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Contents

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Stratigraphic and Structural Controls On Landform Development in the Central Kentucky Karst

By Alan D. Howard

ABSTRACT

The number of sinks and hilltops on Meramec and lowest Chester rocks in central Kentucky was found to be closely controlled by several factors, among which the stratigraphic horizon upon which they are developed and the local and regional structure are most important. Several persistent sink-forming stratigraphic units were found in the basal Ste. Genevieve and upper St. Louis formations which crop out as low-relief "sink escarpments." These stratigraphic units are tentatively identified with bedded chert sequences.

The form of the Big Clifty (Dripping Springs) escarpment is closely controlled by the stratigraphic dip; low dip is associated with an irregular escarpment front of linear ridges and numerous outliers, with solutional valleys below the escarpment. In contrast, the escarpment front is nearly parallel to the structural contours in areas of high dip and there exist few reentrants or solutional valleys.

Little evidence was found within the present topography for any past periods of baseleveling within the karst area. However, striking changes of landforms through time are occasioned by the vertical succession of stratigraphy and the local and regional structure. Clear evidence is lacking within the topography for historical changes in landforms resultant from climatic change or fluctuation of local baselevel. However, evidence for such environmental change might be read from alluvial, surficial, and cavern deposits and cavern levels.

INTRODUCTION

Most of the central Kentucky karst area is now covered by 7½ minute Geologic Quadrangle Maps (table 1 and plate 1). These provide an excellent basis for studies into the nature and degree of stratigraphic and structural controls upon landforms in this classic karst area. The investigations reported below are largely based upon an analysis of these maps together with some literature research and field investigation.

The central Kentucky karst area is part of a fairly continuous karst belt developed on Meramec and lowest Chester strata (Mississippian) which extends from southern Indiana through northern Alabama. The limits of the area considered within the present study are shown in plate 1 along with quadrangle locations, major towns, and main streams and rivers.

The stratigraphic sequence is summarized in figure 1. Estimates of the thickness of the stratigraphic units vary greatly with the source of the estimate. Certainly this is partly due to variations in thickness of the units. However, part of the difference in estimates of thickness is probably due to differences in interpretation of the location of the breaks between formations within the stratigraphic column. Additional variation may have been caused by the poor exposure of the units within the karst area, which necessitates indirect estimates of stratigraphic thickness based upon structural contours. This last source of variation is perhaps the most important of the three for the St. Louis limestone.

The limited exposure of bedrock within the karst area has also precluded precise definition and lateral tracing of sub-units within the formations, especially within the St. Louis and Ste. Genevieve formations. Part of the purpose of this study was to determine whether it is possible to recognize stratigraphic sub-units on the basis of their topographic expression.

For the purpose of this study I have defined six landform "units" within the karst area, which are mapped in plate 1. Landform unit 1 includes those land areas underlain by Big Clifty sandstone or overlying formations. Unit 2 consists of landforms on the Girken formation and upper part of the Ste. Genevieve limestone. This includes steep slopes directly fronting the Big Clifty escarpment, solutional valleys and sink complexes in these formations, and knobs in these strata lying in front of the Big Clifty escarpment and rising above the level of the surrounding sink plains. Units 1 and 2 are not distinguished north of the Green River, but are lumped together under unit 6 because of the predominance of surface drainage. Unit 3 consists of those areas irrespective of underlying bedrock which have low relief coupled with sparse, shallow, and broad sinks and occasional surface drainage ("low" sink plain). Unit 4 similarly includes those areas of low relief having numerous, deep, and steep sinks and no surface drainage ("high" sink plain). Unit 5 includes all areas having surface drainage where the drainage eventually sinks underground. Unit 6 is composed of areas having almost complete surface drainage.

Thus landform units 3, 4, 5, and 6 are defined on the basis of general topographic expression without reference to the stratigraphic horizons upon which they are developed. The escarpment in the northern part of the karst area, which is commonly called

Table 1

Geologic Quadrangle Maps Covering the Central Kentucky Karst Area

Quadrange Name	Map No.	Author	Publication Date
Allen Springs	GQ-285	Moore, Sampel L.	1963
Bowling Green North	GQ-234	Shawe, Fred R.	1963
Bowling Green South	GQ-235	Shawe, Fred R.	1963
Bristow	GQ-216	Gildersleeve, Benjamin	1963
Brownsville	GQ-411	Gildersleeve, Benjamin	1965
Drake	GQ-277	Moore, Samuel L.	1963
Glasgow North	GQ-339	Haynes, Donald D.	1964
Glasgow South	GQ-416	Moore, Samuel L., and Miller, Robert C.	1965
Hiseville	GQ-401	Haynes, Donald D.	1965
Horse Cave	GQ-401	Haynes, Donald D.	1966
Lucas	GQ-251	Haynes, Donald D.	1963
Mammoth Cave	GQ-351	Haynes, Donald D.	1964
Meador	GQ-288	Nelson, Willis H.	1963
Park City	GQ-183	Haynes, Donald D.	1962
Polkville	GQ-194	Gildersleeve, Benjamin	1962
Rhoda	GQ-219	Klemic, Harry	1963
Smiths Grove	CQ-357	Richards, Paul W.	1964
Temple Hill	GQ-402	Moore, Samuel L., and Miller, Robert C.	1965

System	Series	Formation or Group	Thickness (in feet)	Lithology	Characteristic Landforms
Mississippian	Chester	Hardinsburg Sandstone	10-60	Sandstone, fine- to medium grained; with minor shale. Sandstones mostly friable to locally indurated.	Underlies broad uplands above Big Clifty escarpment.
Mississippian	Chester	Golconda Formation: Haney Limestone	10-55	Limestone, fine- to coarsely-crystalline, locally argillaceous.	Underlies uplands above Big Clifty escarpment. Poorly exposed, often solutionally removed in zone of outcrop. Minor sink horizon.
Mississippian	Chester	Golconda Formation: Big Clifty Sandstone	40-120	Sandstone, fine-grained, thin-bedded to massive, well indurated. Variable amounts of shale, especially near top and base.	Most of formation exposed at escarpment face. Upper beds underlie marginal parts of upland escarpment. Blocks of sandstone solutionally lowered at escarpment front.
Mississippian	Chester	Girken	60-200	Limestone, dense to crystalline, locally argillaceous or oolitic, massive- to thin-bedded. Contains sparse chert. Contains a few feet of sandstone and shale locally at top.	Exposed on steep slopes fronting the Big Clifty escarpment and in knobs where Big Clifty cap unit is absent. Solutional valleys and deep sinks developed in these beds where stratigraphic dip is low.
Mississippian	Meramec	Ste. Genevieve Limestone	135-215	Limestone, predominantly oolitic, medium- to massive-bedded. Some argillaceous interbeds present. Contains beds and stringers of chert, especially near base. Persistent bedded chert unit recognized at base of formation.	Underlies low-relief plains with sparse, shallow, broad sinks and minor surface drainage.
Mississippian	Meramec	St. Louis Limestone	230-350	Limestone, fine- to coarsely-crystalline, thin- to medium-bedded. Contains some argillaceous and silty beds, especially near base. Minor dolomitic beds present locally. Chert nodules and beds are abundant throughout formation. Lower contact is transitional.	Underlies low-relief plains with numerous, deep, steep sinks and no surface drainage. Underlies low-relief, dip-slope plain with surface drainage which disappears beneath sink plains.
Mississippian	Meramec	Warsaw-Salem	40-170	Limestone, thick-bedded, variable chert nodules. Also siltstone and dolomitic limestone, thin- to thick-bedded.	Supports surface drainage with moderate relief.
Mississippian	Osage	Fort Payne	230+	Dolomitic limestone, thick bedded, very cherty. Also clastic limestone, thick-bedded, and dolomitic siltstone.	
Devonian		Chattanooga Shale		Not Exposed.	

Figure 1.

Columnar section of rocks exposed in the Central Kentucky Karst. Data on rock types summarized from the geologic quadrangle maps listed in table 1. Characteristic landforms summarized from discussion in text. Thickness figures are maximum and minimum thicknesses reported in the geologic quadrangle map texts.

the Dripping Springs escarpment, is universally associated with a caprock of Big Clifty sandstone. Because of the obvious stratigraphic control of the escarpment, the break between physiographic units 1 and 2 is defined as the contact between the Big Clifty sandstone and the underlying Girken formation. As will be discussed later, however, *all* of the physiographic units have consistent association with stratigraphic units.

Structural contours taken from those shown on the new 7½ minute Geologic Quadrangle Maps are given in plate 1, with a contour interval of 100 feet. The datum for contours south of the Big Clifty escarpment (solid lines) is the top of the Chattanooga shale, while the base of the Big Clifty is the datum for areas to the north of the escarpment front (dashed lines).

The contours show the structure of the karst region in the area of the present study to be gentle flexures superimposed upon a farily uniform dip to the northwest. The flexures are primarily either local monoclinal steepenings of the dip or small, shallow anticlines and synclines.

The first part of the paper will be concerned with the nature and extent of stratigraphic and structural controls in the Kentucky karst landforms. The second part of the paper will present speculation on the historical evolution of the karst landforms insofar as evidence for landform changes through time may be found in the topography, geology, and structure. Particular attention will be paid to evidence relating to the contrasting hypotheses of baseleveling (peneplanation or pediplanation) and "dynamic equilibrium".

Structural and Stratigraphic Controls Upon the Karst Landforms

The existence of controls by underlying bedrock upon the landforms developed thereon is fairly obvious in the case of the Drippings Springs escarpment. The Big Clifty sandstone acts as a cap rock unit over the Girken and Ste. Genevieve formations. Solutional valleys, domepits, and marginal sinks are associated with the escarpment front. These features have been discussed at length by Pohl (1955) and Quinlan and Pohl (1966), and will not be emphasized in this paper.

On the other hand, the nature or existence of geologic controls upon landforms developed on the lower Ste. Genevieve, St. Louis, and Warsaw-Salem formations is not obvious. Regional patterns of landforms in the karst area correlate broadly with the outcrop patterns of underlying formations, so some degree of stratigraphic control seems reasonable. The outcrop pattern of the lower part of the Ste. Genevieve generally may be correlated with zones of "low" sink plain and to a lesser extent with "high" sink plain, as these terms are used in plate 1. The upper part of the St. Louis formation generally underlies broad "high" sink plains. The lower strata of the St. Louis generally support surface drainage which may disappear into the sink plains developed on the upper part of the St. Louis. The lowest part of the St. Louis limestone and the Warsaw-Salem and underlying formations are generally characterized by surface drainage with minor, local sink zones.

However, these general correlations do not define the precise nature of the stratigraphic controls. Relief on the sink plains is so low that any stratigraphic controls are only subtly expressed in the topography. More powerful criteria for stratigraphic and structural controls were sought to demonstrate conclusively their nature and existence.

Preliminary investigations were directed toward the delineation of stratigraphic controls on sinks. If such controls exist, they should be manifested primarily in the density and distribution of sinks with respect to the stratigraphic horizon upon which they have developed. Stratigraphic controls presumably affect other properties of sinks, such as depth, area, and elongation. These morphological factors have been studied and indirectly related to stratigraphic controls by LaValle (1965). The present study will be primarily concerned with stratigraphic controls upon density and distribution of sinks.

The number of sink bottoms and sink rims (or lips) developed upon each ten-foot interval of exposed stratigraphic column was selected for study as an informative, objective, and easily-measured quality of sinks. A parallel study measured frequency of hill summits along the stratigraphic column.

Sampling was conducted along strips one-quarter-quadrangle in width running north-south across the width of the karst belt. The sample strips were terminated to the south by the last exposures of the St. Louis limestone and to the north by: 1, the Green River; 2, unbroken expanse of Chester upland (unit 1 in plate 1); or 3, inavailability of geologic quadrangles. In all but one case the entire belt of sink landforms was sampled. Location of the sample strips is shown in plate 1.

Along each sample strip the altitude of the lowest enclosed contour of each sink was noted, together with the structural elevation at that point as read from the structural contours. Both altitude and structural elevation were read to the nearest 10 feet (except on a few maps where the topographic contour interval was 20 feet). In addition, for each sink noted above the number of enclosed contours composing the sink were noted. Where the sinks were compound (sinks within sinks), only the altitude and depth of individual component sinks were used; that is, the compound sink was not treated as an entity. Along the same sample strips the altitude and structural elevation of each hilltop was noted.

The stratigraphic elevation of each hilltop and sink bottom relative to the reference bed used in the structural contours was computed by subtracting the structural elevation from the altitude for each hilltop or sink. Additionally, the stratigraphic elevation of each sink rim was computed by adding the number of enclosed contours to the stratigraphic elevation of the sink bottom.

Plots were made of the frequency of hilltops and sinks versus altitude and stratigraphic elevation. In figure 2 the stratigraphic elevation of each sink bottom was plotted for each sample transect to give the resultant frequency distribution. Figure 3 shows the frequency of hilltops versus stratigraphic elevation. In figure 4 the difference between the number of sink rims and the number of

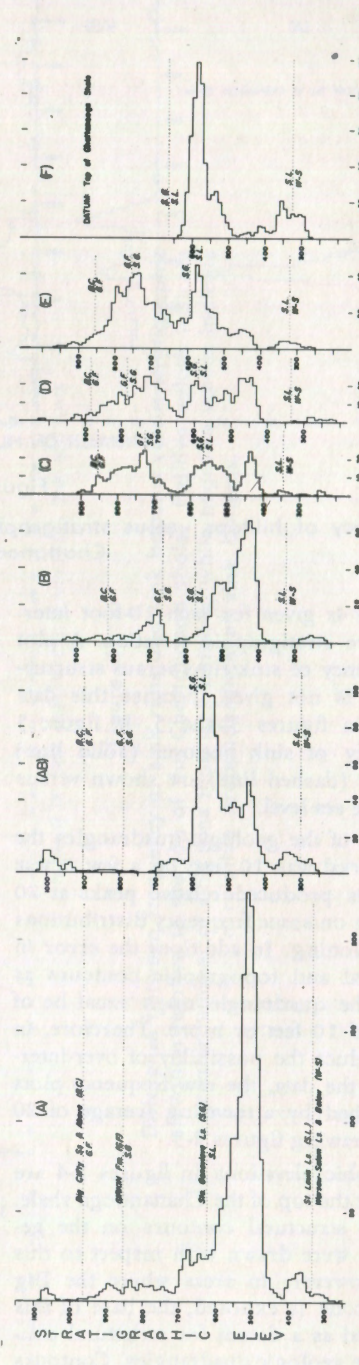

Figure 2. Frequency of sink bottoms versus stratigraphic elevation relative to the top of the Chattanooga shale. Letters in brackets refer to sample transects keyed in Plate 1.

