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September 21, 2019

Ladies and Gentlemen of the Ohio Power Siting Board,

Rockwell Springs Trout Club is a member owned, private fishing club that was established in 1900. Rockwell Springs is not for or against wind farms generally as a viable alternative energy source. However, considering the tremendous numbers of sinkholes in our general area, we are very concerned with the Republic Wind Farm construction process and potential contaminants that could enter our rare aquifer.

In 1988 we were very fortunate to have a graduate student from the University of Toledo, Gary Edward Kihn, write a thesis entitled *Hydrogeology of the Bellevue-Castalia Area, North-Central Ohio, with an Emphasis on Seneca Caverns,* a copy of Mr. Kihn's thesis accompanies this letter. Mr. Kihn's thesis provided extensive research and a greater understanding of our water. The aquifer begins north and a little west of Attica, Ohio and runs north to Castalia, Ohio. Water is added to the aquifer by precipitation runoff into an extensive series of sink holes between the area of northern Attica to Castalia. This limestone formation ends in and around Castalia with the water coming out of the ground in many artesian springs around Castalia. One of these artesian springs is the water source for Rockwell Springs.

Rockwell Springs, as well as two other trout clubs and the Ohio Division of Wildlife Castalia State Hatchery in the area, depend entirely on the artesian springs to operate. Without the constant flow of cold water from the springs our Rockwell Springs Trout Club would cease to exist.

We have several concerns to enter into the record. All these concerns involve construction and operation of the turbines in such close proximity to the many sink holes that provide our water. The project area east of County Road 21 and north of E. County Road 24 contains numerous sink holes and projected turbines, as shown in the attached graphic.

Will the construction and operation of the turbines harm or disrupt water entering the aquifer? What odds exist that a cave-in or ground depression could occur disrupting the flow of water toward Castalia? What materials are used that could enter and contaminate the aquifer? And, if a contamination occurred what are the means to clean up what entered the aquifer?

All these issues lack a clear answer due to unique structure of the greater-Castalia area geography. The experts will be unable to assure us that our concerns will never be realized as they might be able to assure others in a normal geographic area were sinkholes and aquifers are not present.

Once again, please strongly consider the fact that the aquifer and water table is the lifeblood of Rockwell Springs Trout Club as well as the other two trout clubs and the Ohio Division of Wildlife Castalia State Hatchery.

We would respectfully ask the Board to deny the current wind farm proposal and ask that if this project is implemented it be done in a less environmentally sensitive location.

Respectfully submitted,

Jake Smith, Club Manager

A THESIS ENTITLED

HYDROGEOLOGY OF THE BELLEVUE-CASTALIA AREA, NORTH-CENTRAL OHIO, WITH AN EMPHASIS ON SENECA CAVERNS

BY

GARY EDWARD KIHN

THE UNIVERSITY OF TOLEDO

DECEMBER 1988

A Thesis

entitled

HYDROGEOLOGY OF THE BELLEVUE-CASTALIA AREA, NORTH-CENTRAL OHIO, WITH AN EMPHASIS ON SENECA CAVERNS

by

Gary Edward Kihn

as partial fulfillment of the requirements of the Master of Science Degree in Geology with emphasis in Hydrogeology

Thesis Committee:

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December 1988

ABSTRACT

HYDROGEOLOGY OF THE BELLEVUE-CASTALIA AREA, NORTH-CENTRAL OHIO, WITH AN EMPHASIS ON SENECA CAVERNS

Gary Edward Kihn

Submitted in partial fulfillment
of the requiremants of the
Master of Science Degree in Geology

The University of Toledo

December 1988

Detailed hydrogeologic studies of karst areas in Ohio are seriously lacking. To remedy this situation, the Bellevue-Castalia area of north-central Ohio was investigated to determine the relationships of karst geomorphology to hydrogeologic regime. An in-depth hydrogeologic study of Seneca Caverns was also undertaken.

In the study area, the surficial material is considered non-waterbearing because of its discontinuous areal extent and high clay content. The principal aquifer of the Bellevue-Castalia area is a 700 foot (213 m) carbonate (Middle and Upper Silurian - Middle Devonian) sequence. This fractured rock aquifer exists under confined, semi-confined, and unconfined conditions in response to thickness variation of the overlying glacial and glacio-lacustrine sediments. Yields from the carbonate aquifer range from 10 to 1,000 gallons per minute (6.3×10^{-4} to 0.06 m^3/s) and are dependent on the presence of secondary porosity, mainly in the form of solution-widened joints and fractures.

Recharge to the carbonate aquifer occurs as concentrated autogenic recharge directly through dolines and sinking streams, and as diffuse autogenic recharge indirectly through the cover of surficial deposits. The bulk of the

discharge takes place through springs and, to a lesser degree, through pumpage by rural residents and evapotranspiration by plants. Springs were located and classified as contact, fracture, sinkhole, or subaqueous type. Chemical analyses of the springs indicate that Miller's Blue Hole, Rainbow Creek Spring, Crane Spring, and Muddy Spring may be part of a flow system separate from Duck Pond Spring, Castalia "Blue Hole", Railroad Spring, Castalia Trout Farm Spring, Rockwell Spring, and Ransom Spring. Ground-water movement in the carbonate aquifer occurs through joints or fractures and, thus, is considered diffuse flow. The only evidence of conduit flow is the abundance of dolines that are formed in the highly soluble Columbus Limestone (Middle Devonian). A potentiometric surface map was constructed by measuring water levels in over 100 cased, bedrock wells in the study area. The configuration of the potentiometric surface was influenced by a buried river valley. A fracture trace analysis using low altitude (scale 1:24,000), black and white aerial photographs and joint analyses from four quarries -- Flat Rock, Bellevue, Parkertown, and Castalia -- revealed a dominant bedrock fracture trend of N45E and a less dominant trend of N35W. Two dye traces were attempted to delineate ground-water basin boundaries and to determine the direction and rate of ground-water flow through the aquifer. Five pounds of Rhodamine WT and twenty pounds of Sodium Fluorescein were injected into a doline and monitored at down-gradient springs. However, none of the dyes were recovered after one year. Utilizing water level data and fracture trace/joint analyses, the regional ground-water flow direction was determined to be primarily northeast and is governed by a buried river valley.

Seneca Caverns are formed along a major fracture in the Columbus and Lucas formations (Middle Devonian) and possibly the underlying strata. This fracture trends N68W and dips 40NE. Seven of the twelve known rooms or levels were surveyed and mapped, and a detailed stratigraphic analysis was performed. The Caverns are a collapse or breakdown-type cavern probably resulting from deep-seated solution of gypsum in the Bass Island Group (Upper Silurian). Water was found most commonly in the seventh level approximately 100 feet (30 m) below the surface. Fluctuations in the level of the Cavern stream ("Old Mist'ry River") are the direct result of rate and duration of precipitation and soil moisture content. Lag times were calculated for five storm events from July, 1986 to May, 1987, and ranged from five hours to five days, fifteen hours and ten minutes. Shorter lag times were recorded when the storm duration was shorter and the soil was moderately saturated, and vice versa. The geochemical parameters exhibited by the Cavern stream were very similar to those of Castalia "Blue Hole". The drought of 1988, however, eliminated the scheduled dye trace to determine the possible connection of these two features.

ACKNOWLEDGMENTS

I would like to extend many thanks to my adviser, Dr. Lon C. Ruedisili, for aid in the design of the project and for the revision of the text, to Mr. Ronald E. Gallagher, for his assistance with the Cavern survey, the datalogger equipment, the dye test experiments, and also for his editorial comments, and to Dr. Craig Bond Hatfield, for thoughtful review of the manuscript. Gratitude is also extended to the University of Toledo's Graduate School and Geology Department for their financial aid in the purchasing of equipment for this project.

Special thanks are given to Mr. Richard Bell, owner of Seneca Caverns, for the use of his cavern and for the unending time and guidance he has given me in all facets of this project, and to Dr. James F. Quinlan, Mammoth Cave, Kentucky, for his general expertise in the field of karst hydrogeology and dye tracing and also for allowing me to attend the National Water Well Association (NWWA) sponsored Karst Workshop in March, 1987 free of charge.

I am also indebted to Robert Zaleha, my faithful and enthusiastic field assistant, to Paula Gibson, for her editorial comments and continual support, to Dr. Stuart Dean, for help with the fracture trace analysis, to Kathy Skurzewski, for her computer assistance, to Floyd Stewart (deceased, Castalia Trout Club), to Ken Miller, Charles Schmidt and Roger Ritzert (Owens-Illinois), to Jeff Smith (Rockwell Springs Trout Club), to Dan Longnecker (Rainbow Creek Trout Club), to Ron Tipton (Koppertz Stone Company), to Bud Stephenson (Sandusky Crushed Stones' Parkertown Quarry), to those at the Erie County Metropark, to Mark Dietzel (City of Bellevue), and to all those too numerous to mention by name that gave me access to their property and provided me with technical information.

Last but certainly foremost in my heart a deep-felt "thank-you" to my parents whose continual love, support, and encouragement made the completion of this thesis a reality.

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INTRODUCTION

Scope and Purpose of Study

Karst, according to most authors, is the rolling topography formed by the solution of limestone or dolomite. It is typified by underground drainage, sinkholes (dolines), springs, and caves and caverns. It must be noted that caves have natural openings to the surface whereas caverns do not. Carbonate aquifers, in relationship with karst regions, are found widespread throughout the world and generally exhibit a high waterbearing potential. In the United States alone, it has been estimated that carbonate aquifers underlie approximately 20 percent of the land surface (McGuiness, 1963). Ground-water supplies from some of these aquifers, such as the Edwards aquifer in Texas and the Biscayne aquifer in Florida, are sufficient to support major population areas. In spite of the importance of these aquifers, the mechanics of ground-water flow in karst terranes is foreign to many hydrogeologists. The flow in these aquifers is considerably different from that encountered in porous media. The velocity of flow is much greater, which in turn enhances the systems' sensitivity to contaminants.

Karst development in Ohio is fairly widespread throughout the western half of the state. This is evidenced primarily by the location of numerous caves and caverns in this region. These caves and caverns were located and roughly surveyed by G.W. White (1926) and are the subject of a more detailed study by Hobbs (in progress). Detailed hydrogeologic studies of these areas, however, are seriously lacking in the literature. To remedy this situation, a portion of north-central Ohio was chosen for such a study (Figure 1). Seneca Caverns, one of the largest caves or caverns in the state, is located within this area and was included in this study.

The primary objective of this study was to determine the relationships of the karst geomorphology to the hydrogeologic regime of the Bellevue-Castalia area. The specific objectives of this study were to: (1) identify and classify the karst features located in the study area, (2) identify the physical and chemical parameters of the carbonate aquifer, and relate them to ground-water flow, (3) identify the direction(s) of ground-water flow, and (4) perform a detailed hydrogeologic study of Seneca Caverns.

Methodology

The initial stage of the investigation entailed a reconnaissance of all karst features including dolines, sinking streams, caves and caverns, and springs found in the Bellevue-Castalia area. These karst landforms were then examined in detail and classified in terms of their origin and physical properties.

The behavior of the carbonate aquifer was examined. Physical and chemical properties of the springs were used to determine aquifer characteristics. This information, combined with knowledge about surficial and bedrock geology, was used to determine if flow within the aquifer was primarily diffuse or conduit-type as described by W.B. White (1969 and 1977).

By measuring the water level in over one hundred bedrock wells in the study area, a potentiometric surface map of the carbonate aquifer was constructed. Information from this map was utilized to determine the direction(s) and gradient of ground-water flow. Through the use of low altitude, black and white, aerial photography, the bedrock fracture pattern was identified. These data were utilized to aid in the determination of the regional trend of ground-water flow. It was hoped that the flow direction(s) and flow-through times within the aquifer could be determined with the addition of fluorescent dyes to the ground-water system.

