

Base Level, Lithologic and Climatic Controls of Karst Groundwater Zones in South-Central Indiana

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Abstract

Nearly all groundwater movement with the carbonate bedrock of south-central Indiana has been within a karst groundwater zone that occupies the lower part of the vadose groundwater zone above the zone of permanent saturation (phreatic zone) or local base level and that is characterized by a highly fluctuating water table.

Variations of precipitation within climatic cycles and the local relief above base level control the vertical range of the water table and subsequent solution within the karst groundwater zone during that time period corresponding to each subaerial erosion level. Variations in carbonate solubility, thickness of beds, and intensity of jointing determine the texture and shape of the integrated subterranean conduits. The volume of water transmitted through the karst groundwater zone and its acidity determine the size.

Limestone is relatively soluble compared with dolomite and clastic sediments, but it is less permeable than some dolomites or sandstones. Initial permeability in limestone is along joints and bedding planes and varies greatly from bed to bed. Shales and silty carbonate units may form perched water bodies that may be breached by joints or erosion. Sandstone, dolomite, and intensely jointed limestone may form perched water bodies or aquifers within relatively impermeable limestone strata and release the water to less permeable, subjacent limestone.

Introduction

Solutionally-enlarged openings and caverns in dynamic karst groundwater zones carry the greatest amount of groundwater in the non-glaciated portion of south-central Indiana (Fig. 1). Base level, lithology and climate are the three major controls of karst groundwater zones in carbonate rocks of Mississippian age that underlie the Mitchell Plain and Crawford Upland physiographic units of south-central Indiana.

Karst Groundwater Zone

The karst groundwater zone is characterized by a highly fluctuating water table within the confines of any integrated openings, such as joints and bedding planes, in carbonate bedrock (13) (Fig. 2). It is above the zone of permanent saturation (phreatic groundwater zone) and therefore includes the lower part of the zone of aeration (vadose groundwater zone). This zone is somewhat the same as the fluctuating water table zone described by numerous hydrologists, such as Finch (4) and Meinzer (8), and defined by Swinnerton (14) as the zone within which limestone caverns are developed.

Water table fluctuations within karst terrains have been ignored by most theorists concerned with formulating a universal theory of origin of limestone caverns, even though the most intense and frequent water table fluctuations documented have been within karst or cavern areas. A few modern hydrologists argue that a true water table does not exist within the widely spaced openings in carbonate rocks. The fact that the relative grain size between the interstices, usually open joints

