louis range from 69 to 8700 liters per minute per meter of drawdown.

Owing to the development of an extensive conduit system that transmits large quantities of water, the Central Kentucky Karst aquifer has a large tertiary permeability. Under a mapping program conducted by the Cave Research Foundation, in cooperation with the National Park Service, 270 km of cave passages have been mapped within the Flint Mammoth Cave System in Mammoth Cave National Park, as of 1 January 1975.

Stratigraphic and Structural Controls

The effects of stratigraphic and structural controls on the development of caves and, therefore, on the permeability and major drainage lines are discussed by Deike (1967) and by Palmer in Miotke and Palmer (1972). Their discussions concentrate on the controls of cavern development, but, as has been previously pointed out, cavern development and the development of secondary permeability are tied together.

The cave passages are a series of low-gradient, branching conduits with higher-gradient cut-offs and vertical shafts connecting them. They follow lines of least resistance through the rock along the hydraulic gradient, with a variety of relationships to dip and strike, folds, and fractures, depending on local conditions, the

and strike, folds, and fractures, depending on local conditions. The effect of fractures and joints on the location of cave passages is minor. Less than 50% of the length of passages are joint controlled (Deike, 1967).

Palmer has concluded that the variations in stratigraphy and geologic structure influence the trend and gradient of cave passages, but not the passage elevations. Most of the major passages are concordant to the local geologic structure, owing to the fact that bedding-plane partings represent the most efficient paths of flow at any given horizon within the limestone. The presence of bedding planes appears to have a greater influence on passage orientation than do variations in lithology.

Stratigraphy and structure have influenced the regional groundwater flow pattern. The shales in the lower St. Louis and upper Salem-Warsaw formations act as barriers to the flow of ground water and are the primary reason why the southern drainage divide is so close to the Barren River. The flow is generally down the dip (to the north) toward the Green River. Toward the western end of the Sinkhole Plain, the dip changes to the west. The shales in that area lie below the Barren River, and groundwater flow is westward to Graham Spring (Fig. 33).

The Water Table on the Sinkhole Plain

Cushman (1968) has mapped the water table beneath the Sinkhole Plain from data based on late-fall records obtained intermittently over a period of 5 years at more than 300 drilled wells. In general, the map indicates that the water surface slopes north and northwestward toward the Green River. The Green River is at an elevation of 128-129.5 meters above mean sea level within the Central Kentucky Karst. The basal water table slopes from an altitude of more than 213 meters to 129.5 meters in a distance of 17.7 km. The gradient in the Sinkhole Plain is relatively steep, dropping about 79 meters in 8 km from the drainage divide to the escarpment, or about 9.9 m/km. In contrast, the gradient is remarkably low from the escarpment to the Green River, dropping only about 4.6 meters in 9.7 km for a gradient of 0.5 m/km. Thus, nearly 95% of the fall occurs in the upper half of the basin. This marked contrast in water-table slope is believed by Cushman to reflect the influence of structure, as the permeability of the aquifer is nearly the same at both places. The slope of the basal water table is probably controlled by the shale layers in the upper Salem-Warsaw limestone, which dips below the level of the Green River at the position of the Chester Escarpment. These beds do not influence the shape of the water table north of the escarpment.

Haney and Related Aquifers

General Setting

The geology and topography of the Chester Cuesta create favorable conditions for the formation of several perched aquifers above the basal water table. These water-bearing zones are as follows: (1) a generally continuous groundwater body in the Haney limestone, perched on shale layers in the lower part of the Haney, (2) a generally continuous body of ground water in the Big Clifty sandstone, perched on a shale at the base of the Big Clifty, and (3) discontinuous bodies of perched water above local shale beds near the top of the Girkin formation (Cushman *et al*, 1965).

The perched ground water in the Haney limestone is of most interest here. The Haney, the overlying Hardinsburg sandstone, and the underlying Big Clifty sandstone comprise the caprock of Flint Ridge. The Haney is about 12 m thick in the area and is exposed above the Big Clifty at elevations of about 225-232 m. Its aerial extent on Flint Ridge is approximately 10.9 km^2. These caprocks contain numerous joints. Openings along the joints in the Hardinsburg and Big Clifty are small, whereas those in the Haney are enlarged by solution.