Specific dyes used included Rhodamine WT and Sodium Fluorescein.

The hydrogeology of Seneca Caverns was studied. A transit survey utilizing a Wild T1A theodolite was conducted to establish the vertical and horizontal components of the various rooms or levels comprising Seneca Caverns. Detailed maps illustrating the spatial arrangement and unique features of each level were constructed. A detailed stratigraphic section was formed by carefully examining the geology of each level and comparing the results with that summarized by Stewart (1955) from central and north-central Ohio. A study was also instituted to determine the relationships between precipitation events and the water level found within Seneca Caverns. An Omnidata International Easy Logger was installed which simultaneously measured precipitation, water temperature, and the level of the Cavern stream. These data were utilized to determine the lag times between storm events and resulting water level fluctuations during the course of a year.

Description of Area

Location

The study area encompasses approximately 240 square miles (622 sq. km) of portions of Seneca, Sandusky, Erie, and Huron Counties, in north-central Ohio (see Figure 1). It is bounded to the north by the shore of Sandusky Bay (Lake Erie), to the east by State Route 4, to the south by county road 46, and to the west by township road 268. This area lies between latitude 41° 27' and 41° 11' north and longitude 82° 45' and 82° 57' west.

Physiography

The Bellevue-Castalia area lies within two regions of the Central Lowlands Physiographic Province (Burgess and Niple, 1967). The northern two-thirds of the study area is located within the Wisconsinan Lake Plains Region and the remaining one-third within the Till Plains Region. The approximate boundary between the two regions is defined by glacial Lake Maumee beach ridges (Figure 2).

The Lake Plains Region is nearly level to slightly undulating with poorly drained soils developed from clayey and loamy glacial till, and silty glacio-lacustrine deposits. Also included in this region are several fairly continuous beach ridges that mark the changing water elevations of the Wisconsinan lakes formed by the impoundment of glacial meltwaters in the Lake Erie Basin.

The Till Plains Region is gently rolling with poorly drained soils formed of high lime glacial till. Unstratified, glacial outwash sediments deposited by the retreating glacier are characteristic of this region.

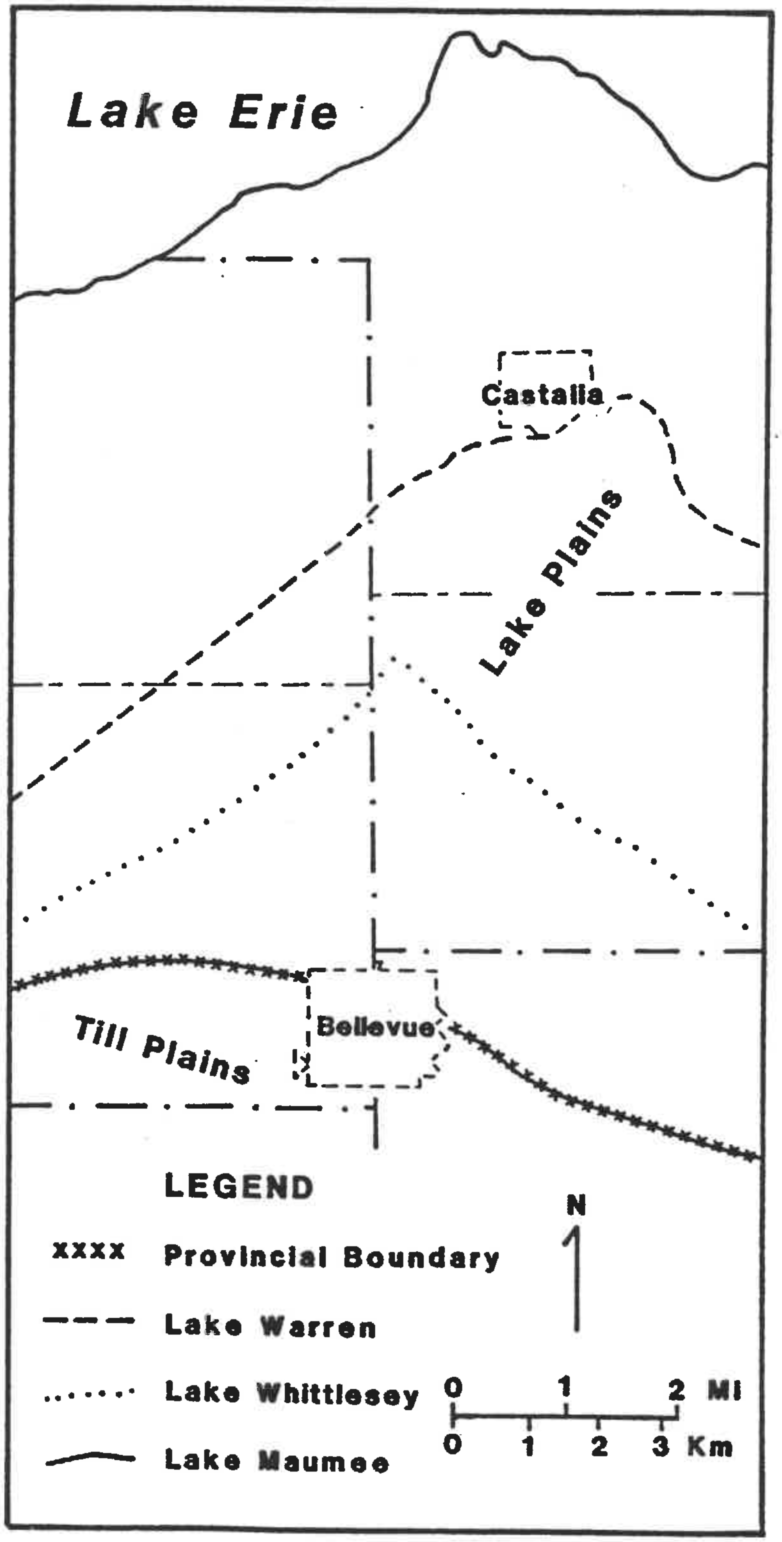


Figure 2. Map showing the Wisconsinan Physiographic Provinces and the major beach ridges (after Forsyth, 1959).

The general surface slope of the study area is northward toward Sandusky Bay at approximately 13 feet per mile (2.46 m/km). The main topographic features in the area are beach ridges and the dissected remains of a bedrock high known as the Columbus cuesta (Figure 4, page 19). The maximum relief of these features is 20 feet (6 m) and 125 feet (38 m), respectively.

Climate

The proximity of Lake Erie and the prevailing southwesterly winds has a moderating effect on the temperatures of the area. For this reason, extremes are seldom recorded. The mean annual temperature is 50.6°F (10.3°C), with the average monthly extremes being 25.7°F (-3.5°C) in January and 73.9°F (23.3°C) in July. The mean annual precipitation and snowfall are 33.90 and 38.40 inches (86.21 and 98.54 cm), respectively. The date of the last killing frost averages May 7, while the first killing frost in the fall occurs on October 12. All climatological data listed above were obtained from the open files of the National Oceanographic and Atmospheric Association (NOAA) station in Sandusky, Ohio (NOAA, 1986).

Land Use

Land in the study area is utilized primarily for agricultural purposes. Approximately 92.4 percent of the land is in farms and 6.7 percent is classified as urban or built-up (Ohio Water Commission, 1966). Cropland comprises 81.3 percent of the farmland, farm woodland 8.7 percent, and pasture 3.4 percent. The balance of the farmland is devoted to other miscellaneous purposes.

Previous Investigations

Numerous studies have been completed regarding the stratigraphy, bedrock geology, and karst features of the Bellevue-Castalia area. A majority of these studies, however, are very general in nature. Hydrogeological studies of this area are generally lacking.

Newberry (1873) first observed a lack of surface drainage near Castalia, while Hubbard (1928) postulated that sinks and conduits in the limestone were formed by solution, thus leading to the lack of surface drainage in the Bellevue area. The most comprehensive study of the area was performed by Ver Steeg and Yunck (1932). This paper included discussions on the topography, glacial history, and bedrock structure of the Bellevue-Castalia region. The artesian springs located in Castalia and sinkhole development in the area were identified. It was determined that the sinkholes were formed by the extensive development of underground drainage which caused the surface rock to slump into cavities below. A dye test was performed to determine the source of the Castalia springs. Two pounds of uranine dye were flushed down a toilet in Bellevue, located approximately nine miles (15 km) to the south. The dye cloud was spotted in a flooded field three miles (5 km) south of Castalia. None of the dye, however, was ever recovered in the Castalia springs. The most recent study concerning karst features in the study area was conducted by Tintera (1980). The major contribution of this study was a list of the sinkholes and their dimensions.

Contaminated ground water in the Bellevue area as a result of improper waste disposal by the towns' residents was first noted by Daugherty (1941). Followup studies by the Ohio Division of Water (1961) and Walker (1962) mapped the extent of ground-water contamination in the region. Sikora (1975) expanded this work further and determined that ground water was no longer contaminated.

Several regional studies were completed which included sections on surface- and ground-water resources of the study area. These studies were performed by the Ohio Department of Natural Resources (ODNR), Division of Water (1966 and 1970) and the Ohio Water Commission (1966). Other reports concerning ground-water resources of the area include Walker

(1962), ODNR, Division of Water (1968), United States Geological Survey (USGS), Department of Interior (1972), Schmidt (1980 and 1982), and Hoover (1982).

A large number of investigations have been completed regarding the stratigraphic characteristics of the Devonian and Silurian strata found in the study area. Mather (1859) was the first to use the name Columbus Limestone, while Newberry (1873) clearly defined the extent of this unit. The name Delaware Limestone was first proposed by Orton (1878). Struble (1952) conducted a petrographic study of these units in northern Ohio. Lane et al. (1909) were the first to name and define the Detroit River Group and subsequently subdivide it into the Lucas and Amherstburg formations in north-central Ohio. A summary of these Devonian units was included in reports by Stauffer (1909 and 1957), Stewart (1955), Dow (1962), Bailey (1968), and Skryzniecki (1972). The earliest work on the Silurian rocks of northwest Ohio was that of Gilbert (1873) and Winchell (1873 and 1874) in which the occurrence of the Lockport Group was mentioned. Lane et al. (1909) described the Raisin River, Put-in-Bay, Tymochtee, and Greenfield dolomites and Carman (1927) assigned these formations to the Bass Islands Group. Additional studies of these Silurian units include Ulteig (1963 and 1964) and Janssens (1968 and 1977). Stout (1941) described the chemical compositions of all carbonate units in western Ohio and gave generalized descriptions of numerous quarry sections. In succeeding years, in-depth studies of quarries in the study area were undertaken to determine the bedrock geology. Janssens (1970) studied the Bellevue, Castalia, Parkertown, and Venice quarries while Sparling (1970) and Forsyth (1971) studied the Castalia quarry. Cunningham (1972) then conducted a petrographic examination of the rock units in the Parkertown quarry.

Several studies have been completed on Seneca Caverns to date. In 1926, G.W. White identified eleven major cave systems in Ohio including Seneca Caverns, known then as Good's Cave. A tape and compass survey was completed along with a general cavern description. According to G.W. White (1926), Seneca Caverns was not formed by faulting. However, no other explanation was given for its formation. Grieves (1953) performed the next study of Seneca Caverns and concluded that the Caverns was formed by faulting. Short (1970) presented a brief cavern description and theorized that Seneca Caverns is a collapse or breakdown-type cavern due to deep-seated solution activity which undermined the area causing the overlying rock to collapse.