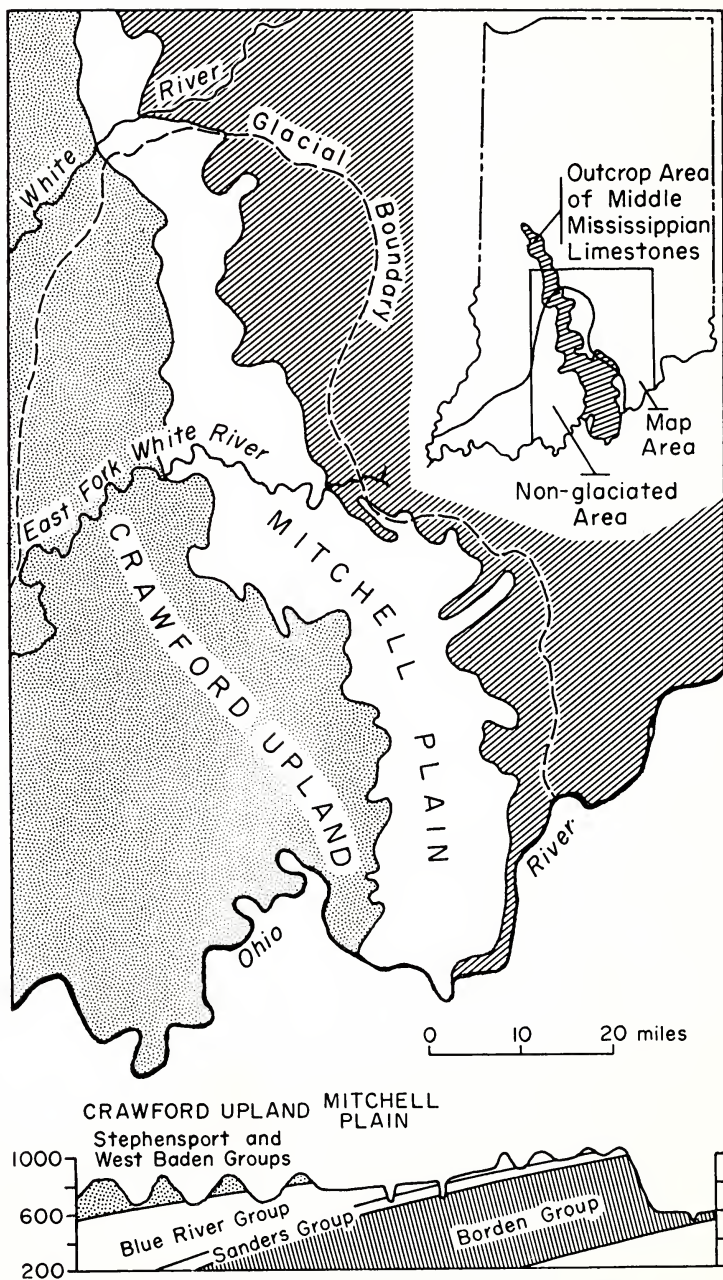


FIGURE 1. Generalized map and cross section of south-central Indiana showing the location of the Mitchell Plain and Crawford Upland and their relationship to rocks of Mississippian age.

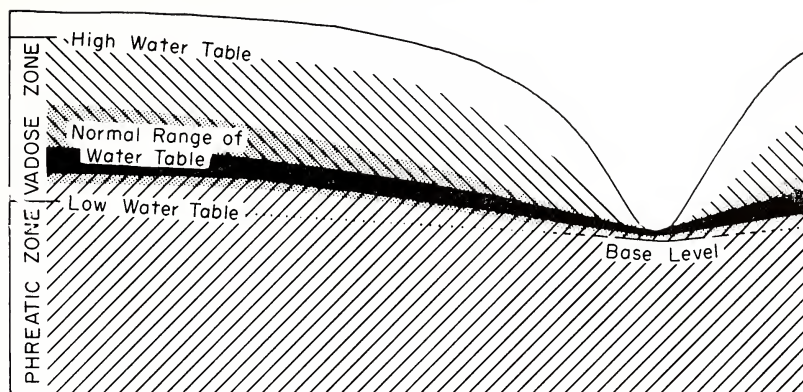


FIGURE 2. Idealized diagram showing the range of fluctuations of a normal water table in a homogeneously permeable medium.

and bedding planes, is measured in feet rather than in microns does not alter the fact that groundwater level fluctuates within the integrated openings. The fluctuations are commonly measurable in feet and tens of feet in Indiana, but rises of the water table up to a few hundred feet are known elsewhere.

A fluctuating water table exists in all permeable strata. The fluctuations of the water table are caused by the difference in frequency and amounts of precipitation, infiltration rates, transmission rates, storage capacity of the bedrock, and discharge rates from the permeable strata or bedrock. Precipitation is variable seasonally, annually, or within climatic cycles and geologic episodes, causing corresponding fluctuations of the water table. The rate of infiltration is dependant upon the presence of exposed surface openings and is affected greatly by the type and density of vegetation. There may be no surface run-off from a forested area while from a denuded area it may be total. The transmission of karst groundwater is dependent upon the size, shape and sinuosity of integrated openings between an intake area and discharge point. The larger the openings, the greater the potential groundwater movement. The amount of groundwater movement and discharge is controlled by size and abundance of integrated openings, and by storage capacity within the fluctuating zone. The rate of groundwater discharge to surface outlets is controlled by the size of the integrated openings and the hydrostatic head that may be developed within the subterranean tributaries to the outlets.