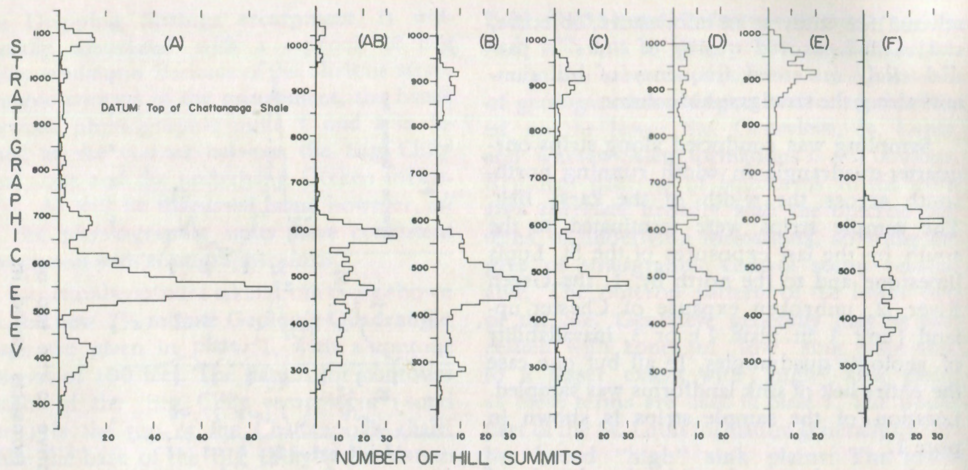

Figure 3.

Frequency of hilltops versus stratigraphic elevation relative to the top of the Chattanooga shale.

sink bottoms is given for each 10-foot interval of relative stratigraphic elevation. A plot of the frequency of sink rims versus structural elevation is not given, because this data is inherent in figures 3 and 5. In figure 5 the frequency of sink bottoms (solid line) and hilltops (dashed line) are shown versus altitude above sea level.

On most of the geologic quadrangles the contour interval was 10 feet; on a few it was 20 feet. This produced relative peaks at 20 foot intervals on some frequency distributions upon raw plotting. In addition, the error in the structural and topographic contours as drawn on the quadrangle maps must be of the order of 10 feet or more. Therefore, in order to reduce the possibility of over-interpretation of the data, the raw frequency plots were smoothed by a traveling average of 20 feet before drawing figures 2-5.

Stratigraphic elevations in figures 2-4 are in feet above the top of the Chattanooga shale. Most of the structural contours on the geologic maps were drawn with respect to this horizon. However, in areas where the Big Clifty sandstone is exposed, the base of this unit was used as a datum for structural contours on the geologic quadrangles. Contours on the Big Clifty sandstone were converted to contours relative to the Chattanooga shale by matching the structural elevations of the two systems of contours at their junction on the map. Where the two systems had a discordant juncture (theoretically indicating variations in thickness of the strata between the two reference horizons), the average value of the junction equivalence over the width of the sample strip was used.

Frequencies of sink rims, sink bottoms, and hilltops derived from these transects are directly comparable only if the structural dip remains constant over all the sample strips. This was, in general, not true for any of the sample strips, so the graphs of figures 2-4 should be used only to make rough comparisons of *relative frequencies*. *Positions* of relative peaks and valleys in the distributions within the stratigraphic columns are regarded as being much more significant than absolute values of the peaks and valleys.

The use of structural contours to define stratigraphic position within the geologic column rests upon several assumptions:

1. That the overburden (regolith) over the bedrock is of constant thickness over the areas sampled. It was assumed that variations in regolith thickness would either be random or would be systematically associated with

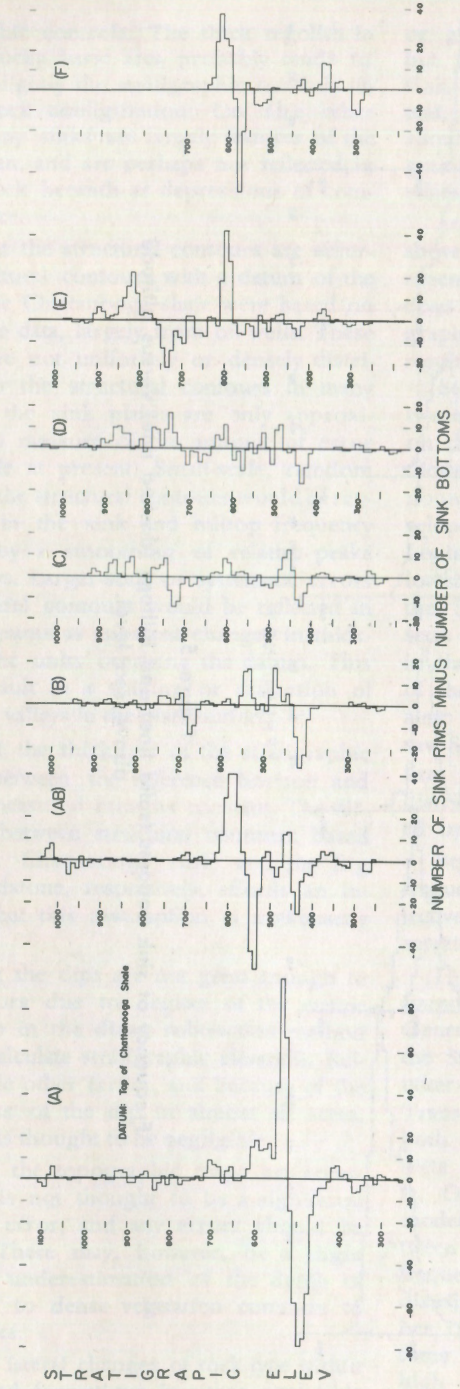

Figure 4.

Difference between frequency of sink rims and sink bottoms versus stratigraphic elevation relative to the top of the Chattanooga shale. Positive values indicate a greater number of sink rims than sink bottoms, while negative values indicate a relative preponderance of sink bottoms.

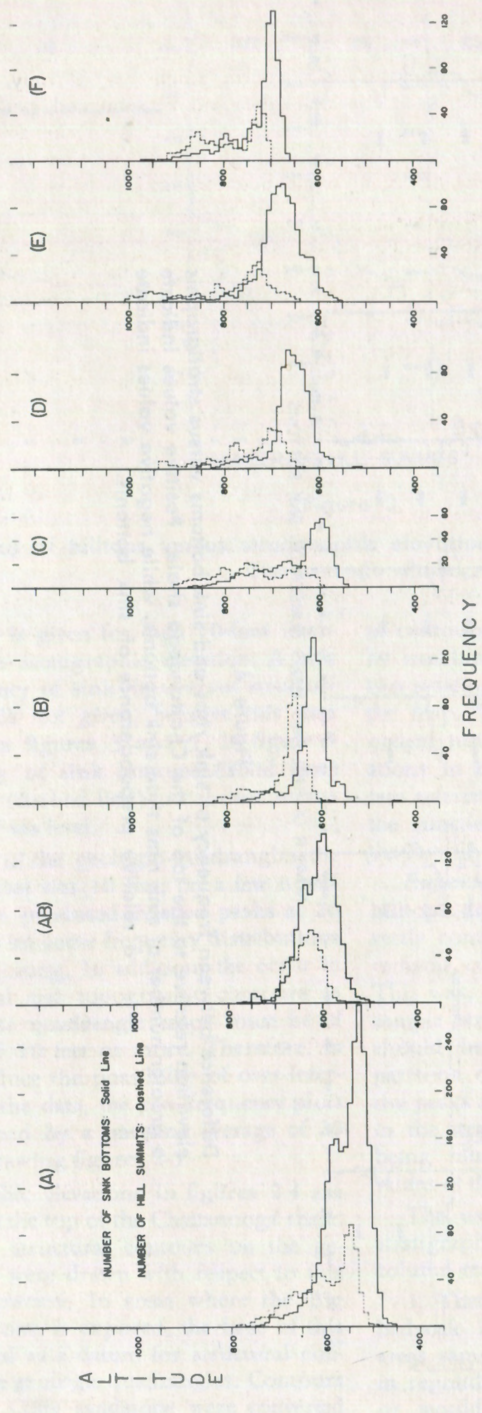

Figure 5.
Frequency of sink bottoms (solid line) and hilltops (dashed line) versus altitude above sea level.

stratigraphic controls. The thick regolith in the Kentucky karst area probably tends to obscure slightly the stratigraphic controls on the bedrock configuration. On the other hand, many sinks are largely features of the overburden, and are perhaps not reflected in the bedrock beneath as depressions of comparable size.

2. That the structural contours are accurate. Structural contours with a datum of the top of the Chattanooga shale were based on subsurface data, largely from oil wells. These holes were not uniformly or densely distributed, so the structural contours in many areas of the sink plains are only approximate. No measure of the amount of error is possible at present. Small-scale, ramdom errors in the structural contours would be represented in the sink and hilltop frequency analyses by a smoothing of relative peaks and valleys. Larger-scale or systematic errors in structural contours would be reflected in the histograms as apparent changes in thickness of the units overlying the datum. This would result in a shifting or distortion of peaks and valleys in the distribution.

3. That the thickness of the stratigraphic column between the reference horizon and the unit measured remains constant. The discordance between structural contours based upon the Chattanooga shale and the Big Clifty sandstone, respectively, affords an indication that this assumption is not exactly met.

4. That the dips are not great enough to cause errors due to neglect of the cosine of the dip in the direct subtraction method used to calculate stratigraphic elevation. Relative to the other errors, and because of the shallowness of the dip in almost all areas, this error is thought to be negligible.

5. That the topographic maps are accurate. This is not thought to be a significant source of error, and any errors should be random. There may, however, be a slight systematic underestimation of the depth of sinks due to dense vegetation common to deeper sinks.

6. That lateral changes of rock type within the sampled formations is either negligible or gradual. If stratigraphic controls exist, but the stratigraphic units which influence sink development are of only very local extent, then no pronounced peaks and valleys should be expected in the frequency histograms, or if present, they should not correlate between sample strips.

An indication of the reasonableness of the above assumptions lies in the degree of consistency of the resultant frequency distributions in figures 2-4. The following paragraphs will concern the consistency of the results.

Stratigraphic positions of the contacts between the geologic formations, as mapped on the Geologic Quadrangle Maps, are indicated in figure 2. This figure shows pronounced zones of high sink frequencies developed on the upper two-thirds of the St. Louis formation, and sink-poor zones in the lower part of the St. Louis and middle of the Ste. Genevieve formations. Some transects encountered and a high density of sinks in the Girken formation and the upper part of the Ste. Genevieve limestone, and minor sink zones in the lowest St. Louis and Warsaw-Salem formations. These observations confirm the general relationships between sink distribution and underlying bedrock suggested by the landform classification of plate 1.

Beyond this point the interpretation of the frequency distributions becomes more subjective, but I will advance three general observations about the distributions:

1. Two zones of high sink frequency are found within the St. Louis and lowest Ste. Genevieve limestones. One of these is near the St. Louis-Ste. Genevieve contact, and the other is near the middle of the St. Louis. Transects (AB), (C), (D), and (E) show both of these zones, while the other transects show only one of the two zones clearly. One or more stratigraphic zones with moderate sink frequency may be present between these two. The two zones of high sink frequency are also reflected in the frequency distribution of the difference between the number of sink rims and number of sink bottoms (figure 4). In this figure the zones of high sink frequency are at about the same structural elevation as transitions downward

from a stratigraphic zone supporting a relative surplus of sink rims to an underlying zone of relative deficiency of sink rims as compared to sink bottoms.

2. Stratigraphic zones of relatively high frequency of hill summits usually overlie by 20 to 40 feet zones of high frequency of sinks within the Warsaw-Salem, St. Louis, and lower part of the Ste. Genevieve formations. Sink zones in the Girken and upper part of the Ste. Genevieve formations, on the other hand, are associated with almost no hill summits, but summits are common about 150 feet above the sink bottoms on the Big Clifty and overlying formations.

3. The total number of sinks or hilltops developed on a given stratigraphic horizon is greater the more gentle is the stratigraphic dip. Profiles (A), (AB), (B), and (F) are in areas of predominantly low stratigraphic dip, while the other profiles are in areas of steeper dip. This is expressed also in the general width of the sink plains in the areas of sampling (see plate 1). The absence of a discernable peak of sink frequency in the middle part of the St. Louis formation in transect (F) is probably due to the local steepening of dip which is present at the zone of outcrop of that part of the formation. The relative lowness of the sink peak associated with the St. Louis-Ste. Genevieve contact in transect (A) may also be due to a steepening of the dip.

In figure 2 a lateral correlation of the histograms was attempted. The frequency histograms of each transect were shifted vertically so that the sink frequency maximum in the middle part of the St. Louis formation, where present, was positioned along a line of constant value of the vertical ordinate. The peak in the upper part of the St. Louis formation in transect (F) was equated with the upper peak in transect (E). These correlations were carried through figures 3 and 4.

It becomes immediately apparent in comparing relative stratigraphic elevations and formation contacts laterally between the transects that the assumptions made during the sampling of the transects are only partially met. The difference in relative stratigraphic elevation of the lower sink zone in the St. Louis formation should presumably represent changes in thickness of the strata between the sink zone and the Chattanooga shale (errors in assumption 2, above) or less likely, systematic errors over wide areas in structural contours (assumption 3). Similar errors are indicated by the variation in the position of the Big Clifty-Girken contact with respect to both stratigraphic elevation and the lateral correlation. Variations in the position of the St. Louis-Ste. Genevieve and St. Louis-Warsaw-Salem contacts are regarded as being less significant because of the poor exposure of these units, and in the latter case because of a transitional zone between the formations. The position of the minor, sporadic sink zone(s) near the contact between the St. Louis and Warsaw-Salem formations varies both with respect to stratigraphic elevation and the lateral correlation. However, in this case the stratigraphic zones supporting sink development may be truly local, and not correlative over long distances.