GEOLOGY OF THE BELLEVUE-CASTALIA AREA

Stratigraphy

The stratigraphic column for the Bellevue-Castalia area is composed of Middle and Upper Silurian and Middle Devonian limestones and dolomites overlain by Pleistocene glacial deposits (Table 1). The surficial deposits, because of their discontinuous areal extent and high clay content, are generally considered non-waterbearing. The underlying 700 ft (213 m) carbonate section is of prime concern because it represents the regional aquifer for the area. At the time these carbonate units were being deposited, approximately 350 to 425 million years ago, a stable marine shelf environment existed within the mid-continental United States. During Middle and Late Silurian time, a quiet and relatively shallow lagoon-type basin existed in the study area, located east of the Findlay Arch, while an open sea prevailed to the west (ODNR, Division of Water, 1970). The strata deposited during this time included, from bottom to top, the undifferentiated Lockport Group, and the Greenfield, Tymochtee, Put-in-Bay, and Raisin River formations of the Bass Islands Group. These units are primarily dolomites of variable thickness containing appreciable amounts of gypsum and other

Table 1. Stratigraphic column for the Bellevue-Castalia area (after Janssens, 1970 and Sikora, 1975).

System	Series	Group	Formation	Average Thickness (ft)
Quaternary	Pleistocene	Wisconsin	Mankato Cary	0 - 100
		pre-Wisconsin	Drift?	
Devonian	Erian		Delaware	35
	Ulsterian		Columbus	60
		Detroit River	Lucas	35
			Amherstburg	50
Silurian	Cayugan	Bass Islands	Raisin River	50
			Put-in-Bay	50
			Tymochtee	150
			Greenfield	45
	Niagaran	Lockport (undiff.)		225
			Rochester	25

evaporites, along with storm-produced clay and silt lenses that were swept across the barrier arch.

The Lockport Group (Middle Silurian) is usually a whitish gray to light blue-gray, porous dolomite. In some areas, however, the upper 30 feet (9 m) of the formation is very white. The basal portion of the Upper Silurian System is the Greenfield Dolomite. It is light gray to buff in color and is very hard. This formation is quite similar to the Raisin River and Put-in-Bay formations. The Tymochtee Dolomite is a medium grayish brown grading from dark gray to black. It is an excellent horizon marker separating the buff-colored Greenfield and overlying Put-in-Bay and Raisin River dolomites. These formations exhibit a myriad of depositional features including tidal channels, shrinkage cracks, ripple marks and algal stromatolites (Kahle and Floyd, 1971).

The lowermost Devonian units in the study area are known as the Detroit River Group. They rest unconformably on the Upper Silurian strata. The Amherstburg and Lucas formations are differentiated based on their fossil content, although neither has an abundance of fossils (Tintera, 1980). Both of these formations are dolomites of fair purity, with the Amherstburg being rather massively bedded and vuggy due to the secondary solution of gypsum, and the Lucas more thinly bedded (Stout, 1941). They are both of marine origin and were deposited in a tidal environment (Janssens, 1971).

The Columbus Formation rests disconformably on top of the Detroit River Group and varies in lithology from a dolomite upward into a limestone (Stauffer, 1909 and Stout, 1941). Summerson et al. (1957) distinguished these two lithologies calling the lower unit Bellepoint and the upper unit Delhi, whereas Swartz (1907) divided the Columbus into the Bellepoint, Marblehead, and Venice members based on fossil content. The Columbus strata range in color from light gray to brown, are massively bedded with some layers and nodules of chert, and are of very pure composition, averaging 94 percent calcium carbonate. The purity of the strata has lent to the intense development of karst features such as dolines, which dot the surface, and sinking streams. These features will be discussed in greater detail in Chapter 3.

The youngest of the Devonian carbonates in the study area is known as the Delaware Formation. The strata are predominantly limestones with some interbedded shale layers and are light to dark blue in color, finely crystalline, hard and dense, and very fossiliferous. Some layers contain considerable chert usually in the nodular form (Stout, 1941).

Bedrock Geology

The outcrop belts of the carbonate units underlying the Bellevue-Castalia area are illustrated in Figure 3. Exposures are uncommon throughout the region due to a blanket of glacio-lacustrine deposits. Bedrock exposures are generally limited to quarries, caves and caverns, and in the bottoms of certain dolines.

One additional area in which the bedrock may be seen at the surface is along the crest of what is commonly called the Columbus cuesta. A cuesta is a ridge which has a gentle slope on one side and a steep slope on the other. Cuestas are common features in eastern Ohio. They are very uncommon in western Ohio, however, where glacial cover is relatively thick and extensive and the dominantly carbonate rock units offer little contrast in rock resistance. Aside from a few islands in Lake Erie, it is only in the Bellevue-Castalia area that such a cuesta is found.

The Columbus cuesta is formed in the Columbus Formation (Middle Devonian) and is oriented in a northeast-southwest direction between Bellevue and Castalia (Figure 4). The slope is very steep to the northwest while much more gradual to the southeast. The elevation of the cuesta ridge, approximately 785 feet (239 m), was high enough that glacial deposits never buried it. This ridge formed a major promontory projecting into the ice-dammed lakes that occurred in the Lake Erie Basin during the retreat of the Wisconsinan glaciers (Forsyth and Kahle, 1983).

This cuesta was formed as a result of the extensive subaerial erosion that preceded glaciation (Forsyth and Kahle, 1983). Prior to the advent of Pleistocene glaciation, when there was no Lake Erie, there was a major river present in the area known as the Erigan River (Figure 4). This northeasterly-flowing river and its valley subsequently channeled the flow of the advancing glaciers causing the river valley to become enlarged and somewhat deepened by the erosion of these glaciers. It is the differential erosion by this preglacial river and its tributaries that is believed to have created this cuesta.

Water-yielding Characteristics of the Bedrock

As was noted previously, the carbonate section represents the regional aquifer for the study area. At the base of this system is the Rochester Shale (Lower Silurian), which acts as an aquitard separating the waterbearing limestones and dolomites from the non-waterbearing shales, limestones, and dolomites. Rock units deposited prior to the Rochester Shale are considered non-waterbearing due to their great compaction and low ground-water storage potential.

Limestones (calcium carbonate) are generally considered as low waterbearing rocks. Secondary processes which result in the formation of joints and fractures greatly increase the permeability making it possible for waters to circulate in large quantities. These circulating waters, moving through joints and fractures, continuously dissolve and enlarge these openings allowing still larger quantities of water to be transmitted throughout the aquifer.

Dolomites (calcium magnesium carbonate) commonly are more coarsely crystalline, contain more void spaces, and consequently possess a higher primary porosity than do limestones. Although dolomites are subject to secondary porosity through the formation of joints and fractures, the magnesium content makes them more resistant to dissolution than limestones.

Table 2 indicates the locations of the water-yielding zones within the carbonate aquifer of the Bellevue-Castalia area. Proven yields from industrial and municipal wells ranging from 500 to 1,000 gallons per minute (gpm) (0.03 to 0.06 m^3/s) have been developed at depths of less than 300 feet (91 m), while farm and domestic supplies of 10 to 25 gpm (6.3×10^{-4} to 1.6×10^{-3} m^3/s) are usually encountered at less than 125 feet (38 m) (Schmidt, 1980 and 1982). Increased yields, however, are normally accompanied by a decrease in water quality. Water quality measurements which exceed the safe drinking water standards for hardness, dissolved solids, hydrogen sulfide, and sulfate are found in many of the deeper wells which penetrate the Silurian strata and the Amhertsburg and Delaware formations (Middle Devonian). This is due to the presence of gypsum (calcium sulfate) and organic sulfur. Shallow wells developed in these units, however, normally yield water with lower concentrations of the aforementioned constituents.

Norris and Fidler (1971b) determined that structure is the primary factor controlling ground-water yields in Silurian rocks in northwest Ohio. A "high-yield" zone was defined (Figure 5) which resulted from the formation of high

Table 2. Water-yielding characteristics of the carbonate aquifer underlying the study area (after ODNR, Div. of Water, 1970).

System	Formation or Group	Water-yielding Characteristics
Devonian	Delaware Formation	Although entire thickness yields water, primary zones are 10 to 30 feet (3 to 9 m) below top of formation. May contain sulfur.
	Columbus Formation	Although entire thickness yields water, primary zones are 20 to 40 feet (6 and 12 m) below top of formation.
	Detroit River Group	Most wells produce at depth of 40 to 60 feet (12 to 18 m) below top of group. May contain sulfur.
Silurian	Raisin River & Put-in-Bay Fms.	Most wells produce at 20 to 50 feet (6 to 15 m) above the Tymochtee Fm.
	Tymochtee Formation	Wells yield at 30 to 70 feet (9 to 21 m) above the Greenfield Fm., 73 percent of the wells yield supplies.
	Greenfield Formation	The zone 20 to 30 feet (6 to 9 m) above the Lockport Group has producing wells, but less than 31 percent yield supplies.
	Lockport Group	Formation yields minimal amounts of water. Primary zones, however, are 50 to 80 feet (15 to 20 m) below the top of the unit. May be mineralized at depth.

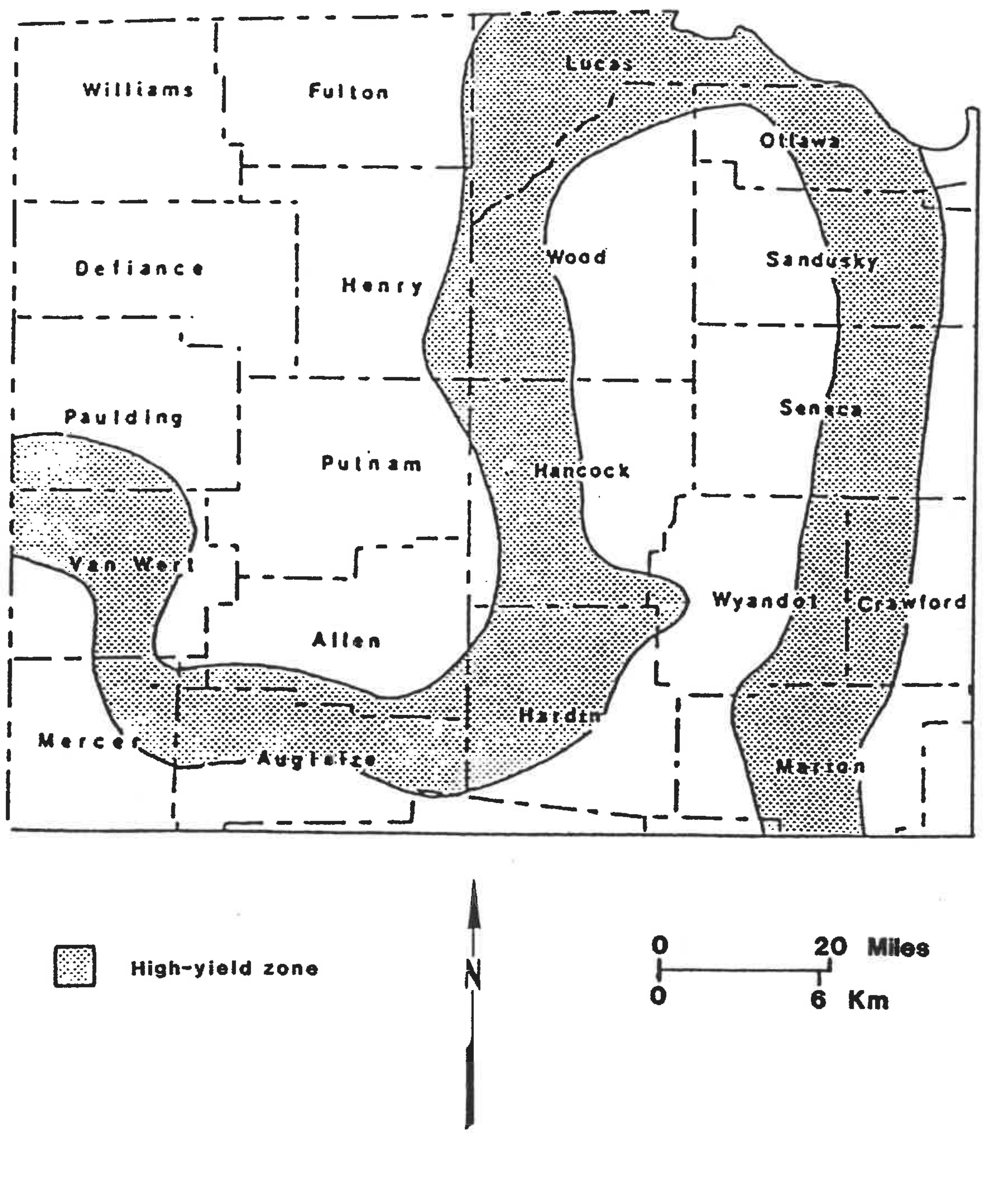


Figure 5. Map of northwest Ohio illustrating the location of a "high-yield" zone (after Norris and Fidler, 1971b).

secondary porosity due to the emergence and subsequent erosion at the crest of the Findlay Arch. Increased solution of this structural high resulted in a region of extensive secondary openings along the flanks of the arch when rocks at the crest were removed by erosion. This is evidenced by caliper-log data which indicate that rocks within the "high-yield" zone contain a greater percentage of openings than do rocks outside this area.