The greatest volume of groundwater movement within carbonate bedrock of a karst terrain occurs within the zone that is most affected by infiltration of acidic meteoric waters and discharge of the accumulated karst groundwaters. The shallow zone just above the zone of permanent saturation within the normal range of water table fluctuation that is affected by even slight infiltration transmits the greatest volume of

fresh karst groundwater (Fig. 2). Therefore, it is the zone of the greatest amount of solution. The amount of karst groundwater transmitted above or below the normal range of the water table decreases proportionally to distance from the normal zone. A decrease in infiltration results in a decrease in the amount of acidic karst groundwater, therefore, a significant decrease in solutional enlargement and a lowering of the water table. An increase of precipitation causing a rise of the water table above the zone of normal fluctuations would increase the hydrostatic head, rate of flow and transmitted acidic karst groundwater in pre-existing openings developed within the underlying zone of normal fluctuations. The normal range of water table fluctuations should decrease to shallower limits during a climatic or geologic episode as openings within the karst groundwater zone enlarged by solution to cavernous routes of proportions capable of containing normal amounts of infiltration.

Heavy precipitation or snow melt may flood integrated openings causing a rise in the water table, an increase in hydrostatic head, and a high velocity within the integrated openings with a corresponding increase in solution rates. The greater the velocity of a solvent, the greater the rate of solution and the greater the amount of solvent, the greater the amount of solution. Thus a large volume of solvent becomes partly saturated much more rapidly than a small volume becomes completely saturated (6). Swinnerton (14) stated that, "A solution of low concentration dissolves more rapidly than a highly concentrated solution. Four volumes of water become one-fourth saturated with CaCO_3 or any similar solute under ordinary conditions in far less time than one volume becomes completely saturated."

Recent papers indicate that groundwater in carbonate rocks which is nearly saturated with calcium carbonate is incapable of a significant amount of solution unless it changes temperature or velocity, or is mixed with other saturated groundwater of a different temperature (1, 15). They have also based nearly all of their cave stream analysis on samples taken during normal flow conditions and none during flood stages, yet the latter condition fills the cavern and produces ceiling solution features that are cited as evidence for development below the water table. Analysis of groundwater taken at spring in Indiana that are discharging from caverns indicate that from about 150 to 300 ppm of carbonate and sulphate ions are removed by solution during normal to low flow stages (9). The concentration of the solution decreases during flood stages, but the great increase of volume of groundwater transmitted results in a net increase in solution of the carbonate bedrock. Nearly saturated groundwater has previously accomplished the significant amount of solution of which it was capable.

Most theorists have noted that many caverns contain evidence of complete water filling, at least in early stages of cavern development, such as the presence of solution features on the upper walls and the ceiling of cavern passages (2). Consequently, they have based their theories on the presumption that the caves were formed at a time when the water table was stable and permanently above the zone of solution

and cavern formation (15). Solution features on the roofs of cavern passages are not proof of permanent saturation nor stable water table conditions. They prove only that the passages were filled when these features were developed. Stable water table conditions are probably non-existent. Climates are not stable, and if the water table were to stabilize, it would indicate that there is a lack of imbalance between intake and discharge within the carbonate bedrock which would suggest a lack of significant groundwater movement. Thus, solution features on cavern ceilings were most likely developed when the passages were water filled with moving groundwater during a temporary high position of the water table.

Base Level Control

Base level has been cited classically as a control of surface geomorphic erosion cycles, as well as a control of levels at which caverns may develop. Ultimate base level is commonly considered as a landward extension of sea level, while regional or local base level is defined in terms of strata or materials resistant to downcutting. Base level in regard to cavern development has been designated as the base level of solution, generally referring to non-soluble rocks below the carbonate host.

Base level is here defined at the top of the zone of permanent saturation (phreatic groundwater zone) or the base of the karst groundwater zone whenever the zone of permanent saturation was relatively stable during any definable episode of cavern development (Fig. 2). Such episodes are usually contemporary with surficial geomorphic episodes. Thus base level is not a flat planar surface, but rather is a gently sloping surface graded in a downstream direction. Surface erosion or downcutting controls the position of the somewhat permanent water table. Subterranean tributaries do not develop significantly below this permanent water table, or base level of significant groundwater movement, or base level of significant solution. Caverns that develop at a designated base level may exhibit irregularities caused by differential solution of various carbonate or non-carbonate lithologies, but generally they reflect the position of the karst groundwater zone that existed when they formed.