Flint Ridge has a relatively level and permeable surface, so that most precipitation either sinks directly, or runs off in short, intermittent streams to the edge of the caprock, where it sinks. The infiltrating water is directed downward to the basal groundwater system via vertical shafts. Water penetrating the permeable surface rock moves downward until it reaches the relatively impermeable shale at the base of the Haney Limestone. It then moves laterally through solution channels to emerge as seeps and springs along the margin of the ridge.

Haney Springs

Concentrated discharge from the Haney Aquifer in Flint Ridge occurs at eight known springs, most of which are located at the heads of deep re-entrants in the sides of the ridge. The flow from these springs crosses the Big Clifty Sandstone and sinks. The USGS has monitored the flow of seven of the springs: Three Springs, Bransford, Blair, Adwell, Cooper, Collins, and CCC No. 1. The combined average flow from these springs is 5.4 l/sec, with a base flow of 4.5 l/sec (Cushman *et al*, 1965). The average base flow for the seven Haney springs is 0.42 l/sec/km^2, which is 26% of the base flow per unit area for the entire Central Kentucky Karst (1.64 l/sec/km^2). These figures indicate that most of the water is discharged as small seeps and flows or leaks downward into the underlying perched aquifer in the Big Clifty Sandstone.

Groundwater Discharge

A complete description of the springs of the Central Kentucky Karst is given in Hess, Wells, and Brucker (1974). This information is only summarized here. A census of springs discharging (and recharging) a karst aquifer is essential to the interpretation of the hydrogeology and geochemistry of the karst area. To verify the location of springs, temperature differences and electrical conductivity were used, which are superior to visual methods alone. The temperature-difference method uses a thermistor to detect changes in water temperature. The electrical-conductivity method uses a specific-conductance bridge to detect changes in the electrical conductivity of the water. These measurements should be made in summer or early fall, when the temperature contrast between the springs and rivers into which they discharge is at a maximum.

Using the above methods to verify the locations of newly discovered springs along the Green and Barren rivers in the Central Kentucky Karst, the authors were able to identify 85 springs along the Green River between Munfordville and Brownsville, Kentucky, and 24 springs along the Barren River between Polkville and Bowling Green, Kentucky. This disproves the common assumption that mature karst aquifer systems are drained by only a few large springs. The Central Kentucky Karst aquifer is drained not only by a few large springs, but also by a large number of smaller springs. The springs were classified as alluviated or as gravity springs, based on their morphology at pool stage of the river, and as "large" or as "local" springs, based on their discharge and specific conductance.

Vertical Shafts

A description of vertical shafts and their role in the movement of ground water in carbonate terranes is given in Brucker, Hess, and White (1972) and is summarized here.

Description and Characteristic Features

The typical shaft is approximated by a right-circular cylinder usually higher than it is wide. These range in diameter from centimeters to tens of meters and in height from centimeters to hundreds of meters. The key feature of the vertical shaft is the vertical wall, which cuts beds of diverse lithology and is not influenced by bedding-plane dip. The bottoms of vertical shafts may be shallow bedrock pans, or they may be flooded or covered with rubble. Where bedrock floors are exposed, they appear as shallow basins of width little more than that of the shaft itself. Likewise, the top of the shaft is a rounded dome, which may contain deeply incised ceiling channels. The shaft walls are characteristically fluted. Water streams down the small channels formed by the flutes. Shafts seldom form as isolated features, but usually occur in groups, or "complexes".

Role of Shafts in the Movement of Ground Water

Vertical shafts can be considered as pipes transmitting ground water by the most effective route through the vadose zone. They are the downstream termini of surface (or underground) catchments in the upper part of the aquifer, and act as the headwaters for an underground drainage network in the base-level part of the aquifer.