Surficial Geology

During the Pleistocene Epoch, approximately ten thousand to two million years ago, this region was covered by numerous ice sheets. Advancing southward out of Canada, these ice sheets planed off the former landscape and deposited a glacial till upon it. Three of the four major glacial ages in North America have left evidence of their passing in Ohio. These are, from oldest to youngest, the Kansan, Illinoian, and Wisconsinan. All of the glacial deposits visible today, however, were derived from the most recent glacial advance, the Wisconsinan. These surficial deposits are composed of a highly calcareous clay glacial till, glacio-lacustrine clays, and sands and gravels of outwash deposits, the thickness of which ranges from near zero feet in the vicinity of Bellevue to over 100 feet (30 m) in the northwestern portion of the study area (Figure 6). The till, which is a heterogeneous mixture of clay, sand, gravel, and boulders was deposited as a ground moraine during the advance of the glacier, whereas, the glacio-lacustrine clays and outwash sands and gravels were deposited during its retreat. The outwash deposits represent the materials that were washed out of the glacier. The glacio-lacustrine clays were deposited when meltwater lakes were formed by the impoundment of the water to the north by the glacier and to the south by a drainage divide. Deposited around the perimeter of these lakes were beach sands and silts. These deposits, known as beach ridges, were the result of the changing water elevations of these meltwater lakes. As the ice sheet continued to recede northward, these beach ridges, which are representative of former lake shorelines, also continued to move northward as the water level lowered over time. The names associated with the fluctuations in lake level are found in Table 3.

Beach ridges resulting from three major changes in water level are apparent in the Bellevue-Castalia area (see Figure 2). The oldest was formed approximately 14,000 years ago by Lake Maumee. It consisted of three levels (I, II, III). Level I, however, does not occur in the area because the ice was still present here. The level III beach ridges are the topographically highest and most extensive of any of the Maumee lake levels in the study area (Forsyth, 1959). As the ice sheet retreated, the water level receded forming Lake Arkona (Levels I, II, III) at a maximum elevation of

Table 3. Wisconsinan lake levels of the Lake Erie Basin (after Forsyth, 1959).

Substage	Years B.P.	Lake Stage	Elevation (ft)	Outlet
Recent	11,000	Erie (modern)	573	
Valders			573	Niagara River
Two Creeks Interval		Erie (early)	540	
	12,000			
Post-Port Huron		Early Algonquin	605	St. Clair - Detroit River
		Lundy	620	Lake Calumet
		Grassmere	640	Lake Glenwood
Port Huron		Warren III	675	
		Wayne	658	
		Warren II	680	
		Warren I	690	
		Whittlesey	735	Grand River, Michigan
	13,000			
Carey		Arkona III	695	
		Arkona II	700	
		Arkona I	710	
		Maumee III	780	Wabash R., Ind.
	14,000	Maumee II	760	Grand R., Mich.
		Maumee I	800	Wabash R., Ind.

710 feet (216 m). Following this, however, the ice sheet made a major readvancement covering the beach ridges produced by Lake Arkona. This new lake, known as Lake Whittlesey, occupied the Lake Erie Basin 11,400 to 13,000 years ago and attained an elevation of 735 feet (224 m). The beach ridges formed by Lake Whittlesey exhibit the greatest relief and are most prominent of all the glacial lake beaches in northern Ohio (Forsyth, 1959). As the glaciers then began to recede, Lake Warren (Levels I, II, III) was formed. Lake Warren reached its peak elevation of 690 feet (210 m) 1,800 years after that of Lake Whittlesey. Beach ridges of the Warren stage are strikingly sandy with minor amounts of gravel (Forsyth, 1959).

Water-yielding Characteristics of the Surficial Deposits

The surficial deposits of the Bellevue-Castalia area consist of a highly calcareous, clay glacial till, glacio-lacustrine clays, sand and gravel deposits situated adjacent to the western flank of the Columbus cuesta, and relatively thick deposits of fine sand and silt corresponding to Wisconsinan stage beach ridges. The water-yielding characteristics of these deposits are listed in Table 4. The sand and gravel deposits, depending on their vertical and lateral extent, generally yield sufficient quantities of water (approximately 10 gpm) for domestic or farm purposes. The glacial till is not a good source of water unless gravel

Table 4. Water-yielding characteristics of the surficial deposits in the study area (after Walker, 1962).

Material	Water-yielding Characteristics
Stringers of sand and gravel interbedded in less permeable glacial material.	Quantity of ground-water available depends upon the extent of the sand and gravel and the source of recharge; usually capable of supplying domestic water needs.
Till, a heterogeneous mixture of clay, sand and gravel with a pre-dominance of clay.	Generally not a source of ground water although, in places, gravel lenses within the till will yield small domestic supplies.
Clay, silt, and fine sand from glacial lakes.	Not generally a water source due to the fine-ness of material; water contained within the sand cannot be recovered.

lenses can be located. These gravel lenses, however, are often not extensive enough to be used as a domestic water source. The glacio-lacustrine clays, and beach sands and silts, are considered non-waterbearing due to the fineness of the material.

Structural Geology

The Bellevue-Castalia area, as well as most of western Ohio, is characterized by relative structural simplicity. Stratified rocks in the area dip southeastward into the Appalachian Basin at approximately 30 feet per mile (5.67 m/km). This dip angle is due to the northern extension of the Cincinnati Arch, known as the Findlay Arch, which is located just west of the study area (Figure 7). The Cincinnati Arch is a regional feature extending as far south as Alabama. In western Ohio, it branches into the Kankakee Arch which trends northwest into Wisconsin and the Findlay Arch which plunges northeast into Ontario, Canada. This Cincinnati-Findlay-Kankakee Arch system divides three basins -- the Appalachian Basin to the southeast, the Illinois Basin to the southwest, and the Michigan Basin to the north. This arch system was formed by the subsidence of these three basins rather than by tectonic uplift (Stout, 1941).

The genesis of the arch system began with the subsidence of the Appalachian Basin in Cambrian time. By late Silurian, the Michigan and Illinois Basins began to subside to the north and west, respectively. During Devonian time, the Kankakee Arch separated the Michigan and Illinois Basins.

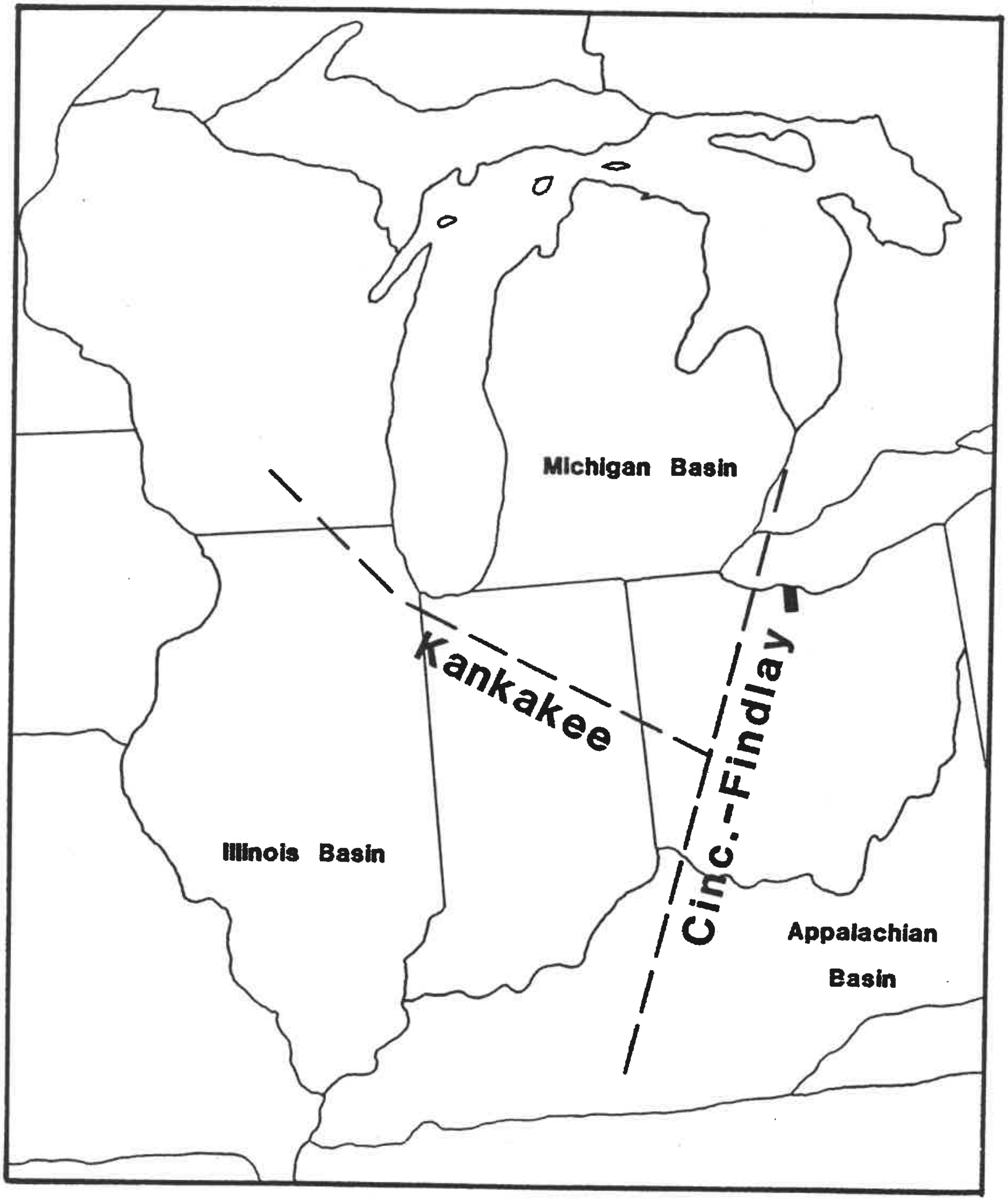


Figure 7. Location of the study area with respect to the Cincinnati-Findlay-Kankakee Arch system (after Riggs, 1960).