Because of both the partial failure of the assumptions used in constructing the frequency traverses and the limited information inherent in characterization of the karst area landforms solely in terms of sink frequency, sink depth, and hill summit altitude, additional techniques were sought to delineate more precisely the nature of the stratigraphic controls. The traverses had indicated two strong sink zones developed on the upper and middle parts of the St. Louis formation, respectively; the outcrop belts of these zones were examined in order to determine whether a characteristic suite of landforms is associated with these zones of high sink frequency.

In areas of low dip the upper and lower sink maxima on the St. Louis appeared to be associated with low, irregular escarpments. Relatively deep sinks are common immediately behind the escarpments. The escarpment relative relief averages perhaps 30 feet. If these escarpments indeed have objective reality, they should be mappable.

Such an attempt at mapping was made to the southeast of the town of Smiths Grove, through which transect (AB) passes. This area was chosen because the high, narrow peaks of transect (AB) gave promise that the structural contours in this area give a relatively accurate measure of stratigraphic position within the St. Louis formation. The method of mapping was to define the escarpment by two criteria:

1. The lips of deep, steep sinks. This was defined as the uppermost enclosed contour which *closely* rings any deep, steep sink.

2. The break in slope at the escarpment front separating flattish upland from steep slopes fronting the escarpment.

It was assumed that the escarpments are developed on a "resistant" layer of some sort, and that the outcrop of such a layer would be at approximately that topographic position defined by the two criteria given above.

The criteria were applied in a statistical sense, in that the average stratigraphic elevation of sink rims and segments of escarpment front over a small area was used to define the stratigraphic elevation of the escarpment within that area. The stratigraphic elevation was determined by subtraction of structural elevation from topographic elevation as was done in the transects. This procedure was employed over the outcrop zones of both the upper and lower sink zones to find the estimated stratigraphic elevation of the two "resistant layers" over the width of the sink plain to be mapped. Note that in this method continuity of the escarpment rather than a constant association of rock layers with a particular stratigraphic elevation is assumed.

The estimated stratigraphic elevations of both resistant layers were found to vary through approximately 30 feet over the width of the mapped sink belt. The estimated stratigraphic elevations of the two hypothetical resistant layers were used together with the structural contours to draw the map given as plate 2, together with a cross-section through the mapped area approximately parallel to the dip. The assumed outcrop pattern of the upper resistant layer is drawn with solid lines, and that of the lower layer with dashed lines. The presence of an additional escarpment intermediate between the upper and lower escarpments discussed above became apparent during the mapping, and was subsequently added as dotted lines.

The resultant map and cross-section form an extremely subjective interpretation of the geomorphology of the sink plain. One positive indication of the "reality" of the interpretation is the close coincidence between the average stratigraphic elevation of the "resistant layers" and the maxima in the sink frequency distribution of sink bottoms in transect (AB) (figure 2), especially for the upper and lower sink zones. The average stratigraphic elevation of the hypothetical resistant layers is, however, slightly above the maxima of sinks in the transect, as would be expected between data based on sink rims on the map and sink bottoms in the transect. Similar positive correlations hold between the mapped horizons and hilltop maxima (figure 3, transect (AB) and inflections in the graph of the difference between frequency of sink rims and sink bottoms (figure 4, transect (AB)).

A series of escarpments associated with approximately the same portions of the St. Louis and Ste. Genevieve limestones were identified and traced to the southeast of Bowling Green, but were not mapped in detail. Their existence, however, forms another partial confirmation of the map units.

Despite the self-consistency of the interpretation of the geomorphology of the sink plains offered in plate 2, the interpretation must be regarded as hypothetical until detailed mapping either confirms or rejects the presence of natural rock units which can be identified with the "resistant layers." A strong possibility exists that the resistant layers may be identified with cherty horizons in the limestone sequence. Malott (1922) recognized the strong influence of cherty layers in the development of the sink plains of southern Indiana. Several of the Geologic Quadrangle Maps (GQ 227, 234, 235, 401, and 558) identify massive or bedded chert units within the limestone near the contact of the St. Louis and Ste. Genevieve formations. Quinlan and Pohl (1966) have identified a persistent bedded chert unit at the base of the Ste. Genevieve formation averaging several feet in thickness, which they feel acts as an impermeable and resistant layer supporting sink topography. This unit may, therefore, approximately correspond in outcrop pattern with the uppermost resistant unit (solid line) in plate 2. Unfortunately, to my knowledge, no persistent rock units have been identified within the St. Louis formation to correspond with the lower two sink maxima, although all quadrangle maps indicate an abundance of beds and stringers of chert throughout the upper and middle part of the St. Louis formation.

The relationship between sinks and escarpments mapped in plate 2 suggests that there may exist at least two basic types of sinks on the St. Louis sink plain. Near the margin of the escarpments occur a large number of deep, steep sinks which breach the supposed resistant layer. Less-steep, shallower sinks which do not penetrate the resistant layer are common above and behind the escarpments.

The portion of plate 2 which is south of the sink plains and north of the Barren River drainage network is dominated by surface drainage which closely follows the dip of the stata and which drains northward into the sink plain, disappearing underground at its margin. This zone of dip-slope drainage on lower St. Louis beds is a conspicuous feature of the central Kentucky karst, as is demonstrated by plate 1. Presumably the dip-slope drainage is developed on top of one or more resistant and/or impermeable beds in the lower portion of the St. Louis formation which therefore cap a southward-facing cuesta. Although the low escarpments of the sink plains appear to be determined by thin resistant beds whose outcrops are adequately represented by thin lines in plate 2, the cuesta escarpment is developed on 30-50 feet of lower St. Louis strata which apparently contain more than one resistant horizon. An arbitrary marker horizon which is associated with a minor sink zone in the lower-right-hand corner of plate 2 has been selected to demonstrate the close approximation of the slope of the dip-slope cuesta to the regional dip (dash-dot lines in plate 2).

The stratigraphic units within the lower Ste. Genevieve, St. Louis, and upper Warsaw-Salem formations postulated in the preceding discussion are summarized in figure 6, along with their most common topographic associations. The lateral persistence of these units (assuming their objective existence) is unknown.

Quinlan and Pohl (1966) also distinguish a chert horizon high in the Ste. Genevieve limestone which they recognize as a sink- and bench-forming horizon. This layer probably correlates with the zone of high sink frequency at the top of the Ste. Genevieve formation and the base of the Girken formation (figure 2).

Regional and local rock structure influences landforms both directly and indirectly. Structural features controlling landforms may include faults, joints, local and regional strike and dip, and folds.

Joints are of undoubted importance in controlling the development of landforms. This is especially true in karst landforms, for much of the subsurface drainage is along solutionally-enlarged fractures. Linear sinks and lines of sinks on the sink plains are most likely developed along strong joints. However, this study is primarily concerned with the larger-scale structural controls on landforms which are subject to analysis on the basis of topographic and geologic maps.

The effects of stratigraphic dip upon landforms is most strikingly displayed by the Big Clifty escarpment. Where the local dip is relatively steep, as it is immediately north of Smiths Grove (plate 1), the Big Clifty escarpment closely parallels the structural contours. In such a case the front of the escarpment is nearly linear, and the few re-entrant valleys in the escarpment are shallow. Solutional valleys behind the escarpment are conspicuously absent. Also, the pro-escarpment ramparts in the Girken and Ste. Genevieve formations (unit 2 in plate 1) are narrow, and correspondingly steep. Escarpment outliers are rare or absent.

On the other hand, where the dip is low, as in the Mammoth Cave, Horse Cave, and Park quadrangles to the east, and in the Bowling Green North quadrangle to the west, the escarpment front is quite irregular, or at the extreme it is broken up into separate ridges and outliers. Solutional valleys and

Formation	Conjectured Subdivisions and Lithology	Thickness in Feet	Mapping Symbols on Plate 1	Associated Landforms (explained in text)
Ste. Genevieve Limestone	Predominantly oolitic limestone			"Low" sink plain grading upwards to escarpment front and knobs
	chert beds		solid line	
St. Louis Limestone	Predominantly limestone	40		Sink escarpment
	chert beds?		dotted line	Sink escarpment
	Predominantly limestone	50		
	chert beds?		dashed line	Sink escarpment; Dip-slope surface drainage disappearing beneath sink plain
	Predominantly limestone	80		Surface drainage with local, minor sink zones
	Limestone and Siltstone, variable chert.	marker horizon	dash-dot line	
		Approx. 40	Contact is approx., not shown on Plate 1.	
Warsaw-Salem				

Figure 6.

Conjectured lithologic subdivisions within the lower Ste. Genevieve and St. Louis formations. The outcrops of these units are mapped in Plate 2.

sink complexes on the Girken and upper Ste. Genevieve formations are common.

To a certain extent the position of escarpment outliers and escarpment reentrants is related to local anticlines or synclines. Solutional valleys tend to be developed in anticlinal flexures, whereas outliers are common in synclines.

Control by dip is less obvious on the sink plains, although the width of the plain is inversely proportional to the steepness of the dip. In areas of steep dip the topographic expression of the sink plain escarpments appears to be less pronounced than where the dip is low.

The degree of development of underground drainage on the sink plains may be a more indirect effect of rock dip. In limestone terrain the stability of subterranean drainage should be related to several factors, including length of underground flow, the physiographic relief, rock dip, and rock types and stratigraphic sequence. These last two factors act together as the major determinant of the degree of sink drainage. A well-developed subterranean drainage is presumably correlative with a high density of sinks and with deep, steep sinks (LaValle, 1965).

The St. Louis and lower Ste. Genevieve formations are composed of a diversity of rock types, in which limestones of varying texture and purity predominate. Chert stringers and beds are the most common secondary rock type, with minor argillaceous and dolomitic zones. Because of the heterogeneity of the limestones, some portions of the stratigraphic column are more soluble than others. This should result in an anisotropy in the "effective permeability" of the limestone to groundwater flow. Groundwater flowing parallel to the bedding can evenly dissolve the limestone along the more soluble layers. In contrast, when the groundwater is constrained to cross the bedding, the flow should be impeded by the less-soluble layers. Therefore, when the stratigraphic dip is approximately parallel to the slope of the water table, the development of subterranean drainage should be most favored. Thus heterogeneous limestone having a moderate dip toward a major stream should have dense sink drainage developed on it, while areas of higher or lower dip, or dip away from major surface drainage, should have fewer and shallower sinks, all other factors being equal.

The area to the southeast of Bowling Green along Drakes Creek may exhibit the above relationships. To the south of Bowling Green and west of Drakes Creek lies a broad area of "low" sink plain where the stratigraphic dip is away from the river (plate 1). Contrasting with this is the "high" sink plain directly opposite on the east side of Drakes Creek, where the dip is moderate and toward the river. A narrow strip immediately west of Drakes Creek also has "high" sink plain, presumably because of the high relative relief near the river. However, the above relationships may not be as exact an example of the effects of dip as the discussion indicates; much of the area to the south of Bowling Green is underlain by the middle part of the Ste. Genevieve formation, which supports a few sinks over the karst belt as a whole, while the lowest Ste. Genevieve and St. Louis formations exposed along and to the east of Drakes Creek are generally associated with "high" sink plain. Both rock type and rock dip may therefore be partial determinants of the contrast of landforms across Drakes Creek.

The drainage network is less obviously controlled by structural dip and fold patterns than are uplands and slopes. The Green River, the Barren River, and major tributaries to these rivers are not obviously controlled by local or regional structure. These rivers discordantly cross bedrock flexures and are not significantly aligned along either the dip or the strike. Low-order streams are more closely controlled by the structure (although they are naturally absent in sink areas). The small streams located south of the sink plains and draining into them are primarily dip-slope streams (see plate 1), although the largest of these disappearing streams are less directly controlled by the local dip.

The conclusion appears inescapable that a high degree of control of landforms by

stratigraphic and structural controls is present in the Kentucky karst area. The significance of these controls to hypotheses of landscape evolution through time is discussed in the following section.

Discussion

Inferences about the evolution of landforms through time may be possible from studies of the present relationships of landforms if appropriate assumptions are made. In the past 70 years the most common assumption has been that the geometrical relationships of hillslopes, summits, and drainage are determined in a simple, direct manner by the historical changes in the factors controlling erosion. Workers subscribing to this assumption have usually emphasized the delineation of upland flats and straths, whose origins were attributed to periods of complete or partial baseleveling. Their evidence has usually been drawn from map studies, with occasional stratigraphic evidence. The usual criterion applied to landforms to delineate erosional levels has been the presence of broad uplands or "remnants" of former upland surfaces which are sub-horizontal and discordant to structure. Particular attention was paid in the present study to testing for the presence of such topographically-accordant levels in the karst area.

Convincing evidence for sub-horizontal controls of uplands is lacking for both the Big Clifty escarpment and the zone of dip-slope drainage on the lower St. Louis formation. On both of these geomorphic units the general upland level closely follows the same geologic units downdip, with the highest topographic elevations corresponding to the furtherest updip extension of the capping strata.

On the other hand, most of the St. Louis-Ste. Genevieve sink plains are broad uplands of rather low relief which are inclined with the stratigraphic dip, but which are not inclined as steeply as the dip. On the average, the sink plains incline at about one-half to one-quarter of the stratigraphic dip. Thus the general pattern of the sink plains does not conclusively prove or disprove the presence of erosional levels, inasmuch as some structural control is present, but it is not the sole determinant of the general level of the sink plain.

"Cyclical" theories of karst development generally postulate the close association of sinks with erosional surfaces. Under this assumption, an unbiased test for any erosional levels associated with sink development would be the distribution of sinks with respect to the altitude of their occurrence. Therefore the frequency of sink bottom with respect to altitude was calculated for each traverse made across the karst belt (solid-line histograms in figure 5). Some of these traverses included sinks associated with the Big Clifty escarpment, but because erosional surfaces are supposedly structure-cutting, this should not compromise the results. The histograms have basically the same single-peaked form which is asymmetric toward higher elevations. A vertical zone of about 100 feet must be constructed for each traverse in order to include a sizable majority of sinks. This is a rather high relief for a postulated erosional surface supporting a karst cycle.

Accordance of hill summits is another criterion commonly used to distinguish erosion levels. The altitude-frequency histograms for hill summit elevations along the sample traverses is given in figure 5 as dashed lines. The hill summits follow the same type of distribution as the sinks, but with an even greater scatter of hill heights. The hill summits, as would be expected, have frequency peaks higher in elevation than those of the sink bottoms. This gives an even greater relief to a postulated erosional surface.