The catchments for individual vertical shafts are usually small, but are quite variable in size and shape. In central Kentucky, many shafts derive their water from the Haney aquifer, and many derive their water from the surface runoff of small catchments a few hectares in extent. Entire surface streams may descend to the water table of the carbonate system through vertical shafts.

Water leaves the vertical shaft through a drain. Drains are small, tubular passages, often too small for exploration, and their detailed pattern is poorly known. Where they can be mapped, the drains are observed to form a complex dendritic pattern of small tubes coalescing into larger passages and eventually emptying into a trunk drain that carries the water into the horizontal conduit system. Most of the drains have steep gradients and are located above the low-water stage of the regional water table, although they may flood seasonally.

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Sinkhole Plain Evolution in the Central Kentucky Karst

Steve G. Wells *

ABSTRACT

The Sinkhole Plain of the Central Kentucky Karst has been developed both by surface and by subsurface drainage. In this study, geomorphic evolution of the Sinkhole Plain is delineated by the following: 1) development of a new method, using best-fit curves of stream profiles, to analyze surface drainage evolution; 2) study of the types and patterns of subsurface drainage systems and their relationships to surface drainage and the structural and stratigraphic setting; and 3) delineation of the boundaries of karst watersheds in the Sinkhole Plain.

Development of the Sinkhole Plain has involved the successive lowering of regional base level and is indicated by two different cave levels. Subaerial portions of sinking streams can be mathematically extrapolated to these cave levels. Both subsurface drainage and the regional slope of the Sinkhole Plain surface are concordant with the direction of maximum groundwater (piezometric) slope and their overall patterns are not influenced by local structural and stratigraphic variations.

Introduction

This paper is concerned with the origin of low-relief karst plains in temperate, humid regions. The Sinkhole Plain of central Kentucky, the area under investigation, is an area of this type (Fig. 34). Previous workers have considered that the dominant denudational process of the Sinkhole Plain is subaerial erosion (Weller, 1927; Lobeck, 1928; Miotke and Palmer, 1972; Miotke and Papenberg, 1972), or subsurface solution and collapse, with low escarpments determined by resistant units (Pohl, 1955; Quinlan and Pohl, 1968; Howard, 1968; Quinlan 1970; White *et al.*, 1970). The purpose of this investigation is to determine which factors influence the evolution of surface and subsurface drainage systems in the Sinkhole Plain.

The portion of the Sinkhole Plain under study is an area of about 170 sq mi within the Interior Lowlands province. It is underlain by about 530 ft of Upper Mississippian limestones, primarily those of the Ste. Geneviève formation. These rocks dip gently northwestward toward the Illinois Basin at about 30 feet per mile. On the southeast, the Sinkhole Plain is bordered by the Glasgow Upland, a low plateau developed on the lower, less soluble members of the Upper Mississippian St. Louis formation. On the northwest, it interfingers with the Chester Cuesta (Mammoth Cave Plateau), a sharply dissected district of limestone ridges capped by the Upper Mississippian Big Clifty sandstone. Local relief in the Sinkhole Plain rarely exceeds 50 ft. Within the adjacent Mammoth Cave Plateau, it reaches 400 ft; the Chester (Dripping Springs) Escarpment rises about 200 ft above the Sinkhole Plain.

Drainage lines in the area head in the Glasgow Upland, on the shaley limestones of the Lower Mississippian Warsaw formation, cross the lower part of the St. Louis formation, then disappear at the edge of Sinkhole Plain. Their waters enter the Barren and Green rivers at Graham, Turnhole Bend, and many smaller springs after crossing beneath the Sinkhole Plain and the Mammoth Cave Plateau.

The study area is located in Warren, Barren, and Edmonson counties of south-central Kentucky (Fig. 34), 100 mi southeast of Louisville, and is represented on the Mammoth Cave, Brownsville, Bowling Green, and Scottsville 15' quadrangles. Western portions of Mammoth Cave National Park are included.