KARST DEVELOPMENT

General

Karst is a terrane with distinctive characteristics of relief and drainage arising primarily from a higher degree of rock solubility than found elsewhere (Jennings, 1971). This definition stresses geomorphic form, but emphasizes process in stressing the role of solution in natural waters and emphasizes structural control in stressing that abnormally soluble rocks are involved.

There are four conditions which contribute to the maximum development of karst (Thornbury, 1969). First, there must be present at or near the surface a soluble rock, preferably limestone. Dolomite may suffice although it is not as readily soluble as limestone. Second, and possibly most important, is that the soluble rock should be dense, highly jointed, and preferably thin bedded. It is often concluded that the most important prerequisite for karst development is the presence of a permeable or porous limestone. Permeability, in the sense of mass permeability, is, however, unfavorable. If a rock is highly porous and permeable throughout, rainfall will be absorbed en masse through the whole body of the rock rather than concentrated along restricted lines of movement (Thornbury, 1969). The third condition is that there is marked topographic relief, with the uplands being underlain by well jointed limestone. This favors rapid downflow of ground water through the bedrock. Good water circulation is extremely important in that moving water promotes solution, while standing water does not. Finally, the region must receive at least moderate rainfall. In general, arid and semiarid regions do not exhibit marked development of karst.

The Karst Process

In order to understand the formation of many karst features, it is necessary to first examine the chemical processes which form them. The solution of calcium is quite complicated, with numerous factors entering into the equation. Only a simplified disussion of the process, however, will be presented here. A more detailed discussion may be found in Picknett (1964).

The generalized steps in the solution of carbonates through the agency of carbon dioxide are as follows (Ewers, 1987):

(1) The absorption of carbon dioxide:

$$CO_2\ (g) \overset{HOH}{\Longleftrightarrow} CO_2\ (aq)$$

(2) The formation of carbonic acid:

$$CO_2\ (aq) + H_2O <=> H_2CO_3$$

(3) The dissociation of carbonic acid:

$$H_2CO_3 <=> H^+ + HCO_3^-$$

(4) The dissociation of water:

$$H_2O <=> H^+ + OH^-$$

(5) The dissociation of the carbonate mineral:

$$CaCO_3 <=> Ca^{2+} + CO_3^{2-}$$

(6) The formation of the bicarbonate ion:

$$CO_3^{2-} + H^+ <=> HCO_3^-$$

These chemical reactions are directly influenced by a number of environmental factors which determine the saturation of natural waters. These factors include temperature, partial pressure of carbon dioxide, mixing corrosion, flow rates, the presence of other ions, open and closed system conditions, and biological effects. A detailed discussion of each factor may be found in Bloom (1978), Jennings (1985), and Ewers (1987).

Karst Landforms

Numerous landforms characteristic of karst are found in the Bellevue-Castalia region (Plate 1). These include the fundamental component of karst topography, the doline or sinkhole, and such features as sinking streams, ponors or swallow holes, and caves and caverns.

Dolines (Sinkholes)

A doline is defined as a near-circular or elongate, funnel-shaped depression, which results from the solution or collapse of underlying strata. Five major classes of dolines -- solution, collapse, subsidence or alluvial, subjacent karst collapse, and alluvial streamsink -- are recognized (Figure 8). The two most contrasting types are the funnel-shaped solution doline and the steep or cliffed collapse doline. The solution doline is located where surface solution is concentrated which is often where two joints intersect. It is nearly circular in plan view with walls dipping anywhere from 30-40° (Bloom, 1978). The collapse doline, on the other hand, is generally caused by the collapse of the roof of a cave formed by underground solution. It will be vertically-walled and may be angular in plan view due to joint guidance. Through time, however, these dolines will become conical or bowl-shaped due to a wearing down of the sides and filling of the bottom.

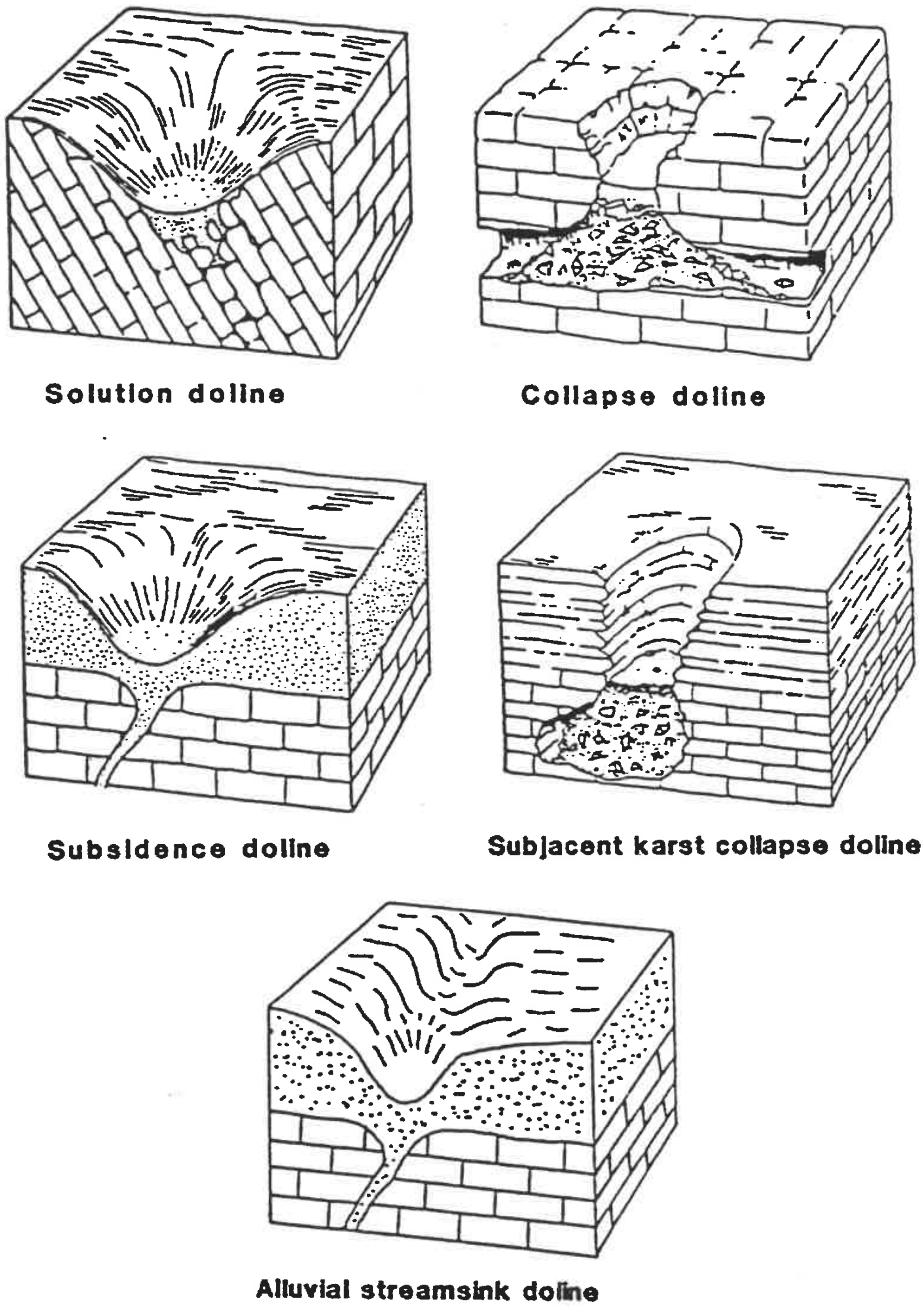


Figure 8. Five major classes of dolines (modified after Bloom, 1978 and Jennings, 1985).

Subsidence dolines, subjacent karst collapse dolines, and alluvial streamsink dolines are surface forms in nonsoluble rock caused by solution of a buried karst. Subsidence dolines are found where thick soils or superficial deposits cover karst rocks and through spasmodic subsidence and piping move downward into widened joints and solution pipes in the bedrock beneath. The size and shape of the dolines varies greatly through time. Subjacent karst collapse dolines are found in other rock formations overlying karst rocks. Often these dolines are formed by the collapse of a cave roof. They are generally elongate in plan view and fairly steep-walled. Alluvial streamsink dolines are found where streams sink through alluvium into underlying karst rock (Jennings, 1985). The processes involved in the creation of subsidence dolines are operative in the formation of alluvial streamsink dolines but, in addition, sinking streams help with the removal of the insoluble alluvium.

Four of the five major classes of dolines are visible in the Bellevue-Castalia area and are differentiated in Plate 1. The vast majority of the dolines belong to the subsidence type. In this case, glacio-lacustrine tills are representative of the superficial deposits overlying the karst rocks. A typical subsidence doline is shown in Figure 9. The largest doline in the study area is of the collapse variety. One such doline, located one mile (1.6 km) south of Castalia, is approximately 20 feet (6 m) deep and one

Figure 9. A typical subsidence doline in the Bellevue-Castalia area.

mile (1.6 km) in diameter. A bedrock cliff is exposed along the northern edge of the doline (Figure 10), while numerous caves are present around the eastern rim. This doline is very angular in plan view and is controlled by numerous joints. Several alluvial streamsink dolines are located along the channels of the two sinking streams; these streams will be discussed in the next section. The location of these dolines is dependent upon the stream stage at a given time. The final class of dolines found in the study area is the solution doline. Only one solution doline was positively identified in the study area, and it is nearly circular in plan view and has a dip of 33° (Figure 11).

Figure 10. View of the bedrock cliff located in the northern portion of the large collapse doline. See Plate 1 for exact location of this doline.

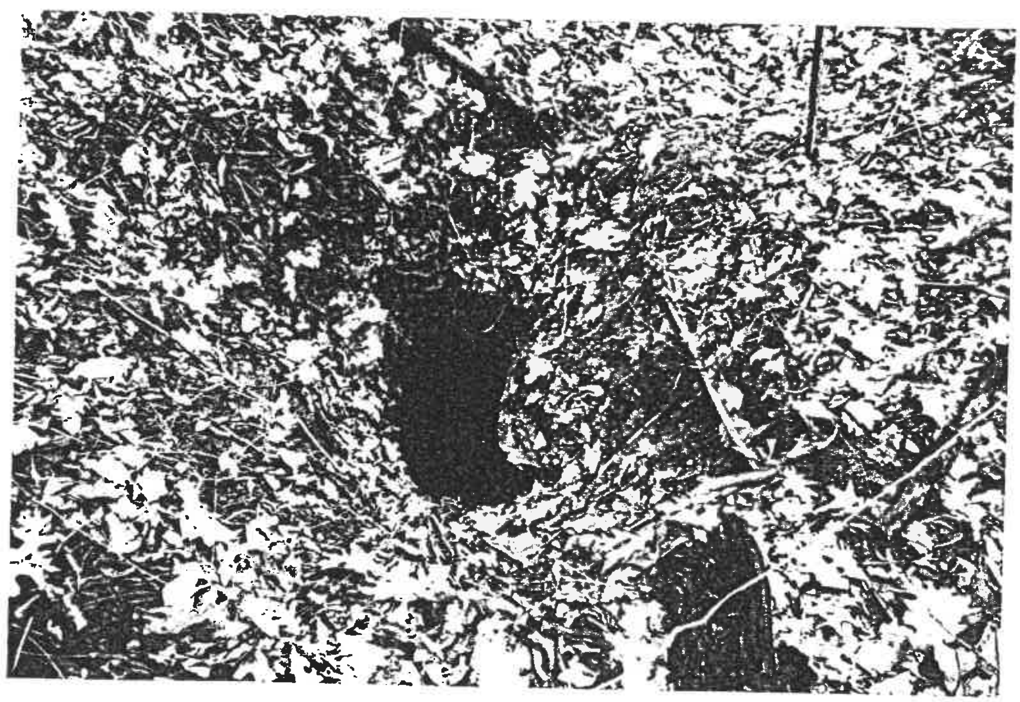


Figure 11. A solution doline located approximately four miles (7 km) southwest of Bellevue. See Plate 1 for exact location of this doline.