Perched water bodies caused by lithologic differences are not regarded as base levels. However, a karst groundwater zone may become perched following surface rejuvenation if the surface stream cuts through an impervious layer or aquitard.

Lithologic Control

A few theoretical papers have considered groundwater flow within a carbonate bedrock as transmission of phreatic groundwater within a homogeneously permeable medium hundreds of feet thick (2, 3). Stratified rocks, including the carbonates, are seldom of uniform lithology, thickness, porosity or permeability for more than a few feet vertically and a few hundred feet laterally. Thick sequences of carbonate strata are usually composed of various thick, medium, and thin-bedded types of

limestone, shaly and sandy limestones, calcareous shales and sandstones and dolomites, and perhaps including non-calcareous (non-soluble) strata, which are all of different solubility rates. The intensity of jointing within each type of lithology also varies greatly from bed to bed and according to the location in relation to local structural features. Thus, any general theory which assumes a homogeneously permeable carbonate bedrock for groundwater movement or cavern development, for other than one bed in a very small area, is ignoring the most important geologic problem concerned.

Most limestones are relatively impermeable rock types, although most are very soluble and some have a high porosity. Most limestones owe their high permeability to integrated openings along bedding planes and joints and subsequent solutional enlargement of these openings. Coarse-grained and porous limestones appear to be more readily soluble than fine-grained or compact limestones. The more dense, compact, thin limestone beds of the lower part of the Sanders Group and the lithographic limestone beds of the upper part of the Blue River Group and some of the limestone units of the West Baden and Stephensport Groups (Table 1) commonly project as resistant ledges into the larger cave passages, or have a comparatively narrow slot dissolved through them if they are thick-bedded. Conversely, large cavern passages formed by solution are more common in thick-bedded, coarse-grained or porous limestones, as in the Salem Limestone Member. Most coarse-grained and porous limestones, even if thin-bedded, have been more extensively dissolved and form recesses in the cave passages formed in strata of the upper part of the Blue River Group and the lower part of the Sanders Group. The compact, denser limestones offer less surface exposure and allow less penetration of solvents than more porous and coarse-grained limestones; therefore, the latter rocks are more easily removed by solution. Rocks low in carbonate content are generally less soluble than those with a higher carbonate content. Purity of the limestones in respect to high calcium carbonate content is of questionable importance. Calcite veins, pure calcium carbonate deposited as crevice fillings, commonly protrude as resistant ledges and dikes in cave passages. Caverns within the St. Louis Limestone Formation, which is generally a heterogeneous unit of argillaceous limestones, are generally unknown, except for small solution tubes. Care should be taken to distinguish between large solutionally-formed cavern passages and large passages which have resulted from collapse enlargement of a small solution passage and solutional removal of the debris at stream level.

Dolomites are less readily soluble than many types of limestone, but they are an important zone of groundwater movement within the Ste. Genevieve Limestone because of their high inherent permeability. Significant amounts of solution occurs in some dolomite beds simply because the dolomite is the groundwater transporting medium which comes into contact with the acidic groundwater. Cavern passages developed in association with dolomite beds are known in several Indiana caverns, notably in Wyandotte Cave, where solution has occurred within or

TABLE 1. *Sequence of important carbonate stratigraphic units of Mississippian age in south-central Indiana.*

Group	Formation
Stephensport	Glen Dean Limestone
	Hardinsburg Formation
	Golconda Limestone
	Big Clifty Formation
	Beech Creek Limestone
West Baden	Elwren Formation
	Reelsville Limestone
	Sample Formation
	Beaver Bend Limestone
	Bethel Formation
Blue River	Paoli Limestone
	Ste. Genevieve Limestone
	St. Louis Limestone
Sanders	Salem Limestone
	Harrodsburg Limestone

through dolomite beds as well as the underlying limestone beds owing to discharge of water from the dolomite beds into the underlying limestone (12). Very permeable porous and vuggy dolomites may project as ledges because the groundwaters pass through them easily, while some very fine-grained, loosely-cemented dolomites are easily removed, perhaps partly by mass wasting and mechanical erosion. Medium beds of dolomite in several Indiana caverns have been widened within cave passages more easily than adjacent limestone beds. Some thick beds of dolomite, in now abandoned passages, have apparently caused passage enlargement by exfoliation or disintegration by weathering.