Surface Drainage

The Sinkhole Plain is bordered on the south and east by subaerial streams that flow toward the plain but pass underground before reaching the center of the study area (Fig. 34). These streams drain westward from their headwater region on the Glasgow Uplands. Previous investigators (Weller, 1927; Jillson, 1927) have suggested that the streams represent the headwaters of a pre-existing axial drainage line on the surface of the Sinkhole Plain. Subterranean capture of the downstream drainage isolated the

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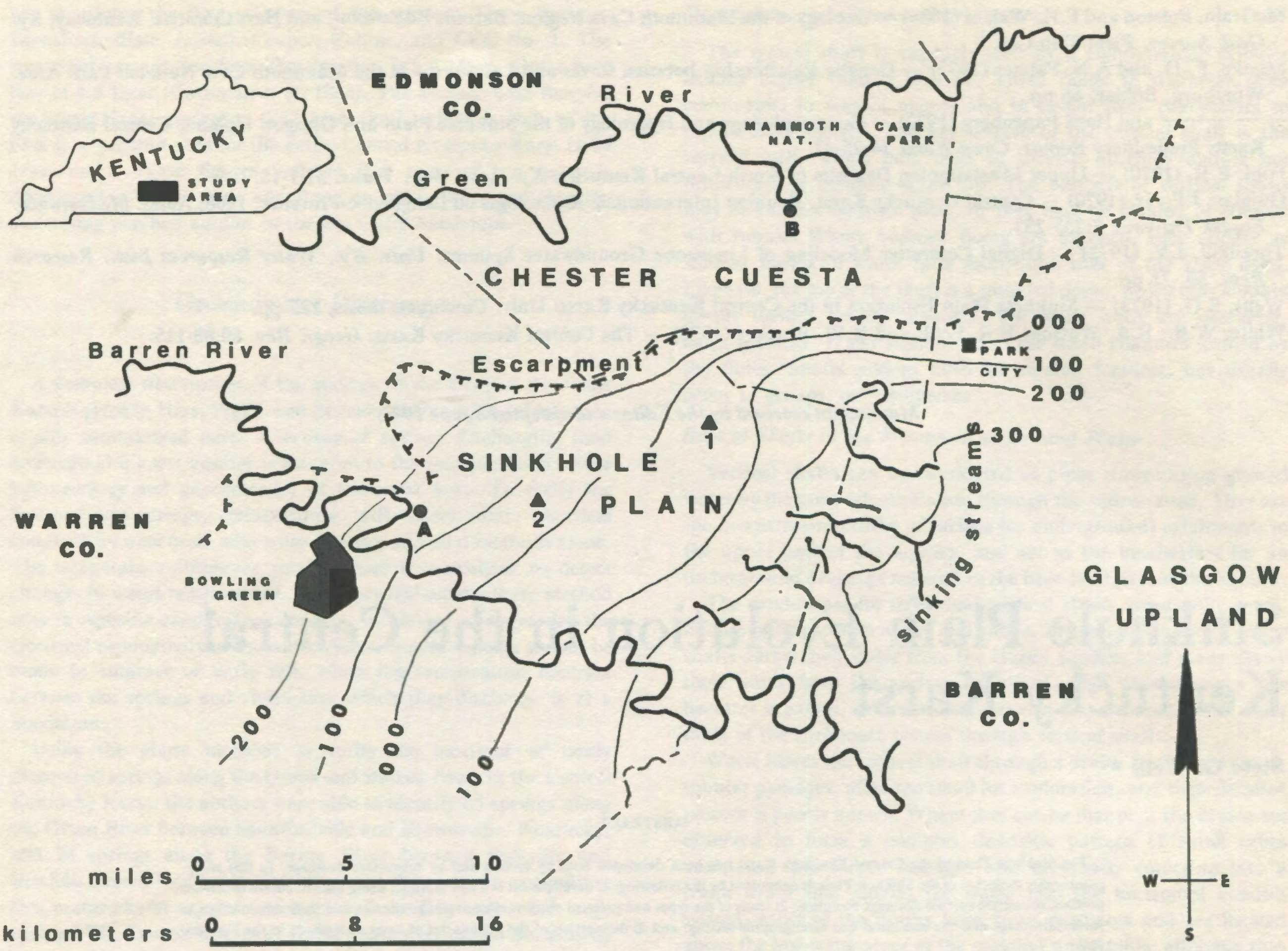