Dolines in the Bellevue-Castalia area are confined almost exclusively to the surficial extent of the Columbus Formation and are arranged in a fairly straight line trending N45E due to joint control. These dolines also form because the purity of the limestone (94 percent calcium carbonate) allows for easier solution by percolating waters.

Dolines in the study area do not appear to have been formed at the same time (Angle, 1987). Those located north of Bellevue contain very little sediment-fill, which might indicate that they were formed after glaciation. Those situated south of Bellevue, on the other hand, have increased sediment-fill, which may indicate that they were formed prior to glaciation and were subsequently filled in with glacial-related deposits.

Sinking Streams and Ponors

The term sinking stream refers to a stream or river which is liable to lose all or part of its volume underground, while a ponor is a low point where surface water enters the ground-water regime. Ponors or swallow holes may vary from cave entrances at the foot of limestone slopes to alluvial streamsink dolines. A stream may branch into numerous ponors or there may be several found in sequence along the course of the stream. Flood ponors may also be found positioned at levels higher than the normal place of engulfment (Jennings, 1985).

Two major sinking streams are apparent in the study area (see Plate 1). The largest of the two, known locally as Speck Creek, is located approximately two miles (3.2 km) northwest of Bellevue. The catchment area for the stream is 10.1 square miles (26.2 sq. km). The discharge of Speck Creek is highly variable. During periods of intense rainfall or snowmelt, the ponors cannot dispose of the runoff quickly enough, thus causing the formation of a lake covering several acres of farmland. It may take several days to a week for the stream to return to its channel. During periods of extreme drought, however, the stream will cease to flow. Numerous ponors are located in succession near the terminus of the stream channel (Figure 12). Several flood ponors are also found in the upper reaches of the channel. Their use depends on the stage of the stream at a given time.

The second of the sinking streams in the study area is known as Schneider's Creek. It is located approximately 4.5 miles (7.3 km) south of Bellevue and has a catchment area of 6.9 square miles (17.9 sq. km). Unlike the sequential arrangement of ponors in Speck Creek, Schneider's Creek branches into two large ponors (Figure 13). After intense periods of rainfall, a lake similar to that of Speck Creek will form. During dry spells only one of the ponors will usually contain water.

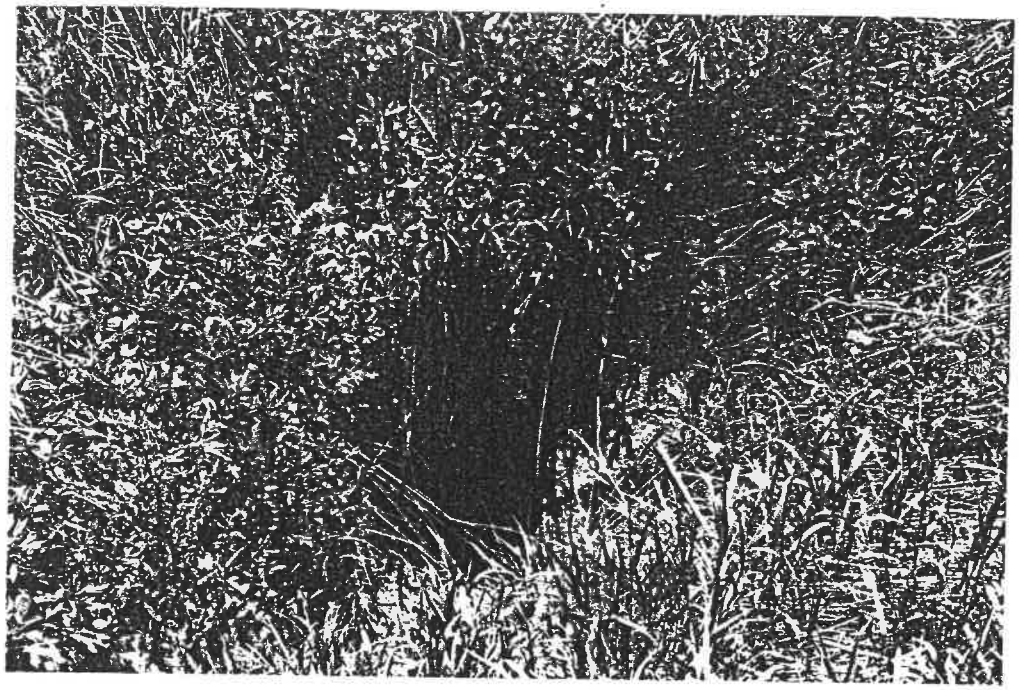


Figure 12. A ponor associated with an alluvial streamsink doline found near the terminus of Speck Creek.

Figure 13. View looking north where Schneider's Creek disappears into two ponors. These are located to the left and right center of the photograph.

Caves and Caverns

Several caves and caverns exist in the Bellevue-Castalia area. These include Seneca Caverns, Crystal Rock Cave, Brewery Cave, and several other small unnamed caves. Seneca Caverns is only mentioned here for it is discussed in detail in Chapter 4.

Crystal Rock and Brewery Caves are located just south of the town of Crystal Rock and are about 1,250 feet (381 m) apart (see Plate 1). Both are located in the Upper Silurian Put-in-Bay Formation. The largest of the two, Crystal Rock Cave, is approximately 40 feet (12 m) in length and 20 feet (6 m) in width with the ceiling being four to six feet (1 to 2 m) in height. Brewery Cave is approximately 20 feet (6 m) long and 10 feet (3 m) wide with the ceiling less than four feet (1 m) in height (G.W. White, 1926). Both caves are fairly void of any substantial solution features. It appears that the floor and ceiling of each cave were at one time together probably due to collapse rather than solution.

Several small caves have been found on the northeastern rim of the large collapse doline mentioned previously. These caves are similar to those of Crystal Rock in that the ceilings and floors appear to have been together at one time. There also is a general lack of any appreciable cave features.

Karst Hydrogeology

General

For many years, it was assumed that water in karst aquifers behaves in essentially the same manner as in porous aquifers. Water was thought to enter the system from the land surface, move downward through a horizon where the void spaces in the rock were dry, the vadose zone, and subsequently into a horizon where all cavities were perenially filled with water, the phreatic zone. The surface of the phreatic zone, or "phreas", was known as the water table, and where this surface intersects the land surface, springs occur. Not until recent decades, however, has this theory been refuted.

Most karst aquifers were found to contain a complicated network of passages controlled primarily by the development of secondary porosity, namely joints, fractures, and bedding planes. In the initial stages of karstification, there was no free intercommunication between these systems nor a single water table. Although each system had a number of input points, the passages joined together underground to feed a single or a few outlets. These systems were very similar in appearance to the dendritic drainage pattern of surface streams. As time progressed, however, these passages became solutionally enlarged and many networks became interconnected. This, in turn, led to the formation of a karst water table.

Aquifer Types and Characteristics

Aquifers, in general, are often regarded as vast underground resorvoirs capable of storing and transmitting large amounts of water. Water enters this reservoir through recharge areas and is ultimately released through discharge areas. In the study area, two types of recharge to the carbonate aquifer are prevalent depending on the precipitation rate. During moderate to heavy precipitation when runoff rates are high, recharge takes place directly through dolines and sinking streams. This is known as concentrated autogenic recharge (Gunn, 1985). When, however, the precipitation rate is low, recharge may occur directly where bare limestone is exposed at the surface or indirectly through the cover of surficial deposits. This type is known as diffuse autogenic recharge (Gunn, 1985). The vast majority of discharge in the study area takes place through springs, seeps, and to a lesser degree artificial discharge, baseflow, and evapotranspiration by plants. Artificial discharge or the discharge of ground water from wells varies immensely from one area to another based on the population's reliance on ground water versus surface water. Baseflow to surface streams is minimal because of the lack of appreciable surface drainage.

Three types of aquifers have been decribed -- confined, unconfined and semi-confined. Confined aquifers, also known as artesian or pressure aquifers, occur where ground water is confined under pressure greater than atmospheric by overlying, relatively impermeable strata. In a tightly cased well penetrating these aquifers, the water level will rise above the bottom of the confining bed. The water level at this point is representative of the potentiometric surface which is defined as the hydrostatic pressure level of the water in the aquifer. A potentiometric surface map may be constructed for a region by examining the water level in cased wells drilled into the aquifer. The resulting contour lines represent slope changes similar to that of a topographic map. The potentiometric surface may lie above or below the land surface depending on the hydrostatic pressure within the confined aquifer. Should the potentiometric surface lie below the land surface, the aquifer is considered to be non-flowing artesian. If the potentiometric surface is above ground level, the aquifer is flowing artesian. When, however, the potentiometric surface falls below the bottom of the confining bed, the aquifer becomes unconfined. This occurs where the confining bed thins appreciably or ends altogether.

An unconfined aquifer is one in which a water table varies in undulating form and in slope, depending on areas of recharge and discharge, pumpage from wells, and permeability. Fluctuations in the water table correspond to changes in the volume of water in storage within the aquifer. The water table normally follows the topographic relief. A semi-confined, or leaky, aquifer represents a combination of the two aquifer types.

The carbonate aquifer of the Bellevue-Castalia area exhibits characteristics of a confined, semi-confined, and unconfined aquifer. In Townsend Township and west-central Margaretta Township, where the confining layer of glacio-lacustrine till is thicker than 60 feet (18 m) (see Figure 6), the aquifer is considered confined. The location of this area is coincident with the zone of flowing artesian wells, which will be discussed later in this section. In the southern two-thirds of the study area where the confining layer is very thin or absent altogether, the aquifer is unconfined. The aquifer is thought to be semi-confined for some distance between the confined and unconfined areas.

Plate 2 represents the carbonate aquifer potentiometric surface. To update our knowledge of water level depths, over 100 bedrock wells were measured in October 1988 with a Roctest Model CPR6 water level indicator (Appendix A). These wells, which included domestic wells and several discharge wells in Bellevue, ranged in depth from 35 to 220 feet (11 to 67 m). The depth of these wells does not, however, cause any variation in the potentiometric surface because the water bearing zones within the carbonate aquifer are hydraulically interconnected. This was concluded by comparing the water levels within two closely spaced wells of different depths.

Ground-water flow direction in the Bellevue-Castalia area is controlled primarily by the bedrock topography of the region. Numerous grooves or troughs have been developed in the bedrock (see Figure 4) possibly as a result of erosion by the preglacial Erigan River system described by Spencer (1894). The large bedrock trough visible just west of Flat Rock and Bellevue and extending northeast and then northwest toward Sandusky Bay coincides nicely with the ground-water low seen on Plate 2. This large ground-water low undoubtedly represents the trunk of the system while other smaller deviations in the water table are the result of bedrock lows produced by tributaries of the main river. The hydraulic gradient for the area averages approximately 10 feet per mile (1.9 m/km).

Plate 2 shows the approximate location of flowing wells in the study area. Such wells, defined previously as those which have a potentiometric level above the land surface, are derived from the confined aquifer overlain by thick glacio-lacustrine sediments in the natural discharge zone. The variation in elevation between the surrounding beach ridges and bedrock highs, and the lake plains descending to the northwest result in the potentiometric surface being at or above the land surface. This, along with the very low permeability of the lake clay, is responsible for the zone of flowing artesian wells in the area.