Even more damaging to the erosional-level hypothesis is the systematic increase in elevation of the histogram peaks from west to east across the karst belt from traverse (A) to traverse (F). It might be argued that this is resultant from tilting of an original erosional surface, but this represents a very high tilt for an area on stable shield basement.

A much less *ad hoc* explanation for the increase of the average sink and hill altitude to the east makes use of the progressively greater separation toward the east of the majority of the upland from major rivers (see plate 1). In the western parts of the study area both the Green River and the Barren River pass close to or through the sink belt. In the eastern part of the area, on the other hand, most of the land area covered by the traverses lies far from the Green River, the Barren River, or major tributaries to these.

It is perhaps a trite observation that in drainage basins of all sizes both the average elevation of the land surface and the height of divides increases the more remote the area from the master drainage stream or river. This is apparently resultant from an automatic adjustment within any drainage basin toward providing approximately equal relief throughout the drainage basin with the corresponding result that the average rate of erosion is made approximately equal throughout the drainage basin, lithologic and structural factors being equal.

More subtle techniques have been used in attempts to define erosion levels in present-day topography (*e.g.*, Clarke, 1966). Among these might be mentioned altitude-areas histograms and shoulder or col altitude-frequency histograms. While I have not applied these techniques to the central Kentucky karst area, I believe that these refinements would show little support for sub-horizontal erosional surfaces in view of the negative results of the above tests.

The absence of topographic evidence for the classic type of ersional surfaces in the karst belt does not preclude the existence of former episodes of baseleveling in this area, nor does it preclude the possibility that stratigraphic evidence for such periods might be found. It does, however, indicate that the inheritance of topographic forms from any such baseleveling has been negligible.

If the presence of inherited erosional surfaces of large scale within the karst belt thus seems doubtful, what are the alternative ways to look at landform evolution within the karst belt? One attractive alternative which has gained recent popularity proposes that erosion may have been relatively continuous over a geologically-long period of time under approximately constant climatic conditions, with the result that rates of downwasting of divides, slopes, and valleys would be approximately equal (subject to stratigraphic controls). The term "dynamic equilibrium" is generally associated with this thesis, and its main proponent is Hack (1960, 1965, 1966). The excellent correlation between landforms and structural and stratigraphic controls in the Kentucky karst area strongly support the idea of long-continued erosion leading to apparently complete adjustment between landforms and underlying geology.

The configuration of the Big Clifty escarpment most clearly shows the close control by geology which suggests extensive erosion under conditions approximating those of the present. In order to interpret historically the Big Clifty escarpment the only major assumption that need be made is that, as the escarpment retreats, approximately the same landforms are developed on the same geologic units. The present perfect correlation of the escarpment front with outcrops of the Big Clifty sandstone supports the hypothesis of continuous landform correlation with underlying geology.

Additional indirect evidence that this correlation between landscape elements of the escarpment and stratigraphy and structure has been maintained through a considerable period of erosion is afforded by the numerous rounded hills, or knobs, in the Girken and upper Ste. Genevieve formations which generally are positioned updip from the last outliers of the present Big Clifty escarpment, and which are especially numerous where the dip is low. These are designated in plate 1 as isolated areas of geomorphic unit 2. The Girken and upper Ste. Genevieve formations are usually exposed only in a narrow belt in front of the Big Clifty escarpment, suggesting that this limestone sequence is relatively easily eroded, and hence is rapidly stripped off once the protective cap of the Big Clifty sandstone is removed.

By analogy, it is postulated that these outlying knobs indicate the former presence of a caprock of Big Clifty sandstone. Many of these knobs lie up to tens of miles beyond the present outliers of the Big Clifty escarpment; these indicate that escarpment retreat with attendant landforms similar to the present ones has occurred over a considerable period of geologic time and during the removal of a considerable volume of rock.

Residual knobs are present at a greater distance from the escarpment front in areas of low stratigraphic dip than in areas of high dip. Assuming a sub-equal rate of removal of residual knobs upon stripping of the Big Clifty caprock, then it follows that: 1, the rate of horizontal retreat of the Big Clifty escarpment is more rapid the more shallow the dip; 2, escarpment retreat occurs in a direction approximately normal to the structural contours; and 3, reentrant valleys and solutional valleys tend to form first in anticlinal flexures, while outliers tend to persist in synclinal flexures. These relationships are apparent in plate 1.

Similar generalizations can probably be applied to the more muted escarpments on the Ste. Genevieve-St. Louis sink plains and to the dip-slope cuesta with surface drainage to the south of the sink plain, but these relationships were not investigated in detail.

It would be mistaken to conclude that the evidence for fairly continuous erosion in the study area precludes changes in landforms through time. Some changes may be required in adjustment to differences in stratigraphy and structure as erosion continues. Likewise, changes of climate and fluctuations of local baselevel may have required past adjustments of landforms on a limited scale.

Landform changes of the first type are dictated by the horizontal migration of landform belts with maintenance of structural and stratigraphic controls as erosion continues. Any point in space may be associated with landforms which are in turn: upland above the Big Clifty sandstone; escarpment front on the Girken and Ste. Genevieve formations; sink plain on the lower Ste. Genevieve and St. Louis formations; dip slope surface drainage on lower St. Louis limestone, etc.

The rate of vertical downwasting at any one point is not uniform, but changes with the stratigraphic unit at the surface. The geometry of an escarpment dictates that the upland behind the escarpment front is lowering more slowly than is the pro-escarpment rampart. "Resistant" beds tend to act as a local base level to the less-resistant strata above them, producing low-relief landforms on top of the escarpment. In complementary fashion the weaker units below the escarpment cap are held high above the general land level by the caprock. Thus when the retreating cap unit exposes the underlying weaker units, the weak units develop steep slopes and are eroded rapidly toward the general level of the surrounding land. Such non-uniform rates of erosion are presumably characteristic of the Big Clifty escarpment, the sink-plain escarpments, and the dip-slope cuesta on lower St. Louis beds.

In addition to these "local" fluctuations in rates of vertical downwasting, stratigraphic and structural controls may introduce larger-scale historical changes in landform evolution. For example, the breaching of a resistant layer along a river might lead to intrenchment upstream. With the Kentucky karst area the only such geologically-controlled landform change of regional extent which has been identified resulted from changes in patterns of flow of subterranean drainage.

In the area lying approximately between Smiths Grove on the west and to beyond Horse Cave on the east the drainage from the dip-slope on the lower St. Louis formation, from the St. Louis-Ste. Genevieve sink plains, and from parts of the Big Clifty escarpment complex presently flows as subterranean drainage to the north and west beneath the sink plain and Big Clifty escarpment to exit into the Green River (Cushman, 1966). The present drainage divide between the Green and Barren drainage networks lies asymmetrically close to the Barren River. This pattern of groundwater flow could not have been present before the Green River had cut below the sandstones and shales of the Big Clifty and overlying formations. Until such a time, groundwater flow north to the Green River would have been impeded by the shaly beds. Subsequent to the incision of the Green River into the Girken formation, a strong, unimpeded groundwater gradient through the underlying limestones was established, and subterranean drainage to the north was initiated.

This mechanism is similar to the tapping of static groundwater proposed by Gardner (1935). According to the mechanism proposed here, the degree of subterranean drainage development should be directly related to the gradient of the flow through soluble layers and the absence of barriers to flow. A suitable measure of the degree of subterranean drainage is probably the density of sinks (other stratigraphic factors being equal).

Within the area discussed above, the establishment of subterranean drainage north to the Green River was probably, therefore, accompanied by the development of an increased number of sinks, and probably, on the average, of steeper and deeper sinks. The initiation of northward-flowing subterranean drainage to the Green River presumably occured earlier in the eastern part of this area than to the west because of the slight tilt of the beds to the west. Northwest-trending subterranean drainage may be in the process of establishment or enlargement in the area to the northeast of Bowling Green.

The outcrop pattern of the lower sink-supporting horizon in the St. Louis formation (dashed lines, plate 2) offers possible indirect support for this proposed sequence of events. Apparently this unit both underlies the sink plain to the northwest and caps some of the cuesta ridges to the southeast. Further to the east of plate 2 this unit caps even larger tracts of surface-drained areas while supporting a sink plain further to the north. This situation could not have arisen if erosion through time had produced only a continuous dipward migration of landform belts where each geologic unit constantly supported a characteristic landform. Presumably, before initiation of subterranean drainage to the northwest, landforms over the outcrop belt of the St. Louis and lower Ste. Genevieve formations within the area described above were either surface drainage or "low" sink plain. Upon initiation of a strong water table gradient to the northwest, intense sink drainage probably developed in the area immediately adjacent to the Big Clifty escarpment, while further away from the Green River on the dip slope of the St. Louis cuesta surface drainage became or remained the stable hydrologic regime. The lowest sink plain unit on the St. Louis formation presumably underlay parts of both of these zones. Stratigraphic dip, distance from the Green River, original drainage patterns, and types of geologic units exposed at the surface are major factors which determined the difference in drainage patterns between the present sink plain and the dip-slope cuesta.

*Quinlan and Pohl (1966) have detailed what they they feel to be an ancestral surface drainage network generally flowing to the northwest within this area on the basis of the present topographic relationships of ridges and lowlands. Therefore this would presumably represent the drainage conditions prior to initiation of subterranean drainage to the Green River.

The concept of "dynamic equilibrium" in its most extreme formulation includes the postulate that landforms are so completely adjusted to the present climate and to the local base level that the landscape contains few, if any influences from or indications of past changes of climate or base level. This is probably too strong a restriction. Although the landforms of the central Kentucky karst area do not seem to contain a record of past episodes of baseleveling, and a long-continued erosional history seems indicated, it should not be ruled out *a priori* that the area has not been subjected to fluctuations of local base level and/or climatic changes which have left their imprint on present-day landforms.

The topographic expression of landforms is a notably poor indicator of past changes of environment. Much more information about changes of erosional processes is present in soils, alluvial deposits, caves, and cave fills than is present in topographic forms. For example, Ruhe (1966), working with soils and alluvial deposits, has demonstrated a striking sequence of gullying, alluviation and stability in Iowa due presumably to Pleistocene and Recent climatic changes which are belied by the simplicity and regularity of the topography. Such changes are not clearly expressed in the topography both because landform modifications occasioned by moderate changes of climate or local base level take place without changes in the position of drainage divides or drainage basins, and because changes of landforms in response to changes in environmental factors tend to obliterate the topographic forms characteristic of the former conditions either by burial or erosion.

Caves and their deposits, by viture of their protected position and sensitivity to changes of base level and climate are excellent potential indicators of detailed erosional history. Many caves form in a restricted vertical zone surrounding the basal water table. This basal water table is often closely controlled by the regional base level. Levels of stability of the basal water table should then be reflected in a high frequency of cavern levels (*e.g.*, studies by Sweeting (1950), Davies (1957), White (1960), and Wolfe (1964).

In summary, the central Kentucky karst area has landforms which are closely controlled by stratigraphy and structure. The karst area has undergone a long period of fairly continuous erosion in which the broad patterns of landform evolution have been determined by geologic factors. Presumably superimposed upon the general history of erosion are periods of fluctuation and stability of local base level and climate. These details of erosional history must be studied through examinations of soils, alluvial deposits, caves, and cave deposits, because they have not left a simple record in the present topography.

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Dept. of Geography
217 Cabell Hall
University of Virginia
Charlottesville, Virginia 22903

Dynamics of Sediment Transport In Limestone Caves

By Elizabeth L. White* and William B. White**

ABSTRACT

Groundwater moving through maturely karsted limestone aquifers may carry, in addition to a dissolved load extracted from solution of the wall rock, clastic material as suspended load or as bedload. All of the insoluble residues from the solution of the limestone and in some cases, large quantities of material from overlying or adjacent clastic rocks must be transported out of closed drainage basins by the action of cave streams. Portions of the transported material are deposited en route to form the richly varied clastic cave sediments. Application of standard engineering formulae for sediment transport indicate that flows with a threshold velocity on the order of tenths of a foot per second are necessary to transport the coarse sediments. Suspended load is important in the transport of fine sediments but requires flows at least in the turbulent regime. A tentative conclusion is drawn that clastic load transport by fast-moving water is an integral part of the development of many karst drainage nets and that extensive development of integrated drainage nets by percolating waters is unlikely.

INTRODUCTION

Cave passages in limestone are conventionally believed to have formed by the slow percolation of groundwater below the water table with the removal of limestone by transport in dilute solution. Much has been written about the limestone solution process and it is probably fair to say that its chemistry is moderately well understood. However, the solution process cannot provide a complete explanation. In nearly all caves one finds a sequence of clastic sedimentary deposits which must have been carried to their site of deposition by mechanical transport. These clastic sediments are rather glossed over in most American writings on caves. They tend to be regarded as "secondary deposits" transported by later high gradient streams that have little to do with the primary process of excavating the cave. Although many writers from Davis onward have considered the effect of mechanical abrasion (corrasion) in the enlargement of cave passages, they do not seem to have considered whether the power needed to mechanically transport the clastic material out of the system was compatible with the postulated slow percolating flows. The exception is Davies who recognized as long ago as 1957 that cave sediments are much more complex than "unctuous clays" derived from local limestone weathering. The data of his papers (Davies, 1957, Davies and Chao, 1959) will form an important part of the present discussion.

Our object in this paper is to examine the role of mechanical transport in the excavation of caves and the development of drainage nets in limestone aquifers. We will attempt to show that the clastic sediments play an important part in the primary development of the drainage conduits and in the reduction of the land surface in certain types of karst terrains.