Calcareous shales that are particularly susceptible to solution occur stratigraphically within the upper part of the Blue River Group. These shales are very soluble and passages within several caves have developed partly by solutional removal of the shale, perhaps aided by some mechanical erosion. Shales, including calcareous and non-calcareous varieties, are generally compact and appear to lack significant open joints; thus, they are commonly barriers to vertically moving groundwater. Perched water bodies occur on the top of the shale within its unbreached limits. However, once breached the perched water is discharged into the underlying carbonate strata. If the relative relief between the perched water body on the shale and the normal water level of the karst groundwater in the underlying carbonate strata is sufficient and if the descending water is transmitted through a vertical joint or set of joints that penetrates several beds, a vertical shaft or pit may result from water running down the walls. Progressive upstream migration of the breached edge

of the shale may result in a laterally elongated shaft, or successive breaching at different points may cause successive abandonment of the former horizontal cave passage and each preceding shaft.

Vertical solution shafts or pits are formed below any lithology such as shale, sandstone, limestone, or dolomite which is relatively resistant to solution in comparison to carbonate strata below, and which is overlain with relatively permeable strata or material. The overlying permeable unit serves as a perched water body reservoir for groundwater to be discharged through an opening in the resistant bed. An opening, such as a joint, through the resistant bed allows somewhat regulated discharge into the underlying carbonate rocks. The character of the reservoir rock and the opening in the resistant bed determines the rate of groundwater discharge into the underlying strata. Vertical shafts or pits are enlarged primarily by regulated flows or seepage down the walls of the vertical opening rather than by complete flooding and filling which is probably possible only in the initial small openings. Some vertical solution shafts are now fed directly by diverted surface drainage and they may have originated by this means.

Thin sandstone beds, particularly the Popcorn Sandstone Bed at the base of the Paoli Limestone Member, serve as resistant beds to form perched water bodies within overlying limestones, discharging water through restricted outlets into underlying strata, commonly forming vertical solution shafts, or laterally elongated shafts. Perched groundwater within thick sandstone units, such as those within the West Baden and Stephensport Groups, is the prime source for the groundwater which develops caverns in underlying limestones (11). Groundwater movement from a perched water body in a permeable sandstone into a limestone as a means of cavern development is probably the major exception to the concept that most caverns are formed by groundwater solution within a karst groundwater zone.

Groundwater movement, and subsequent cavern enlargement, in south-central Indiana is primarily along the joints and bedding planes in the carbonate strata. Indiana caverns commonly have developed parallel to the dip of the strata following sets of joints that allow the steepest gradient to the nearest or lowest surface outlet. Most Indiana cave passages trend with the regional structure which has average dips of about 25 feet per mile to the southwest. There are local structures, however, which deviate from this pattern, and caverns situated on these structures follow the local dip. Some caverns follow the strike of the structure where the karst groundwater zone within the particular cavernous strata has no outlets in a downdip direction along streams which dissected the cavernous strata or karst groundwater zone.

Each bed, depending upon its lithologic type and thickness, has a unique pattern of joints or joint sets. Generally any particular set of joints does not extend into overlying or underlying strata, but some joints are common to several strata, and some master joints apparently extend through many beds of dissimilar lithologies. Thin-bedded strata are

usually more intensely jointed than thick-bedded units. In general it seems that there is a joint spacing of a particular bed that is related to the unit thickness and type of lithology. Although the joints from bed to bed may not be superposed, they do cross and this allows vertical groundwater movement without necessitating significant flowage along bedding planes.