Springs

Spring Characterization

Discharge from the subsurface takes place largely through springs. The rate of discharge via springs has been classified by O.E. Meinzer (1923, 1927, 1939). This classification was based on the magnitude of each spring's mean annual discharge (Table 5). Meinzer (1927) listed 24 first order magnitude springs occurring in limestone in the United States, while Linsley, Kohler, and Paulhus (1949) purported that there were hundreds of second order magnitude springs and thousands to tens of thousands of each of the lower magnitude types. In addition to the magnitude of discharge, springs may be classified based on their location and physical attributes (Bryan, 1919). Contact, sinkhole, fracture, and subaqueous type springs have been identified in the study area (Figure 14). Where permeable rock units overlie rocks of much lower permeability, a contact spring may occur. The lithologic contact between these two units is often marked by a line of springs at the contact of the main water table or perched water table and the land surface. It is not necessary that the underlying layer be impermeable, but merely that the difference in hydraulic conductivity be

Table 5. Meinzer's spring classification (after Meinzer, 1927).

Magnitude	English Units	Metric Units
First	Greater than 100 cfs*	Greater than 2.83 cms+
Second	10 to 100 cfs	0.283 to 2.83 cms
Third	1 to 10 cfs	28.3 to 283 liters/sec
Fourth	100 gal/min to 1 cfs	6.31 to 28.3 liters/sec
Fifth	10 to 100 gal/min	0.631 to 6.31 liters/sec
Sixth	1 to 10 gal/min	63.1 to 631 ml/sec
Seventh	1 pt/min to 1 gal/min	7.9 to 63.1 ml/sec
Eighth	less than 1 pt/min	less than 7.9 ml/sec

*cfs = cubic feet per second
+cms = cubic meters per second

great enough to preclude transmission of all of the water that is moving through the upper horizon (Bryan, 1919).

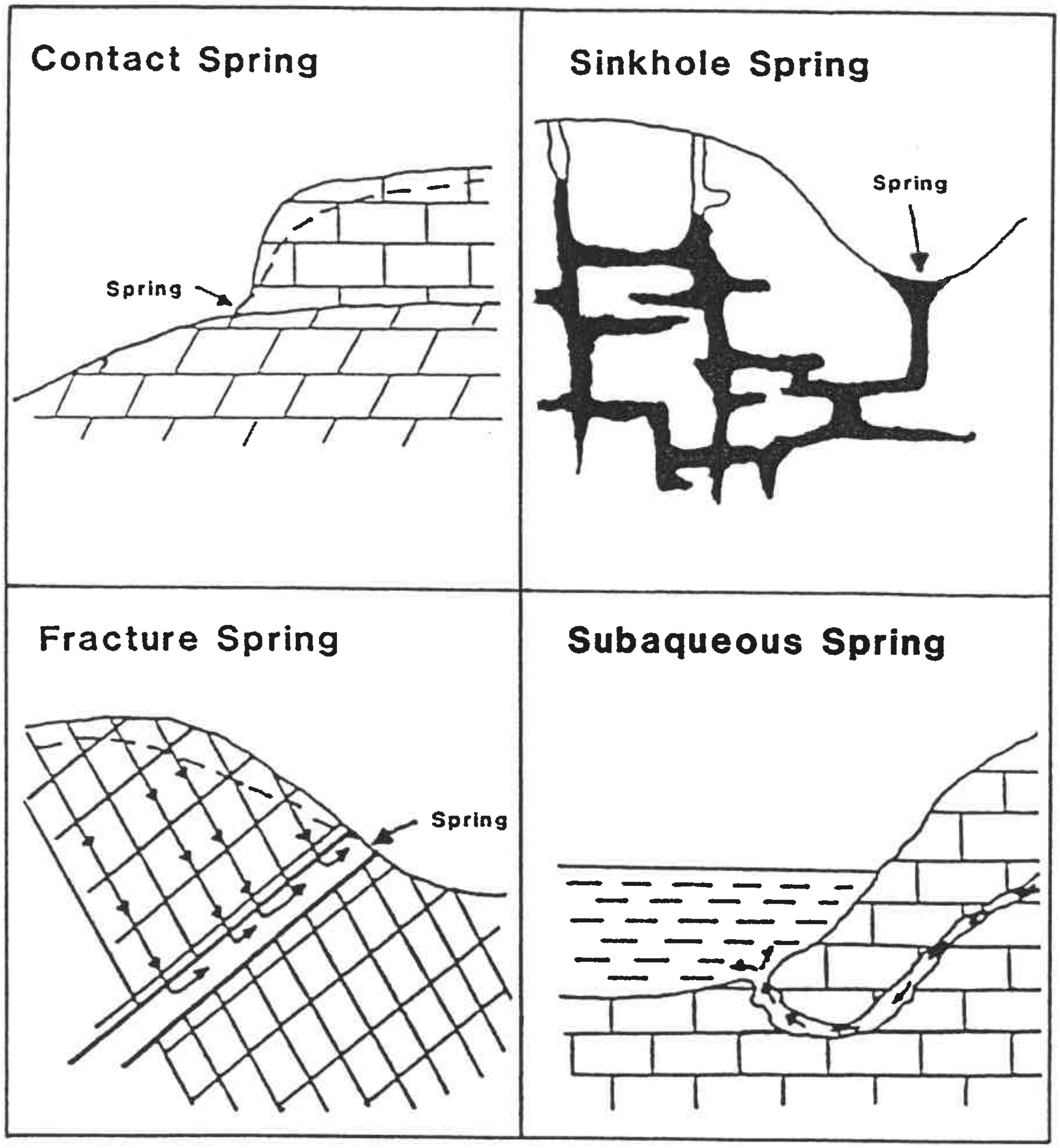


Figure 14. Spring types found in the Bellevue-Castalia area using the classification designed by Bryan (1919).

The second type of spring, the sinkhole spring, represents some of the largest springs in the world. In areas of carbonate bedrock, the runoff is carried primarily as subterranean flow through fractures in the rock or in caverns. These springs may be found where a cavern is connected to a shaft that rises to the surface. Water in these sinkhole springs is most often under artesian pressure. A specialized type of spring is found exclusively in collapse-type dolines produced by the solution of gypsum by artesian water in underlying formations (Quinlan, 1967).

Fracture springs may occur from the existence of jointed or permeable fault zones in low-permeability rock. Water movement within such rock is principally through fractures, and springs can form where these fractures intersect the land surface at low elevations (Bryan, 1919).

Subaqueous springs occur where a spring outlet is located below the water level of a lake or sea. Springs of this type may have been situated, at one time, above water level, but subsequently drowned by rising lake or sea levels.

Ten springs were located in the study area during the summer of 1986. These springs are located throughout the northern portion of the region between the villages of Castalia and Vickery (see Plate 1). A list of the springs along with their dimensions is included in Table 6. Discharge values were obtained for each of the springs during

Table 6. Names and dimensions of springs in the Bellevue-Castalia area.

Spring Name	Diameter (ft)	Depth (ft)
Duck Pond Spring	300	20
Castalia "Blue Hole"	75	45
Railroad Springs	175	10
Castalia Trout Farm Springs	100	45
Rockwell Spring	60	30
Rainbow Creek Spring	150	50
Ransom Springs	20	6
Miller's Blue Hole	500	30
Crane Spring	75	20
Muddy Spring	40	4

high (June) and low (October) flow periods in 1986 (Table 7). A Teledyne Gurley Model No. 625 pygmy meter was utilized for all of these measurements. The variability of a springs' discharge is determined through the use of the coefficient of variation (CV), which is defined as:

$$CV = 100\ s/\bar{x}$$

where s is standard deviation and $\bar{x}$ is the mean. The resulting coefficient of variation for each spring was computed and is in Table 7. These values range from 5.57 to 8.91 percent indicating a low variability in spring discharge.

The largest springs in the study area are close to Castalia (Figure 15). They are all located on or very near the contact between the Columbus and Lucas formations. This is due to the high solubility contrast between the very pure and more permeable Columbus Limestone and the relatively impure and less permeable Lucas Dolomite. The largest of these springs is Duck Pond Spring (Figure 16). Three conduits contribute waters to Duck Pond Spring (see Figure 15), two of which are located in the northwest corner while the remaining one is centrally located. Water may be seen 'boiling' at the surface above these outlets for much of the year.

Table 7. Discharge values for springs in the Bellevue-Castalia area.

Spring Name	Discharge, in gpm (cms)		CV (%)	Magnitude
	June	October		
Duck Pond Springs	6,775 (0.43)	6,407 (0.41)	5.58	Second
Castalia "Blue Hole"	4,494 (0.28)	4,032 (0.26)	5.57	Second/ Third
Railroad Springs	Trace	———	———	Eighth?
Castalia Trout Farm Springs	3,053 (0.19)	2,690 (0.17)	6.32	Third
Rockwell Spring	3,575 (0.23)	3,175 (0.20)	5.92	Third
Rainbow Creek Spring	754 (0.05)	650 (0.04)	7.41	Third
Ransom Springs	660 (0.04)	552 (0.03)	8.91	Third
Miller's Blue Hole	1,980 (0.13)	1,714 (0.11)	7.20	Third
Crane Spring	787 (0.05)	661 (0.04)	8.70	Third
Muddy Spring	105 (6.6×10^{-3})	90 (5.7×10^{-3})	7.70	Fourth/ Fifth

Cold Creek, a stream which never freezes, flows from Duck Pond Spring and empties into Sandusky Bay of Lake Erie. Three other Castalia area springs -- "Blue Hole", Railroad Spring, and Castalia Trout Farm Spring -- discharge into

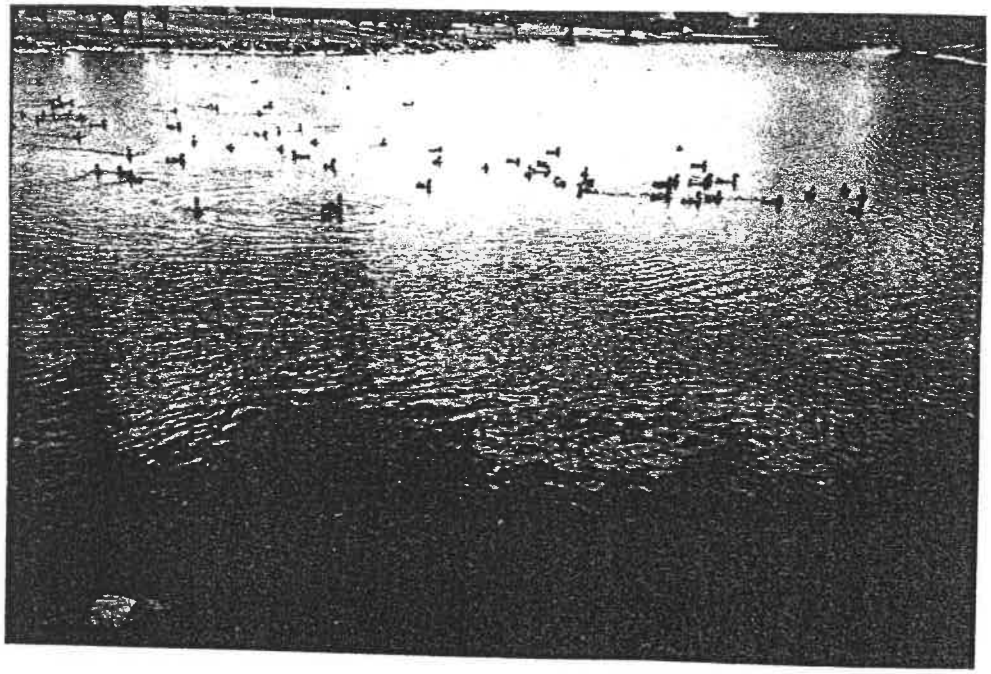


Figure 16. View looking south across the Duck Pond Spring located in the town of Castalia. Note water rises to the surface through the deep hole or conduit seen in the lower foreground.