*Cave Research Foundation
**Materials Research Laboratory and Department of Geochemistry and Mineralogy, Pennsylvania State University, University Park, Pennsylvania 16802

Classification of Cave Sediments

There is no generally accepted classification of cavern sediments. European workers, mainly Schmid (1958) and Kukla and Lozek (1958) have been concerned with the paleoclimatic significance of cave sediments and thus they tend to emphasize the "entrance facies" which consist in large part of hill slope debris in the entrance talus fan. Naturally it is in this facies that one is most likely to find animal or human remains. In the present study we are concerned with the hydrologic significance of cave sediments and thus are interested mainly in the "interior facies." Frank (1965) separated the "interior facies" sediments into "tardigenic" (slowly deposited) and "torrenigenic" (rapidly deposited) types but this separation does not seem to be the most useful for present purposes. In an earlier paper (White, 1964, in press) a petrological classification was proposed as follows:

Clastic Sediments

Autothonous Deposits
- Breakdown
- Weathering Detritus
- Organic Debris

Allothonous Deposits
- Transported clay, sand and gravel, loess

Chemical Depoits

Carbonates
Evaporites
Iron and manganese hydrates
Ice
Phosphates

Of these clastic sediments, breakdown, being mostly limestone can be dissolved and removed in solution. Organic debris generally does not accumulate until late in the cave's history and so is of minimal importance in the early solutional stages. The weathering detritus from the limestone may consist of clay, sand, silicified fossil fragments, chert nodules, shale interbeds and other insoluble debris. In addition to the local load which must be carried out of the aquifer by mechanical transport, there is a distantly derived mechanical load which originates in other parts of the basin drained by underground routes. Some of the sources of sediment load are outlined schematically in figure 1. These materials are observed in the cave passages and are listed under the heading of allothonous deposits.

Significance of Sediment Transport in Limestone Terrains

When treating cave development one must consider the entire karst drainage basin. Usually only a portion of the basin will be on

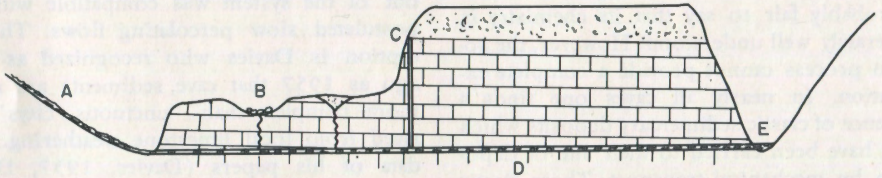

Figure 1.

Various sources of sediment load in a cave system occurring in a heterogeneous aquifer of flat-lying rocks.

A. Sediment transported from a distant source by a sinking surface stream.
B. Surface soil infiltrating through sinkholes.
C. Debris from overlying clastic rocks falling directly into the channel through vertical shafts.
D. Insoluble detritus from the limestone wall.
E. Sediment from base-level backflooding of the nearby river.

limestone; in many cases a portion will be on clastic rocks. This is particularly true in the karst aquifers of the Appalachian Highlands and the Interior Plateaus. There are, then, three sources of clastic load to the subterranean stream: (i) insoluble material in the limestone which supports the cave and insoluble beds within the limestone section (ii) insoluble rock debris from clastic rocks overlying the underground drainage system and (iii) eroded material from adjacent clastic rocks which must pass through the subterranean drainage system to escape from the basin. If long surface streams in the headwaters must drain through subsurface routes near the mouth of the net, the transported load can be considerable. We shall illustrate each of the three situations described above with real examples.

All limestone contains a certain percentage of insoluble residue but perhaps the most extreme example of extensive cavern development in inpure limestones is the Chestnut Ridge group of caves in the Loyalhanna Limestone in Western Pennsylvania. The Loyalhanna limestone is actually a sandstone with a sparry calcite cement. Published analyses give 50-70% insoluble residue of which most is quartz sand with a grain size of 0.2mm. And yet the Loyalhanna is quite cavernous with caves such as Bear, Copperhead, Coon and Laurel Caverns. The caves have sandy floors and a thick weathered layer on the walls but most of the usual evidence of fast-moving water such as scallops and deeply entrenched canyon cross-sections are absent. Only Bear Cave takes a perennial stream. A small stream flows through the lower level of Laurel Caverns and has cut a narrow trench in the wide main passage. However, Bear and Coon caves have a network pattern, the classical evidence for slow phreatic circulation. Whatever the flow regime, enough momentum must have been present to allow for the transport of some 50% of the cavern volume as mechanical load.

Many of the limestone aquifers of the Appalachian Plateaus from West Virginia to northern Alabama and in the Interior Plateaus area are capped with thick sequences of Mississippian and Pennsylvanian shales and sandstones. The normal processes of slope retreat insure that the weathering debris from all of these rocks will move downward into the limestone valley below. Many of these limestone valleys are so arranged that the only route by which this clastic material can leave the basin is through the cave system.

One such example is the Central Kentucky Karst in the region near Mammoth Cave and the Flint Ridge cave system as illustrated in figure 2. Here the main baseleveling stream is the Green River which is entrenched in a deep canyon in the sandstone-capped Mammoth Cave Plateau. The river is separated from the main recharge area by the Chester (Dripping Springs) Escarpment. All drainage from the Sinkhole Plain flows to the north through the lower levels of the big cave systems to springs on the Green River. All debris from the retreat of the escarpment must have been flushed through the cave system since no surface streams penetrate the area. The amount of sediment remaining in the abandoned higher levels of the cave system cannot account for the tremendous quantities of rock which have been removed. Watson (1966) discusses this point at greater length.

A second closed basin example is the Newsome Sinks in the plateaus of northern Alabama (fig. 3). This karst valley is entirely enclosed by clastic rocks except for a narrow underdrained saddle at the mouth. Here again all material from the retreat of the plateau edges must be transported through the underground routes. Varnedoe (1963) develops a convincing argument that the underground routes have been active for a long time.

In addition to disposing of its own insoluble debris and its overburden the karst drainage net must often carry a large load brought in by sinking streams which flow from adjacent clastic rocks. The basin of Hills Creek is shown in figure 4. Only about 30% of the basin is floored by limestone and much of the Hills Creek basin is a high gradient valley on sandstones and shales. All of the clastic load carried by Hills Creek and Bruffy Creek from the erosion of the Allegheny Mountains is transported

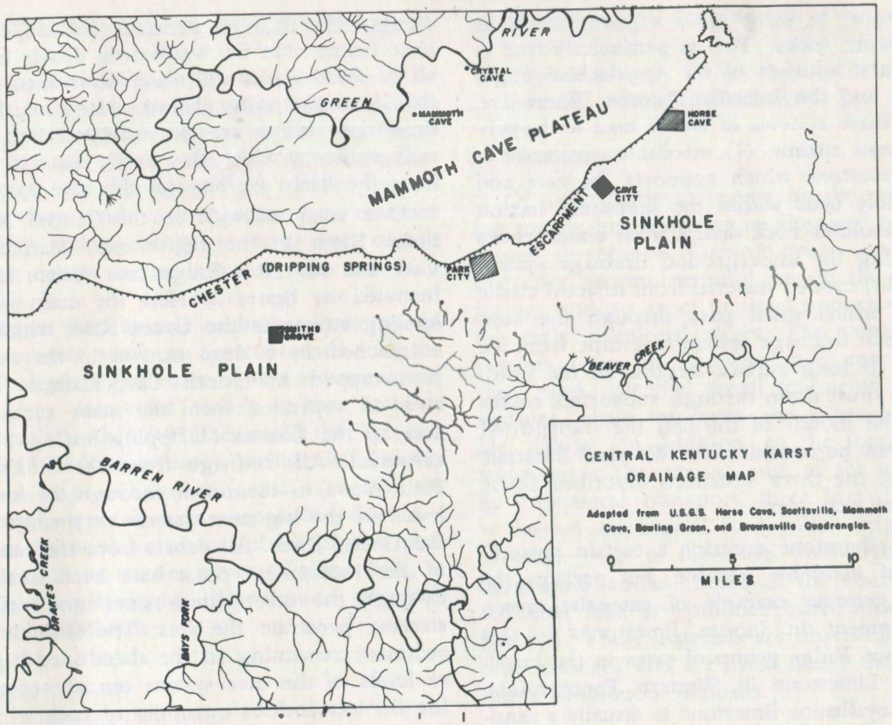

Figure 2.

Drainage basin of the Central Kentucky Karst.

through the cave systems to Locust Spring. As indicated by White and Schmidt (1966) this subterranean channel is flooded over a large portion of its route.

The conclusion seems inescapable: that in these and other similar basins, much clastic material has been and is now being transported through the cavern system. The cave systems in these examples do not exhibit high gradient free-surface streams. Most of the transport seems to take place by conduit flow at or below the present water table.

Description of Cave Sediments

Quantitative descriptions of "interior facies" sediments are extremely rare in the literature. The outstanding work of Schmid (1958) and of Kukla and Lozek (1958) deal mainly with the entrance facies and these results are not applicable to the present discussion. The same may be said for other studies of the European Paleolithic. The data to be used in our calulations are taken from Davies (1957) Davies and Chao (1959) Brain (1958), Bögli (1960), Deike (1960), Helwig (1964-1965), and Frank (1965). These data are summarized below. Every attempt has been made to select data which are representative of sediment moving laterally through main drainage conduits. Localized concentrations of extremely coarse sediment in the vicinity of vertical shafts and terminal breakdown have been excluded.

Davies (1957) measured 37 feet of stratigraphic section in Cumberland Caverns between the Henshaw passage and the Volcano Room. Interbedded throughout the section are "pebble gravel" beds 1-2 feet thick with a diameter of 12-35 mm. This section of

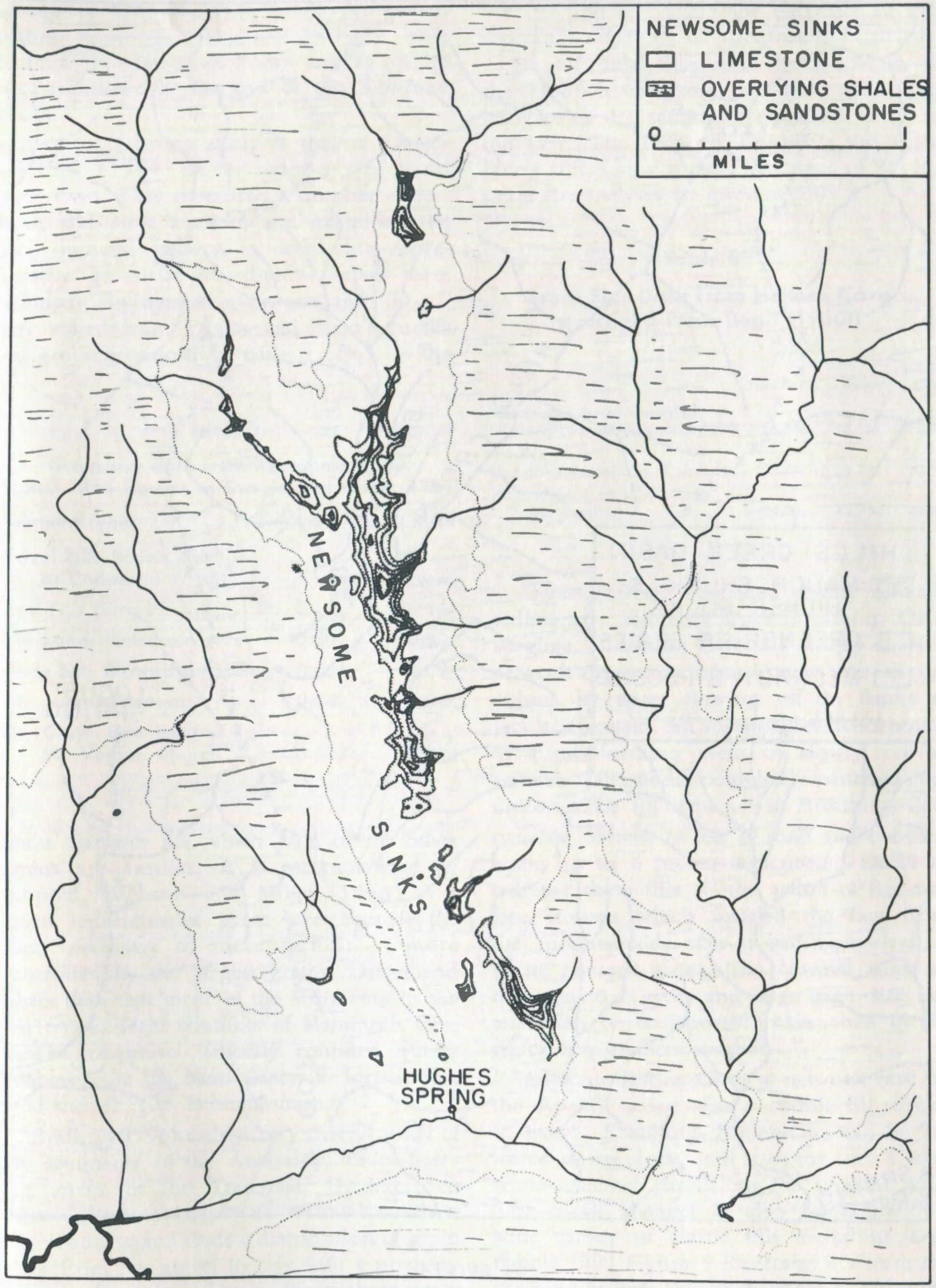

Figure 3.

Drainage basin of Newsome Sinks.