Fig. 34. Location and physiographic map of the study area in the Central Kentucky Karst. Shaded circles, A and B, are Graham Springs and Turnhole Bend Springs, respectively. The shaded triangle, 1, is Smiths Grove Cave. Triangle 2 is Graham Springs Cave. Structural contours in the Sinkhole Plain are drawn on the top of the Chattanooga Shale; the contour interval is 100 ft.

subaerial portions of the sinking streams in the headwater region. In order to reconstruct the role of subaerial drainage in the development of the Sinkhole Plain, it is necessary to project the surrounding surface drainage across the sinkhole plain.

In the present study, a new method is devised to determine whether or not the subaerial portions of these sinking streams were once graded to an ancestral surface river on the Sinkhole Plain. This method of fitting curves to the longitudinal profiles of sinking streams has never been used previously in geomorphic analysis of karst terrain. Best-fit curves of exponential form have been successfully applied to fluvial systems in both arid and humid regions (Lattman, 1973; Shulits, 1941).

The longitudinal profiles of eight sinking streams were determined from USGS 7½ minute topographic quadrangles, and an equation of the form $Y = Ae^{-Bx}$ was fitted to the profiles. Several different equation forms were fitted to the data, the highest regression coefficients in each case are obtained by the equation $Y = Ae^{-Bx}$, where Y = elevation in feet, x = horizontal distance in miles, and A and B = regression coefficients. The minimum regression coefficients for all eight sinking streams are extremely high (Table 5), which indicates that curves can be extrapolated in a valid manner beyond the swallow holes of the sinking streams.

The subaerial portions of the stream profiles were projected x distance in miles from the terminal swallow holes to a point in the center of the Sinkhole Plain. The elevation of the projected profile was determined from the equation (Table 5). The elevation at the head of each stream is represented in the equation as the coefficient A, whereas B varies with drainage basin size.

TABLE 5. Equations of curves fitted to the longitudinal profiles of the subaerial portions of the sinking streams in the Central Kentucky Karst, projected to the nearest surface drainage.

SINKING STREAMS	REGRESSION EQUATION ($Y : Ae^{-Bx}$)	PROJECTED ELEVATION Y(ft.) at x(mi.)		REGRESSION COEFFICENT (r)
		Y	x	
Sinking Creek	$Y : 726e^{-.03x}$	433	10	$-.984 = r$
Pondsville Creek	$Y : 678e^{-.06x}$	444	7	$-.999 = r$
Doty Creek	$Y : 671e^{-.06x}$	437	4	$-.982 = r$
Meiser Creek	$Y : 621e^{-.09x}$	437	3	$-.972 = r$
Three Forks Creek	$Y : 881e^{-.10x}$	451	2	$-.993 = r$
Sinking Branch	$Y : 705e^{-.07x}$	538	2	$-.985 = r$
Little Sinking Cr.	$Y : 730e^{-.03x}$	548	2	$-.985 = r$
Gardner Creek	$Y : 693e^{-.06x}$	445	2.5	$-.992 = r$

These projections indicate that the sinking streams are not subaerially graded to an axial drainage at or above the present surface of the Sinkhole Plain. Five sinking streams were found to be graded to the altitude of the present groundwater level in the Sinkhole Plain (roughly 440 ft), and two sinking streams were found to be graded to a higher-level cave system at an altitude of 540 ft. In contrast, the surface of the Sinkhole Plain is at an altitude of 600 ft.

Subsurface Drainage

The subsurface drainage patterns were delineated by surveying cave systems and by dye-tracing the karst ground-water systems. Two types of cave systems are present. One type consists of large, segmented trunk stream passages at an elevation of 440 to 450 ft (at grade with the Barren River) represented by Graham Springs Cave (Fig. 34). Dye-tracing indicates that these serve as master subsurface drains for the Sinkhole Plain and drain to the Barren River. The second type of cave consists of dry, upper-level, abandoned trunk passages at elevations of 540 to 550 ft, represented by Smith's Grove Cave (Fig. 34). The two caves cited here lie at the two elevations to which the sinking streams are projected.