Multi-level cavern passages that have developed in progressive stages as tributaries related to adjacent surface drainage levels or geomorphic episodes commonly reflect the effect of joint patterns shifting from bed to bed. Passages formed along a master joint which dissects numerous beds, usually have a canyon-like appearance with wide places representing temporary halts in solutional deepening or more soluble strata. In addition, passages in one stratum may form an orientation pattern related to somewhat different intake and discharge areas than passages at a different level within another stratum. Goss Cave, in Washington and Harrison Counties, is an excellent example. The upper levels have developed within one set of joints in the upper part of the Salem limestone owing to a general infiltration of meteoric water, while the lower level within the lower part of the Salem Limestone has developed primarily by surface run-off diverted into sinkholes in addition to water diverted from the upper levels. Similar subterranean diversions in Wyandotte Cave have been described (12).

An important cause of the down dip development of caverns and groundwater movement is the dissection of permeable strata by entrenched surface drainage in places down dip from intake or recharge areas. Gardner (5) based his "static water zone" theory of cavern development on the concept that dissection of permeable strata was responsible for initiating groundwater movement and solution of caverns. Although much of his theory is correct, he places too much importance on structural control of cavern development. Other aspects which he mentions, such as climate, are equally important.

Climatic Control

Numerous authors have either suggested that the misfit streams in modern caverns are not large enough to have dissolved the cavern, and that therefore they were either formed in the past when the passages were water filled below a permanently higher water table (3), or they have been formed by diversion of surface streams to subsurface routes at flood stages (7) or by subterranean stream piracy (16).

Although the streams in most Indiana caverns are now misfit, it is obvious that each level was essentially filled with water at times during its period of solutional enlargement. The thick and extensive cave sediments present in Indiana caverns are further evidence of drastic changes in cave stream regimen that have taken place, particularly where the deposits are within abandoned upper levels which no longer contain a stream. Some of the passages which contain misfit streams now flood with even light precipitation, but most are known to flood only in-

frequently during exceptionally heavy periods of precipitation. Some water filled caverns now exist only because they are flooded owing to alluvial sediments which have dammed their outlets (10).

Three factors, misfit cave streams, abandoned cave levels and extensive cave sediments, are proof that the carbonate bedrock area of south-central Indiana has been subjected to multiple cycles of wetter and drier conditions. Further proof of significant climatic change is available in the interpretation of the Pleistocene history of the immediately adjacent areas. At least three ice sheets of the Kansan, Illinoian and Wisconsinan glaciations were close enough to the area to effect moister conditions than during the interglacial stages. These wetter epochs undoubtedly caused more frequent temporary filling of the solutional openings within the karst groundwater zones existing during each glacial episode, with solutional enlargement taking place at a corresponding rate. Waning of the glacial climatic conditions would result in a decrease of precipitation, a change to drier conditions and the subsequent misfit stream conditions.

Ancient climatic changes would also have caused changes in the amount and type of vegetation in south-central Indiana. A dense woodland cover during a wet climatic episode would have caused a lack of surface run-off with a corresponding increase in groundwater infiltration and an increase in groundwater acidity, whereas a drier climatic episode would cause a decrease in vegetal cover, an increase in surface run-off and perhaps in sedimentation in surface and subsurface drainage routes.

Although some evidence is available to correlate certain cave levels, cave sediments or karst groundwater conditions with specific geologic events, data are generally lacking to devise an exact correlation of all cavern development episodes with known geologic events. Data available at this time suggest that episodes of cavern solution and deposition perhaps are the only evidence of geologic or climatic episodes which are yet to be detected as surface geomorphic features. That is, evidences of past climates and geomorphic events may perhaps be best interpreted from evidence obtained in caverns where the evidence has not been destroyed by subsequent events. Multiple cavern levels in south-central Indiana and the cave sediments deposited within them appear to record a sequence of events ranging in age from late Tertiary to Recent times. Each of the cavern levels was obviously initially formed by solution within a karst groundwater zone and perhaps later filled with cave sediments as each succeeding cave level or karst groundwater zone became re-established at another level, higher or lower than the preceding level, owing to changes in base level and climate.

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