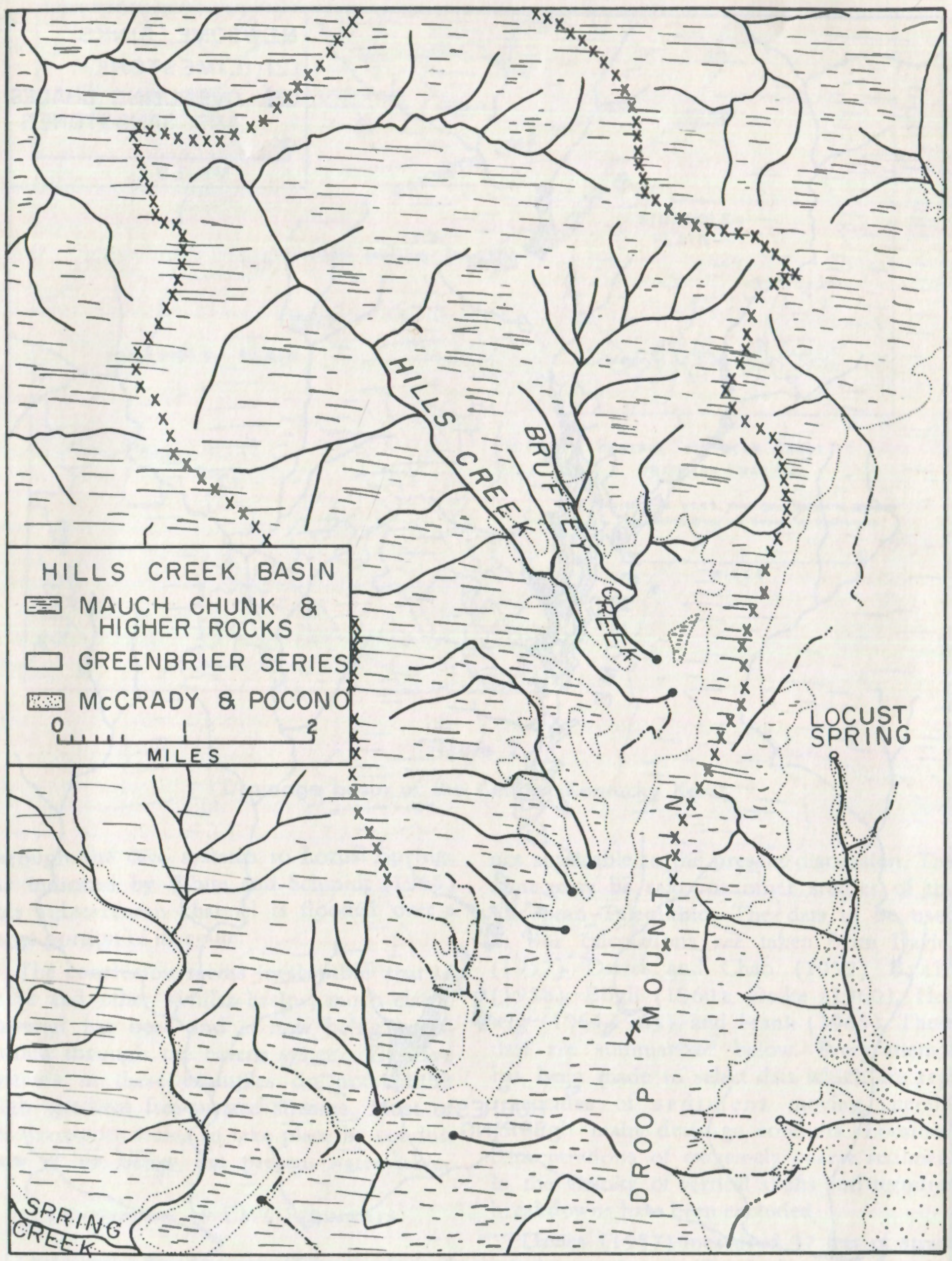

Figure 4.
Drainage basin of Hills Creek.

passage is part of a main trunk conduit extending from the Henshaw Entrance back through the Ten-Acre Room and is lost in the breakdown at the end of the Ten-Acre Room.

A most valuable study is that of Davies and Chao (1959) on the sediments of Mammoth Cave. They measured a number of sections, did sieve analyses and examined the heavy mineral fraction in detail. Six representative grain-size distribution curves were included. The median grain size and "D_{84}" were scaled from Davies and Chao's curves and are reproduced in table 1. D_{84} is the

Table 1

Grain size data from Mammoth Cave (Scaled from figures in Davies and Chao, 1959)

Sediment type	D_{median}	D_{84}
Gravel Fill: Boone Ave at Chaperon	5 mm.	18 mm.
Sand Fill: Sims Pit Ave.	0.18	0.26
Silty Sand: Pensacola Ave	0.10	0.30
Sandy Silt: Richardson Spring	0.025	0.09
Silt: Simms Pit Ave.	0.04	0.08
Silty Clay: Sink west of Mt. Vernon Church	0.0034	0.023

grain diameter for which 84% of the other grains are smaller. It is recommended by Leopold, Wolman and Miller (1964) as a more representative grain size because the shear necessary to move bedload is more controlled by the largest grains. Davies and Chao find that most of the sediments in the big trunk drain conduits of Mammoth Cave consist of gravel (mostly rounded quartz pebbles from the basal Caseyville formation), sand and silt. Clay is uncommon.

Brain (1959) made a very careful study of the sediments of the Australopithecine-bearing caves of the Transvaal. He interprets these sediments largely as residual weathering products and finds a distribution of grain sizes from the gravel to clay with a predominance of sand and silt. Since these caves do not exhibit conduit-like passages, they are less pertinent to the main argument here but do illustrate again the necessity of mechanical transport of the insoluble debris.

Bögli (1960) discussed the sediments of the Hölloch Cave and presented grain size analyses of 19 samples from three areas of the cave. The Hölloch sediments are again sandy silts and silty clays. Averages of Bögli's grain size analyses are given in table 2.

Table 2

Grain Size Data from Hölloch Cave (averaged from Bögli (1960))

	Fine Sand	Silt	Clay
Floodwater Zone (4 samples)			
Styx and Grosser Burkhaltersee	56.8%	30.1%	13.8%
Inactive Zone			
Domgang-Regenhalle (5 samples)	21.0%	55.7%	23.7%
Himmelsgang and SAC-Gang (10 samples)	34.8%	43.7%	21.6%

Deike (1960) described a very unusual sedimentary sequence from Breathing Cave, Virginia. Breathing Cave is a steeply dipping network apparently formed under artesian conditions by water flowing off the flanks of Jack's Mountain. All solutional evidence points to a cave-forming event by slowly moving artesian water under completely saturated conditions. The fill sequence in Breathing Cave consists of tens of feet of sand and cobbles, some up to 6 inches in diameter. Deike attributed these fills to the action of free-surface streams which invaded the cave from the mountainside after it had been drained of its phreatic water. The channel slope in this system is steep and large grain-size material can be transported easier than in the typical low gradient conduit.

Adjacent Butler Cave, a network cave on the syncline slope, has a cobble fill similar to that in Breathing. Butler Cave can be followed down slope until it opens into a long trunk channel paralleling the syncline axis. The trunk channel is also floored with a wide variety of clastic fills including large cobble fills. Figure 5 illustrates a stream deposit cut by later stream channeling. Although this system is more complicated and all conclusions regarded as tentative, one would still

Figure 5.

Typical cobble deposit in Butler Cave, Virginia.

suspect that clastic material has always moved through the system even when it was water filled. A much more detailed study of this system and its sediments is warranted.

Helwig (1964,1965) reported on the sediments of Carroll Cave, Missouri, which is a rather different situation. Carroll Cave consists of a pair of long low-gradient trunk channels in which a piracy has taken place. There is no overlying caprock or adjacent clastic basin and recharge is derived from the diffuse influx of water in an overlying karst terrain. The sedimentary section contains much chert gravel thought to be derived from the Roubidoux and Jefferson City formations which overlie the Gasconade dolomite in which the cave is formed. Although there are numerous possible local sources for the sediment, transport out of the system is by the cave stream.

Frank's (1965) study of the sediments of a selected group of caves in Texas provides a completely different type of material. In the Texas caves clays are the bulk constituent of the fills; transported sands and gravels are uncommon. In the Edwards Plateau there is neither capping nor adjacent beds of clastic rocks and only transport of the residual detritus from the limestone is necessary. The sedimentary sequence of the Texas caves differs so markedly from the temperate climate sediments of the Appalachian and Interior Plateau caves that further discussion will be limited to caves in those eastern provinces.

Mechanism of Sediment Transport

For more than a century the movement of sediments and the stability of sedimentary channels has been of importance to civil engineers concerned with rivers, harbors, and canals. A vast literature has been built up on this subject which we make no attempt to review. The theory of sediment

movement as it is presently understood is found in a number of texts of which we chose Chow (1959), Leopold, Wolman, and Miller (1964) and Lindsley et al. (1949) mostly on the basis of convenience. The equations used for deriving the threshold of motion of the sediment bed were taken from a very recent task committee report of the American Society of Civil Engineers (Vanoni et al. 1966).

The movement of solid material by flowing water takes place by two mechanisms: the material may move in turbulent suspension in the stream (suspended load) or it may be dragged along the bed by the shearing action of the water above it (bedload). Bedload is the more important and will be considered first.

Bedload

The bed material under a stream is subjected to a shearing action by the water flowing over it. The only requirement for the presence of the shear is that the water be in motion; the flow may be either laminar or turbulent. However, there is a resistive force caused by the weight of the particles lying in their sockets in the bed which opposes the shear force. Only when the shear force becomes great enough to rock the particles from their equilibrium positions will the particles move. There exists, then, a critical shear stress below which the particles will not move and above which the bed is set in motion.

Vanoni (1966) expresses the relationship of the critical boundary shear to the mean sediment diameter as

$$\frac{\tau_c}{(\gamma_s - \gamma)\, d_s} = f\left(\frac{U_{*c}\, d_s}{\nu}\right) \qquad (1)$$

where: τ_c = critical boundary shear
γ_s = specific gravity of sediment
γ = specific gravity of fluid
d_s = mean sediment diameter
$U_* = \frac{\tau_c}{\rho}$ = "friction velocity"
ν = kinematic viscosity of fluid

Both sides of the above relationship are dimensionless. The left side can be thought of as a particle Froude number and the right side as a function of boundary Reynolds number. The relationship between the two was determined experimentally long ago and is known as the Shields' Relation. The Shields' curve is reproduced from Vanoni (1966) in figure 6. The straight line portion on the left describes very fine grained sediments

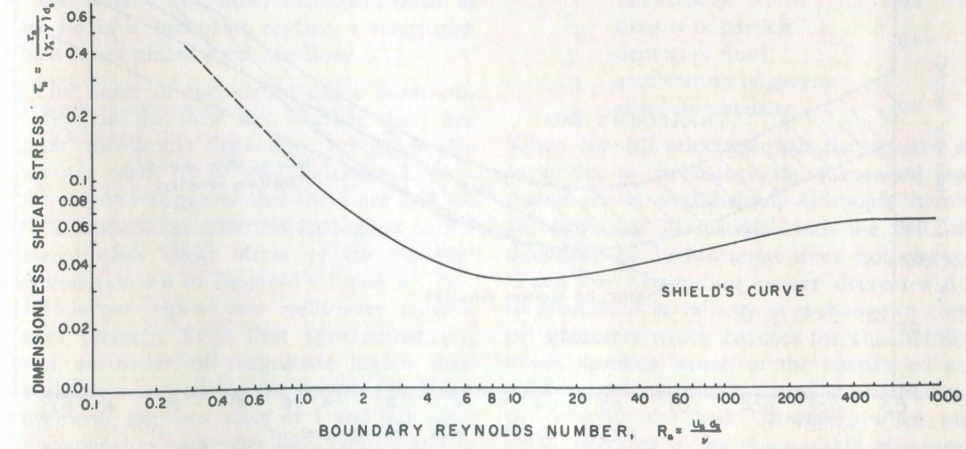

Figure 6.

Shield's Curve (from Vanoni, 1966)

where cohesive forces begin to play a role and where the surface roughness of the channel is completely immersed in a laminar boundary layer. At the extreme right, the relation again straightens out as the surface of the bed is exposed to totally turbulent flow.

Although the threshold of bed movement is rather well described by a critical boundary shear, it is not easy to convert this into a critical threshold velocity. This difficulty arises partially because of the ambiguity in calculating bottom velocities from average channel velocities. The average velocity always occurs at a constant fraction of the depth but as equation (2) below will show, the velocity profile near the bed depends only on bed roughness factors. Thus channels of different flow depth may have the same boundary shear and the same velocity profile but will have different average flow velocities.

One method to estimate the critical flow velocity is to use the velocity profile for two-dimensional free-surface flow over a flat bed as given by:

$$\frac{u}{U_*} = a_r\left(\frac{U_* k_s}{\nu}\right) + 5.75 \log \frac{y}{k_s} \qquad (2)$$

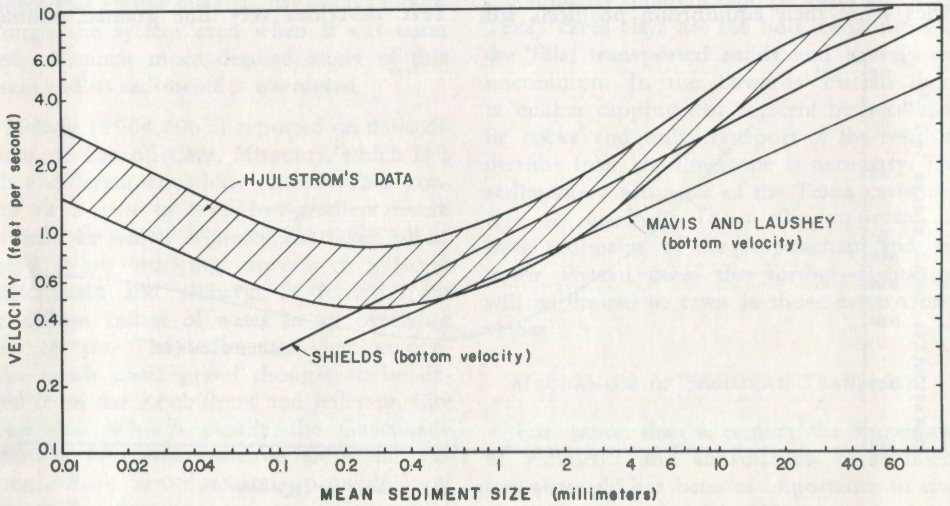

Figure 7.

Critical threshold velocity curves (from Vanoni, 1966)

where: u = velocity at distance y above the bed

k_s = characteristic roughness size

The function a_r is a function of the boundary Reynolds number. When R_* is greater than 90, $a_r = 8.5$ and remains constant. When R_* is less than 3.5 the boundary is hydrodynamically smooth and the function is given by:

$$a_r = 5.5 + 5.75 \log \frac{U_* k_s}{\nu} \qquad (3)$$

One then substitutes the grain diameter to get the velocity at the edge of the boundary layer and uses the Shields' Relation to relate U_* and R_*. The result of this is shown on figure 7 as the curve marked "Shields".

A second estimate can be obtained from the empirical Mavis and Laushey (1949) equation:

$$u_c = 0.5\left(\frac{\gamma_s}{\gamma} - 1\right)^{1/2} d_s^{4/9} \qquad (4)$$

Their results are shown on Figure 7.

A third result is obtained in the Vanoni report by plotting a set of experimental data on the same graph as the equations and these results fit within the band labeled "Hjulstrom's data" in Figure 7. The Hjulstrom curve shows a pronounced minimum at a mean grain size of 9.2 mm which corresponds to a critical velocity of 9.4 feet per second. The rise in the curve on the left side of the diagram is again due to the cohesive forces in the sediments becoming dominant.