Scallop data and gradients indicate that Smith's Grove Cave drained to Barren River when the cave served as an active ground-water flow path. The profiles of two sinking streams that now drain to Green River have been extrapolated across a ground-water divide to this cave. This relationship suggests that the two sinking streams were once part of the Barren River drainage, and that, by subterranean piracy, they have been diverted to the Green River.

The regional subsurface drainage of the Sinkhole Plain was delineated by fluorescein dye-tracing. Results indicate two karst basins—Turnhole Bend Springs basin, which drains northward to the Green River, and Graham Springs basin, which drains westward to Barren River. Both karst drainage basins are oriented orthogonal to the two rivers at their discharge points. That is, subsurface drainage is in the direction of the maximum piezometric slope. Maximum subsurface stream gradients were determined from the dye-tracings.

Stratigraphic Control of Subsurface Drainage

No cave passages were observed to be confined to a single stratum, and there is no apparent correlation between orientation of cave passage and direction of bedrock dip. The dip of the strata in the Graham Springs basin is nearly twice the maximum subterranean gradient of the flow paths, which indicates that the magnitude of the bedrock dip does not significantly affect the flow-path gradient. Subsurface drainage in the Turnhole Bend Springs basin is oriented in the down-dip direction, but the maximum subsurface gradients for these flow paths is only one-third as great as the dip of the bedrock. Although the ground-water flow paths coincide with bedrock dip direction, they are not controlled by the amount of dip.

A similarity of passage elevations in cave systems distributed across the Central Kentucky Karst indicates a regional control of the subsurface drainage elevation that is independent of the stratigraphy. Cave-passage elevations in the Sinkhole Plain are similar to those in the Chester Cuesta. It is postulated that two pauses in the entrenchment of the Barren and Green Rivers were accompanied by stabilization of ground-water flow at river level, leading to development of the two cave systems mentioned previously.

The regional slope of the Sinkhole Plain, as determined from the altitudes of sinkhole divides, is concordant with the direction of the piezometric slope but discordant to the bedrock dip.

Conclusions

Based on this study, the geomorphic evolution of the Sinkhole Plain is delineated as follows:

The Sinkhole Plain was developed by surface and subsurface drainage systems. The recent geomorphic history of the karst plain is recorded in its surface and subterranean drainage features. The evolution of the karst plain involved successive lowering of the regional base level, as indicated by cave levels at 540 to 550 ft and at 440 to 450 ft. After the development of the 440 to 450 ft cave level, diversion of drainage from Graham Springs basin to Turnhole Bend Springs basin occurred. Concordance of the regional slope of the Sinkhole Plain surface with the direction of subsurface drainage illustrates that the karst plain was not significantly influenced by its structural and stratigraphic settings.

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Selected Abstracts From Recent Meetings

NSS Convention, Angels Camp, California, 27 June 1975

The Mineralogy of Lilburn Cave, California: A Preliminary Report

Bruce Rogers* and Kathleen Williams*

Lilburn Cave is a large (over 12 km) cave situated in the southern Sierra Nevada, at an elevation of 1600 m. It is formed in Paleozoic/Mesozoic marble, in a roof pendant with extensive contact-metamorphism-induced metaliferous tactite nearby. Due to this setting, the cave has a varied mineralogy:

Aragonite: copper-, zinc-, and nickel-doped, sea-green-to-blue needles and stalactites.
Axinite: clear-brown crystals found on lavender chert.
Azurite: crests, blebs, and up-to-1 mm crystals.
Calcite: stalactites, stalagmites, soda straws, draperies, flowstone, helictites, micro-helictites, cave pearls, gours, cave rafts, moonmilk, and crystal forms.
Chrysocolla: pebbles and cobbles in intermittent streams.
Epidotite: pebbles in intermittent streams.
Goethite: black and brown stalactites, flowstone, and crystals.
Gypsum: crusts, crystals, and angel's hair.
Hematite: aggregates as cave pearl nuclei, staining agent in orange calcite.
Hydromagnesite: moonmilk.
"Limonite": coloring agent in calcite speleothems—some of which are as much as 20 m in height.
Malachite: crusts, rings, blebs, and angel's hair.
Sphalerite: crystals in a goethite/hematite dike (?).
Tremolite: acicular bunches and as "Mountain Leather" felted sheets.
"Wad": unidentified black flowstone-coloring agent; crusts on fills, on marble, and on granite walls.
Witherite (?): white crusts.