A conclusion can now be drawn: For sediment to move as bedload in an underground channel, the flow velocities must be on the order of tenths of a foot per second. Sediment transport will not take place at lower velocities. The distinction between pipe flow in a flooded channel and open channel flow will change the velocity profile and thus the details of the transport but will not change the order of magnitude of the effect since only the bottom velocity is involved.

Suspended Load

Suspended load is that portion of the transported sediment which is carried in suspension in the water. The restrictions on suspended load are somewhat more severe at large grain sizes than are the restrictions on bedload. Since the particles are heavier than water, the net downward pull of gravity must be counteracted by an upward force which is provided by the turbulent eddies in the water. The flow, therefore, must at least be in a turbulent regime, a restriction which is not placed on the bedload.

The onset of turbulence sets a minimum velocity on the flow but whether there are higher thresholds depending on grain size does not seem to be well understood. Bagnold (1966) suggests that there are and offers a suspension criterion analogous to the dimensionless shear stress of the Shields' Relation (shown in Bagnold's figure 8). Particles larger than one millimeter require shears (velocity to a first approximation), about an order of magnitude higher than the shears required for the onset of bedload movement. Between sizes of 1 and 9.1 mm, the suspension parameter falls rapidly and at grain sizes near 0.1 mm the shear needed to suspend a particle is on the same order as that needed to move it in bedload.

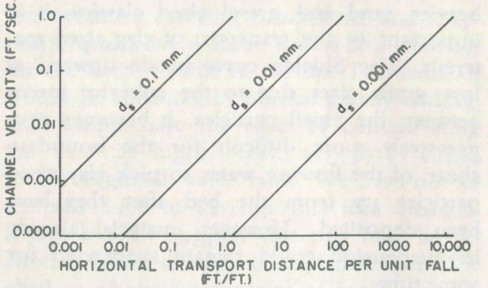

Figure 8.

Travel distances for small particles before settling out of a suspended load. Calculated from Stokes law assuming water at 50° F. and a particle density equal to that of quartz (2.65).

The description of the suspended load is unusually given in terms of the fall velocity of a particle which is described by Stokes law:

$$v_f = \frac{2(\rho_g - \rho)\, g\, d_s^2}{9\mu} \qquad (5)$$

v_f fall velocity
ρ_g density of particle
ρ density of fluid
g acceleration of gravity
μ absolute viscosity

When the fall velocity equals the upward velocity due to turbulence, the suspended sediments are in equilibrium. Although there is an exchange of material with the bed, the thickness of bed-material does not change. When the suspending power decreases due to a decrease in velocity or a change in channel geometry which reduces the channel Reynolds number, some of the suspended material settles out and is added to the bedload or thickens the bed. However, when very small particles in the clay and silt size range are in suspension, it takes a long time for the material to fall out and the load can be transported a long way before the water clears.

This situation is illustrated roughly in figure 8.

Although the suspended load plays a relatively minor role in the transport of the heavier sand and gravel sized clastics, it is important in the transport of clay sized materials. The Shields curve bends upward at low grain sizes due to the cohesive forces between the small particles. It becomes progressively more difficult for the boundary shear of the flowing water to pick clay sized particles up from the bed after they have been deposited. However, material already in suspension could remain suspended for some time.

Material from the overlying soil horizon settling into the conduit through very small fractures and joints could contribute to the suspended load. If this material settled directly into still water (or water moving at the so-called groundwater velocities of feet per year) one would expect a relationship between the thickness of the deposited beds and the location of the open joints. Such a relationship has not so far been observed but it must be admitted that careful observations in search of this effect have probably not been made. The non-existence of this relationship would suggest that at least some reworking of the material derived from the soil horizon does occur and that there is some lateral transport by suspension in the moving water of the conduit.

Another source of suspended load is in the headwaters of the sinking streams that feed into the karst aquifer. These streams usually have higher gradients in their headwaters and thus can pull more material into suspension. When the waters reach the low gradient conduits of the cave system, the velocities will decrease both because of decreased gradient and because of the transition in some cases from channel flow to pipe flow. The coarser sized material will settle out very quickly and continue its journey through the cave as bedload. The very fine material will remain in suspension for a much longer time because of its lower fall velocity and some of it may be transmitted entirely through the aquifer system in suspension.

This mechanism may be of importance in explaining the silting up of side passages and blind passages which have been blocked by breakdown. If fast moving water carrying a fine-grained suspended load is backponded into these passages, the velocities decrease and there is time for the suspended load to fall out.

The same applies to water carrying a sediment load which is back-flooded from surface streams. Collier and Flint (1964) have called attention to the silting of the lower levels of Mammoth Cave which they attribute to silt and clay carried into the cave from the backflooded waters of the Green River. Watson (1966) has questioned this mechanism and proposed that the silt is obtained from the sediment load being carried westward from the Sinkhole Plain. The mechanism discussed above favors the Green River as a source. The Sinkhole Plain is some miles to the east of the Echo River portion of Mammoth Cave and the waters reach the low gradient portion of their course several miles upstream from the sampling points. The flow regime in the lower reaches of the underground conduits may be below the suspended load threshold and most of the sediment in suspension would have been dropped somewhere upstream and would be moving as bedload in the downstream reaches. The Green River is about half a mile from the Echo River sampling points and contains a large suspended load in high flood. When these waters backflood into the cave velocity falls rapidly and much of the suspended load would have an opportunity to drop out. Thus there may exist in Mammoth Cave the situation in which sediment is being added to the cave at the top of the floodwater zone at the same time that a much larger volume of sediment is being transported through and out of the cave at the base of the main water-carrying conduits. The data and calculations are both too tenuous at the moment to attempt to resolve the argument. Measurements of grain size distributions as a function of distance from the Green River could provide some critical evidence.

The over-all importance of suspended load in the transport of sediments through cave channels cannot be evaluated easily. Cave channels typically have low gradients and thus suspended load cannot play a very important role in the transport of sand and gravel sized material. It may play a role in the transport of clay sized material particularly if the travel distances are relatively short. Most limestone springs are clear during most of the year and are only muddy if at all during high floods. Locust Spring, the main discharge point of the Hills Creek Valley has been observed by the writers to be extremely muddy during high flow. This is a short reach of underground stream which is fed by a long reach of high gradient surface stream and it might be a good example to use for quantitative measurements.

Implications and Discussion

The demonstration that mechanical transport plays a role in all stages of cavern development and that there exist moderate velocity requirements to exceed the threshold of sediment movement has certain implications for other aspects of cavern development.

The aspect of cavern development about which least is known is the early stage between the un-modified joint or bedding plane and the solution conduit of enterable size. It does not seem likely that sediment moves through the original secondary mechanical openings in the limestone because the flow velocities through these small crevices would be too low to exceed the threshold of transport. Thus groundwater in the initial stages would seep through the proto-cavern routes and the sediments would be filtered at the surface and constrained to move by surface routes. At some point in the development of these initial openings the diameters will become large enough to permit some sediment movements and at least the fine grained material will begin to work its way into the solutionally enlarged openings of the aquifer. If there do exist flow patterns curving to various depths in these initial flows where the deviation from Darcy flow has not yet become overwhelming, the longer, more deeply curving paths would be expected to have the lower velocities because of higher wall frictions. These paths would also have less sediment carrying competence and it appears, qualitatively, there would be a tendency for the deeper paths to silt up. The tendency would be enhanced by normal gravity settling, and would have the effect of concentrating the flow in higher shallower paths closer to the regional water table. We have not attempted here to develop this idea quantitatively, but do suggest that this may be yet another factor to offer in explanation for the shallow phreatic movement of karst groundwater. Meisler (1963) also used sedimentation at depth to explain the shallow occurrence of solution openings.

The concept of a moving bedload allows a revision of some of the older statements that appear in the speleological literature which imply that the sediment on cave passage floors is washed in late in the history of the passage. If much of the mechanical load is transported as bedload, then a sedimentary layer on the floor is a natural state of the passage at all stages in its development. Bedrock floors should be rare and indeed they are. The clastic bed in the passage loses its water, and when that happens the bedload is frozen and is left in the now-abandoned passage. The bed, whether moving or not, serves an important role in insulating the soluble bedrock floor from the solutional attack of the moving water. If the competence of the subterranean stream is only sufficient to move the bedload but not competent to lift it into suspension and expose the floor to solutional or abrasive attack, then the bed serves as a self-perching mechanism which forces the cave stream to flow at its original level and not downcut its channel. Then as baselevel is lowered, a hydrostatic head is generated within the aquifer due to the perched stream. Eventually this head will provide the driving force for cutting a new solution channel on a lower level. The original channel may then be abandoned or carry only a fraction of its former flow. This provides an additional mechanism for the formation of two distinct passages separated vertically from each other rather than one deep canyon. According to this argument canyon passages can form only when the flow which they contain is sufficient to allow some bed scour either mechanically or by solution or both. The way in which the self-perching mechanism of sediment-carrying low-gradient cave streams fits into the generally accepted relationship between cave levels and river terraces is not known but it does provide a mechanism by which, in some cases, cave levels could form without pauses in the lowering of regional base level.

The motion of clastic sediment through the channels of the drainage net immediately implies the possibility of mechanical scour (erosion) as a factor in the enlargement of the cave passage. Such mechanisms have been discussed in the European literature and indeed it is likely that mechanical erosion may play an important role in the high gradient Alpine caves. Whether this is true in low gradient temperate karst is open to question. The load moves very slowly, much of it moves as bedload. The bed may be armored with large diameter material which moves only during exceptional floods and which at other times protects the bed. Cave passages retain their rounded, "solutional" shape although abrasional attack would be principally active against the floor. Mechanical erosion must play some role, particularly in canyon passages, but there seems to be no good reason for assigning new importance to this process.

Since sediment transport is very intimately related to flow velocities and discharge, it is clear than fluctuations in flows due to changes and derangements of recharge areas and to fluctuations in rainfall over long time periods must be important and that a record of these events should be preserved in the sediments. It has been suggested that the massive sediments of many northern caves originated from periglacial climates with massive spring runoffs from snows and glaciers. In this paper we have limited ourselves to a consideration of the generalized mechanism of sediment transport and its importance in cavern development. The use of these concepts as tools for the interpretation of individual karst drainage basins will be an important future topic of research but is not treated here.

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Cave Research Foundation
and
Materials Research Laboratory and
Department of Geochemistry and Mineralogy,
Pennsylvania State University,
University Park, Pennsylvania 16802

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An Electrolytic Field Device for the Titration of CO_2 in Air

By F. Delecour*, F. Weissen*, and C. Ek**

ABSTRACT

A technique for determining the CO_2 content of cave air is described. The titration set, used by agronomists and soil scientists for the analysis of soil atmosphere, has proved to be practical and easily manipulated underground. The procedure is based upon an electrolytic titration of CO_2 absorbed by a 0.1 N NaCl solution. The apparatus is enclosed in a portable wooden case.

Some prelininary results are reported. Attention is given to the gradients of the amount of CO_2 in the atmosphere of some caves: along a shaft, along a scree fan extending from a fractured zone, and along a subterranean stream. The observed range of CO_2 content of cave air ranges from that in the free atmosphere to nineteen times more in a fissure. The most frequent values are two to eight times in spacious caves, compared to free air.

INTRODUCTION

An interesting method for the measurement of CO_2 in soil air has been described by H. Koepf in 1952. It has been used by (among others) F. Hilger (1963), and at the Centre d'Etude des Sols Forestièrs de la Haute Belgique, Gembloux, Belgium. This method is also suitable for atmospheric CO_2 determinations and has been shown by F. Delecour (1965) to be accurate and reproducible.

The determination is based upon the electrolytic titration of the carbonic-acid gas of a known volume of air, absorbed in a 0.1N NaCl solution, that has been colored with phenolphthalien. The electrolysis of this solution produces NaOH, which neutralizes the absorbed carbon dioxide. The current is kept constant during the whole titration. One measures the time necessary to produce enough NaOH to neutralize the absorbed carbon dioxide, which is indicated by the phenolphthalien turning red. The carbon dioxide content of the sample can be computed from the current (mA), the time (seconds) and the air-sample volume (ml).

The apparatus, packed in its case, is shown in figure 1. It weighs 15 kg. It is easily carried, so that measurements can be made in the field (or in a cave).

APPARATUS

Referring to figure 2, the components of the apparatus shown in figure 1 are: A - voltammeter, B - battery compartment, C - rheostat, D - NaCl solution storage flask, E - compartment for titration cells, and F - cover

*Centre d'Etude des Sols forestiers de la Haute Belgique, I.R.S.I.A., Faculte agronomique, Gembloux, Belgium

**Laboratoire de Geologie et Geographie physique, Universite, Liege, Belgium

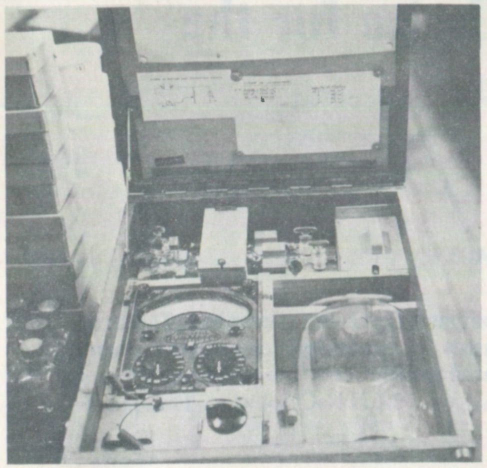

Figure 1.
The titration set in its wooden case.

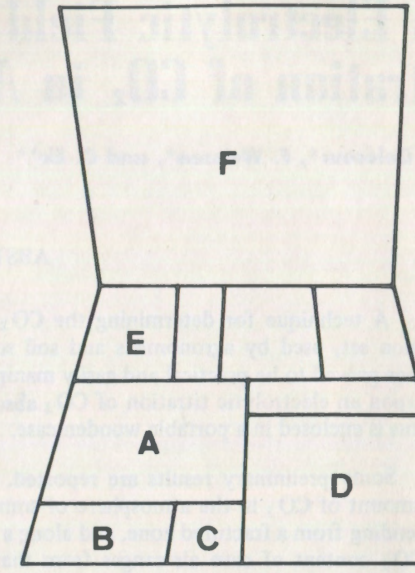

Figure 2.
Components of titration set (see text).

with a white screen for comparing colors. During the electrolysis, the apparatus is arranged as shown in figure 3.

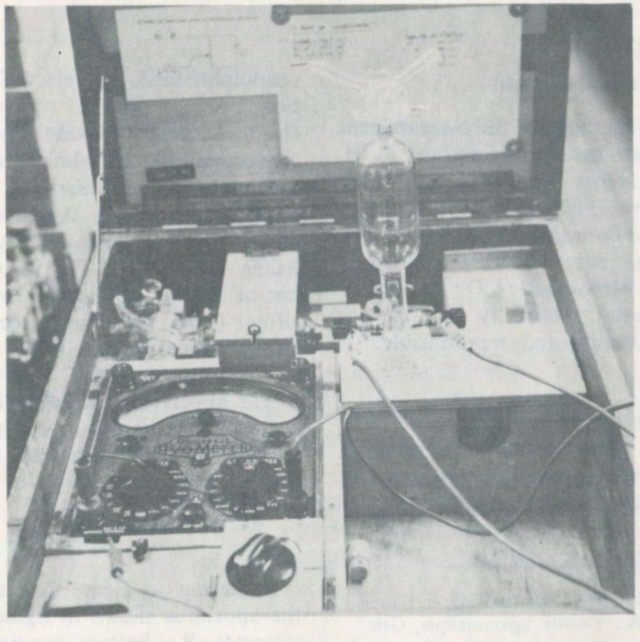

Figure 3.
Arrangement of the apparatus during electrolysis.

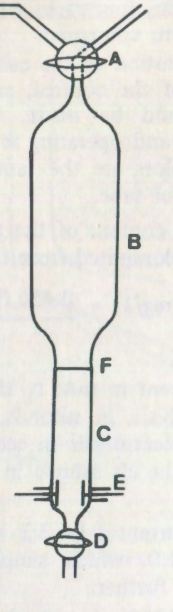

Figure 4.
The titration cell (see text).

Figure 4 is a schematic of the titration cell. Its components are A - three-way stopcock, B - cylindrical bulb (about 3.5 x 9.5 cm), C - lower tube (about 1.5 x 6.5 cm), D - lower stopcock, E - platinum electrodes (1 cm^2, 1 cm apart, and at least 3 cm below F), F - guage line. The total volume of parts B and C is about 100 ml. This must be accurately determined. The volume of C below the guage-line is 10 ml.

The NaCl storage flask is a 3 to 4 liter glass or polyethylene bottle with a stopcock at the bottom. During an analysis, the upper opening is fitted with a soda-lime absorption tube.

The transistor-stabilized current source (0-10 mA) is shown schematically in figure 5. The current strength is regulated by an adjustable 10-k ohm rheostat. Temperature effects are cancelled by the Zener-diode ZL 3.9. Power is supplied by drycells (12-14 volts).

Several reagents are required. These are a) a 0.5% solution of phenolphthalein in 0.1 N NaOH; b) 0.1 N NaCl, adjusted to a pH between 0.9 and 9.1 by means of solution a, for which about 3±.15ml per liter of solution are required. The NaCl solution then exhibits a definite pink color; c) 2.5% KCN-AgCN, prepared by dissolving 2.5 g KCN in about 70 ml of distilled water, dissolving 2.5 g AgCN in this solution, and bringing the volume to 100 ml ; d) 2% KCN, prepared by dissolving 2 g KCN to a volume of 100 ml.

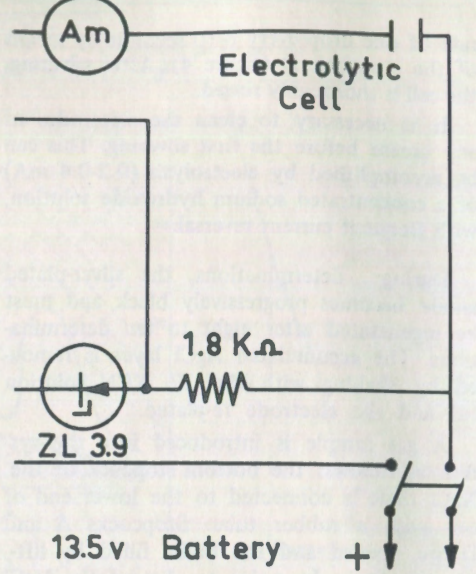

Figure 5.
Stabilized current source.

Procedure

Before use, the anode of the electrolytic cell must be silvered. This silver layer reacts with the chlorine released at the anode, removing it from the reaction system. Silvering of the anode is accomplished using the KCN-AgCN solution (a above), with a current strength of 10mA, for one hour. This may be done with the current-source already described, connected in reverse to the electrodes. About 50 ml of the silver solution is pipetted (caution!) into the titration cell and the circuit is switched on. The solution is allowed to drip from the cell at the rate of one drop every 8-10 seconds by means of the stopcock D (figure 4). After silvering, the cell is thoroughly rinsed.

It is necessary to clean the electrodes of any grease before the first silvering. This can be accomplished by electrolysis (0.2-0.4 mA) of a concentrated sodium hydroxide solution, with frequent current reversal.

During determinations, the silver-plated anode becomes progressively black and must be regenerated after eight to ten determinations. The accumulated AgCl layer is removed by shaking with the 2% KCN solution (d) and the electrode re-plated.

A gas sample is introduced into the system as follows: the bottom stopcock of the NaCl flask is connected to the lower end of cell with a rubber tube. Stopcocks A and D are opened and the bulbs filled by lifting the flask. In order to discard the NaCl solution which has been in contact with air, at least 10 ml of the solution is allowed to flow out through A. Both A and D are then closed, and the cell may be taken to the place where the sample of air is to be taken.

Sampling is accomplished by opening A and D, allowing the NaCl solution to run out until it comes to the level of the guage line F. The stopcocks are then closed. The air-sample volume is that of bulb B.

By vigorously shaking the cell, for four or five minutes, the CO_2 in the sample is absorbed in the NaCl solution, which becomes acidic and hence colorless. The cell is then connected to the current source (figure 5) for electrolysis. A current of 2.2 mA is advantageous for the subsequent calculations. The duration of the electrolysis is recorded to the second (for example, with 2.2 mA, the duration of electrolysis is of the order of 50-55 seconds for a sample of free air). During the electrolysis, NaOH is formed at the cathode and the solution becomes red again. The cell is then disconnected and shaken, as above, for four or five minutes. The color of the solution is then compared with that of a control tube. If the solution in the cell is less colored than that in the control, it should be electrolysed further, until the colors are the same. After every new electrolysis, the cell is shaken, for 30 to 45 seconds, before the colors are compared.

If the solution in the cell is more colored than that of the control, an acidifying electrolysis should be made, by reversing the connections and operating as described earlier, until the colors are the same in the cell and in the control tube.

The CO_2 content of the sample, expressed as mg/1, is computed from the formula where

$$CO_2\ (mg/l) = \frac{0.456\,(i)\,(t_1 - t_2)}{V}$$

i is the current in mA; t_1 the duration of direct electrolysis, in seconds, t_2, the duration of reverse electrolysis in seconds; and V the volume of the air sample in ml.

If the current i is 2.2 mA, (0.456 i) is very near 1.0, which simplifies the calculation even further.

Koepf (1952) found that the range of maximum precision lies between 0 and 5 mg CO_2/1. Beyond 5 mg/1, the results are somewhat low (about 5% for CO_2 content of 7 mg/1).

Results

Some preliminary measurements, carried out during the spring and summer of 1966, displayed encouraging results. All were recorded in Belgium, in limestone caves, at altitudes ranging from 100 to 250 meters above sea level. Additional observations will be reported in the Annales de Speleologie.

In figures 6 and 7 are shown the results of measurements made along steep slopes in Comblain-au-Pont Cave and Rochefort Cave. It can be seen that the CO_2 content of the air may increase downward or upward. The top of the Comblain shaft is open to the outer atmosphere while the top of the "Hell" chamber in Rochefort Cave is connected by fissures with the soil of a wooded doline.

Two factors may induce these CO_2 gradients. In the Rochefort "Hell Chamber", diffusion from a fissure is probably the main factor. At Comblain, the CO_2 gradient can

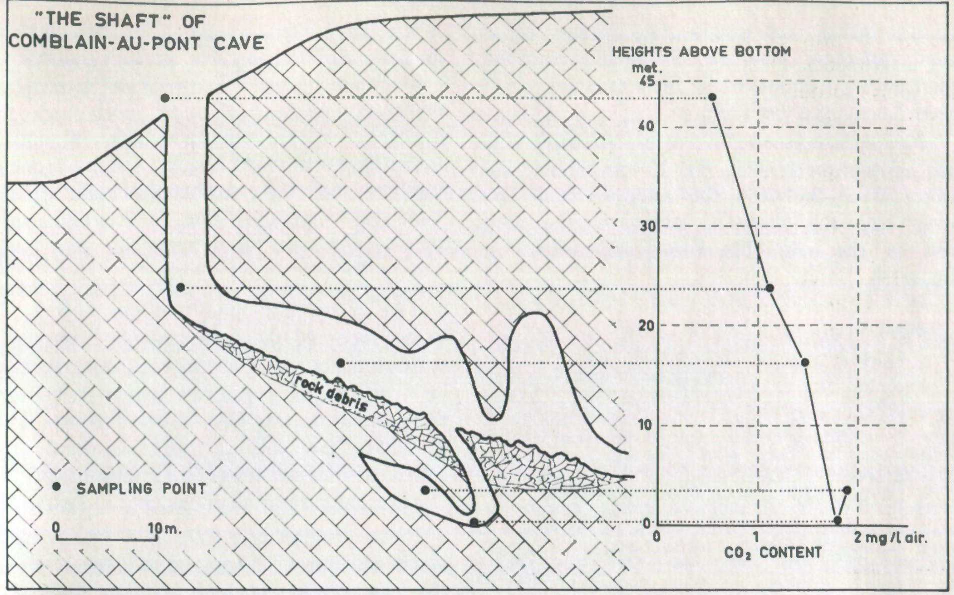

Figure 6.

Cross section from Comblain-au-Pont Cave showing sampling locations and CO_2 measurements.

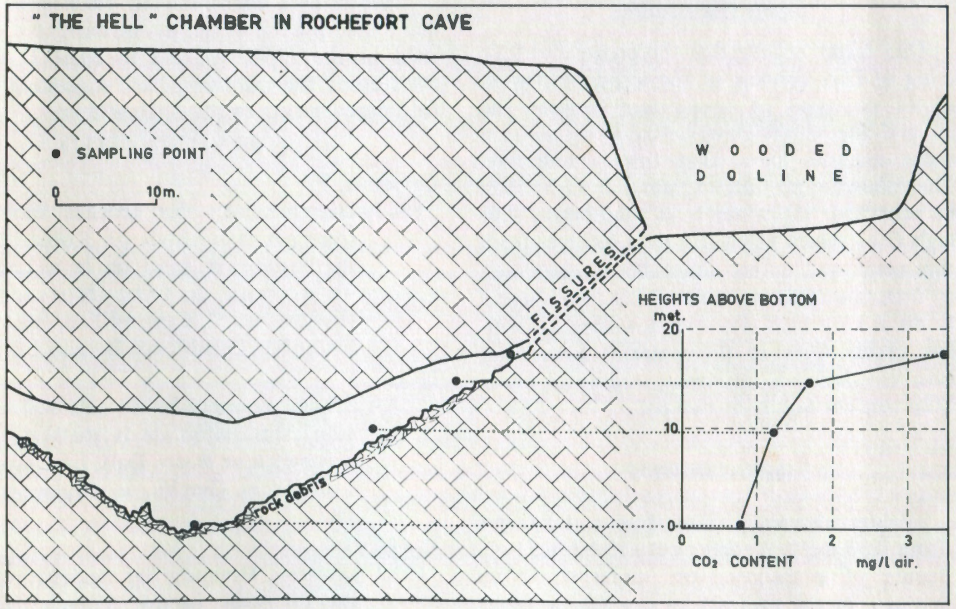

Figure 7.

Cross section from Rochefort Cave showing sampling locations and CO_2 measurements.

be due to an accumulation by gravity. However diffusion may be responsible for the higher CO_2 content in the little chamber situated 3.5 m over the bottom.

Five measurements were carried out along an underground river, at 20 cm above the water level, between the exsurgence of the river and the terminal sump, at the very end of the cave. The results are shown in figure 8.

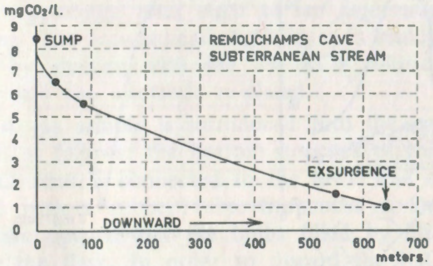

Figure 8.

Subterranean stream in Remouchamps Cave. CO_2 content of air 20 cm above water level. The distance is measured along the stream.

One might assume that the progressive lowering of CO_2 content of the air, downstream, shows variations of equilibrium between the air and the water. However calculations from water analyses show that this is not true. The following possible explanation might then be proposed: the water is in equilibrium with high CO_2 values when the river emerges from the sump and enters accessable passage and, at first, the diffusion rate of CO_2 from water to air is very high. Further downstream, both the excess CO_2 in the water is lower and cave ventilation is improved, leading to lower CO_2 concentrations in the air.

Conclusions

Values observed (75 measurements) ranged from 0.55 mg/1, for free air, to 10.62 mg/1 in a fissure of a shallow cave, under wood and pasture. In the majority of cave atmospheres, values were between 1 and 3 mg/1; they can be a little lower in some passages of big caves, or somewhat higher in small cavities.

The high CO_2 content in fissures (frequently 10, or even 19, times higher in narrow fissures than in the free atmosphere) seems to indicate that air with high CO_2 content flows through these fissures, and that its diffusion in caves is slow. A faint, increasing gradient of CO_2 downslopes was observed twice. It is possibly due to accumulation by density, but the values were not very high at the bottom (less than 3 mg/1).

One might conclude that most of the features reported here were either known or suspected, but the scarcity of measurements in cave air has prevented general conclusions from being drawn. One hopes that the device presented here will contribute to the multiplication of measurements in all seasons, in other climates, at various altitudes, and so forth.

Since atmospheric CO_2 is a *prime mover* in dissolution of limestone, we firmly believe that measurements such as described above will prove to be valuable for furthering our general understanding of karst processes.

Acknowledgements

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