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Selected Abstracts from Recent Meetings

NSS Convention, Angels Camp, California, 27 June 1975

A Hydrologic Study of an Alpine Karst, Flathead County, Montana

Jerry F. Ayers*

The Silvertip Mountain karst, located in southeastern Flathead County, Montana, is underlain by an integrated drainage network of cave passages. The geometry of the cave system is, basically, multi-level; the levels are interconnected by vertical shafts. Vadose streams flow through the lower reaches of the system. Dye tests have shown that these streams are directly related to risings located on the northern edge of the area. Though the system is currently forming under vadose conditions, the morphology of individual cave passages, particularly those in the upper levels, suggests a phreatic origin which was strongly influenced by local geological features.

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Some Animal Remains found in the Saddle Butte Cave System

Ellen M. Benedict*

The Saddle Butte Lava Tube System, in southeastern Oregon, is more than 8 mi long. It consists of a number of caves, several of which are from 1000 to 5000 ft long. Skeletal, scat, and mummy remains are evidence of the use by mammals, birds, and reptiles of these caves as refugia in an arid locality. Kit Fox skulls in the Owyhee River Cave establish a significant state record for a threatened species.

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Papers discussing any aspect of speleology are considered for publication in *The NSS Bulletin*. We particularly welcome articles describing important caves and cave areas, on the history of caves and of speleology, on problems and techniques of cave conservation, and critical reviews of current literature, in addition to papers on the more traditional subjects of cave geology, geography, anthropology, fauna, and ecology. The material presented must be original and of lasting interest. Authors should demonstrate the significance of their work to speleological theory and should elucidate the historical antecedents of their interpretations by reference to appropriate literature. Presentations consisting of raw data, only, will not be accepted.

A narrative style of writing is preferred. Fine prose is terse yet free from lacunae, sparkles without dazzling, and achieves splendor without ostentation. Data and interpretations blend effortlessly along a logical continuum so that the reader, having read, neither knows nor cares how many pages he may have turned while following the author's exposition.

As written language must communicate through time as well as across space, neologisms should be introduced only if needed to express new concepts or to record new percepts. Standard usage, therefore, is required of all authors. For general style, refer to papers in this *Bulletin* and to the following handbooks: "Suggestions to Authors" (U. S. Geological Survey), "Style Manual for Biological Journals" (American Institute of Biological Sciences, Washington, D. C.), and "A Manual of Style" (The University of Chicago Press).

Articles on earth sciences (including pseudokarst), life sciences, conservation, social science (including history), and exploration should be sent directly to the appropriate specialist on the Board of Editors (see masthead); articles not clearly falling into any of those categories may be sent to the Managing Editor. Potential contributors, especially those not professional scientists or writers, are invited to consult with the editors for guidance or aid in the presentation of their material.

Two double-spaced, typewritten copies of each manuscript, including all illustrations, are required. Manuscripts should not exceed about 10,000 words in length (approximately 40 pages of typescript), although this limit may be waived when a paper has unusual merit. Photographs must be sharp, high in contrast, and printed on glossy paper. All line drawings should be neatly rendered in "india" ink or its equivalent; the smallest lettering must be at least 2 mm high after reduction. Typed lettering is not satisfactory. Captions will be set in type and added in proof. The dimensions of original drawings and of cropped photographs should be made some multiple of the length and width of a column or of a page, when possible, in order to avoid problems with the layout. In case of doubt regarding length or illustrations, consult with the editors.

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