Cold Creek along its course. The Castalia "Blue Hole" empties 4,032 to 4,494 gpm (0.26 to 0.28 m^3/s) into Cold Creek via two discharge streams (see Figures 15 and 17). This "Blue Hole", which has been developed as a tourist park, is undoubtedly the most famous of all the springs in the study area if not the state. "Blue Hole" was formed around 1820 when a dam and grist mill constructed on Cold Creek (see Figure 15) caused a rise in elevation of Duck Pond Spring, which increased pressure and weakened the strata above the "Blue Hole" causing it to collapse. Continuous undermining caused another cave-in in early 1914 which left "Blue Hole"

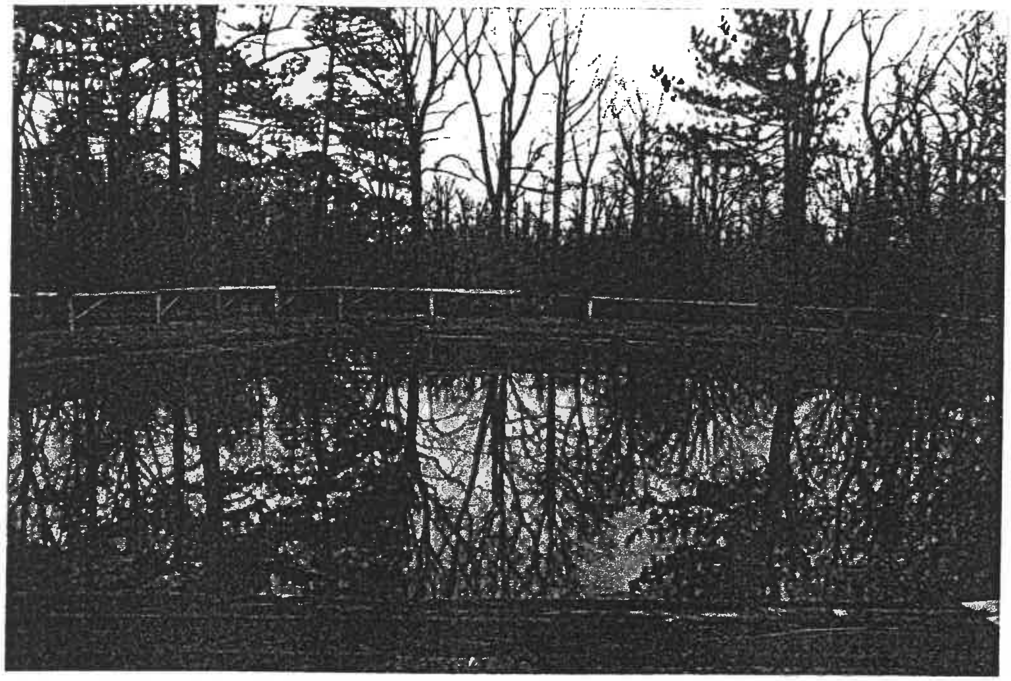


Figure 17. View looking east across Castalia "Blue Hole". Note one of the two discharge streams leaving "Blue Hole".

with its present dimensions, roughly 75 feet (23 m) in diameter and 45 feet (14 m) in depth (Hubbard, 1895). Hubbard (1895) reported that "Blue Hole" and Duck Pond Spring were hydraulically connected. When water was released from Duck Pond Spring through the mill dam, "Blue Hole" ceased to flow. Today, a small hatchery which raises trout for a private fishing club (Castalia Trout Club) is located adjacent to "Blue Hole" (see Figure 15) and utilizes the unfreezing waters which maintain a fairly constant 49°F (10°C) temperature throughout the year. This consistency in temperature is critical, for the trout cannot tolerate any drastic changes in temperature. In addition, trout water must contain sufficient dissolved oxygen; usually around 7 to 10 parts per million (ppm) of dissolved oxygen are necessary. However, water discharging from "Blue Hole" contains less than 1 ppm dissolved oxygen. This general lack of oxygen was thought to be: (1) from the presence of bacteria related to previous contamination from Bellevue; (2) from plants, fungi, and other microorganisms; and/or (3) from the conversion process of iron to iron oxide. All of these factors could consume large quantities of oxygen in ground water. To remedy this situation, surface aerators, water wheels, and waterfalls were installed to help add free oxygen to the spring waters.

Railroad Spring contains two conduits which contribute minimal amounts of water to Cold Creeks' volume through one discharge stream (see Figure 15). The only noticeable evidence of flow from these conduits is during the winter, at which time the pool is frozen except for directly over the conduits (Figure 18).

The last spring to empty its water into Cold Creek is the Castalia Trout Farm Spring (Figure 19). It contains two conduits located approximately 60 feet (18 m) apart. Presently, the Owens-Illinois Company commercially raises nearly 800,000 trout a year along a single discharge stream (C. Schmidt, 1987). The Owens-Illinois Company also operates a private fishing club along the lower reaches of Cold Creek.

Figure 18. View looking southeast across Railroad Spring. The water above the conduits is unfrozen.

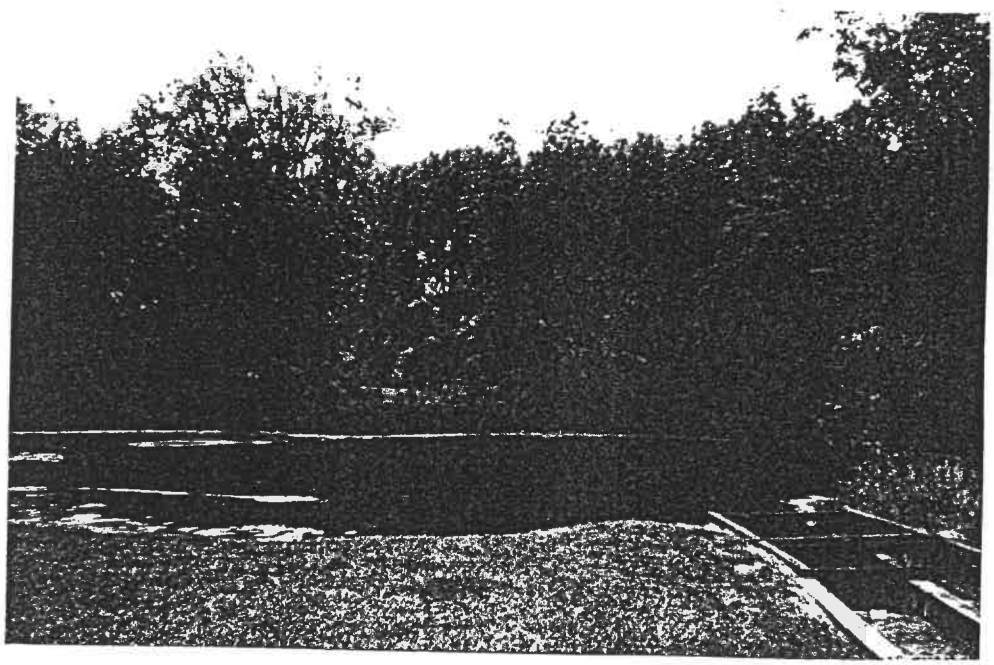


Figure 19. View looking north across the Owens-Illinois Company Castalia Trout Farm Spring. The two conduits are located to the left and right center of the photograph.

The remaining contact spring in this area is known as Rockwell Spring (Figure 20). The water from this spring forms Little Pickerel Creek and flows northwest into Sandusky Bay (Figure 21). A trout club, known as the Rockwell Springs Trout Club, has been established approximately one mile (1.6 km) downstream of the spring outlet (see Figure 21).

In the study area, three springs -- Miller's Blue Hole, Crane Spring, and Muddy Spring -- are classified as sinkhole springs. These springs are all developed in collapse dolines caused by the solution of gypsum in the underlying Bass Islands Group. Large deposits of marl, a crumbly soil consisting of clay, sand, and calcium carbonate, have been identified ringing these springs (Angle, 1987). These marl deposits are commonly known as spring mounds. Miller's Blue Hole (Figure 22) is the largest sinkhole spring with a diameter of of 500 feet (152 m), depth of 30 feet (9 m), and a discharge of 1,714 to 1,980 gpm (0.11 to 0.13 m^3/s). A large spring mound, approximately 10 feet (3 m) in thickness, is found encircling it. It does not, however, appear that marl is currently being deposited (Angle, 1987). The second largest of these springs is Crane Spring (Figure 23), with a diameter of 75 feet (23 m) and a depth of 20 feet (8 m). The spring water is bluish due to the color of the clay surrounding the spring (see Figure 23). The smallest of the sinkhole springs is Muddy Spring (Figure 24). It is only 40

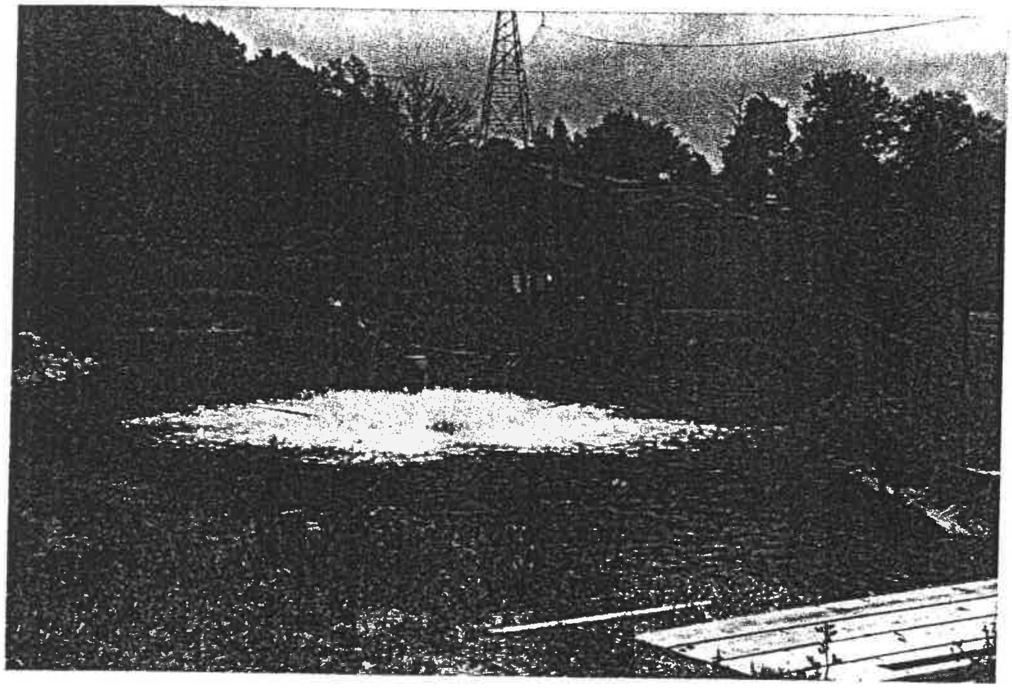


Figure 20. View looking southeast across Rockwell Spring, which forms Little Pickerel Creek. Note surface aerator in center of spring. The large hill in the background is part of the Columbus cuesta.

feet (12 m) in diameter and 4 feet (1 m) deep. Minimal amounts of marl have been deposited around the latter two springs (Angle, 1987).

Two springs -- Rainbow Creek Spring and Ransom Spring -- are of the fracture variety. Both springs are situated at the terminus of northwest trending fractures as was determined through the use of low altitude aerial photographs which will be discussed in detail later. The largest of these fracture springs, with a diameter of 150 feet (46 m), depth of 50 feet (12 m), and a discharge of 650 to 754 gpm (0.04 to 0.05 m^3/s), is Rainbow Creek Spring (Figure 25).

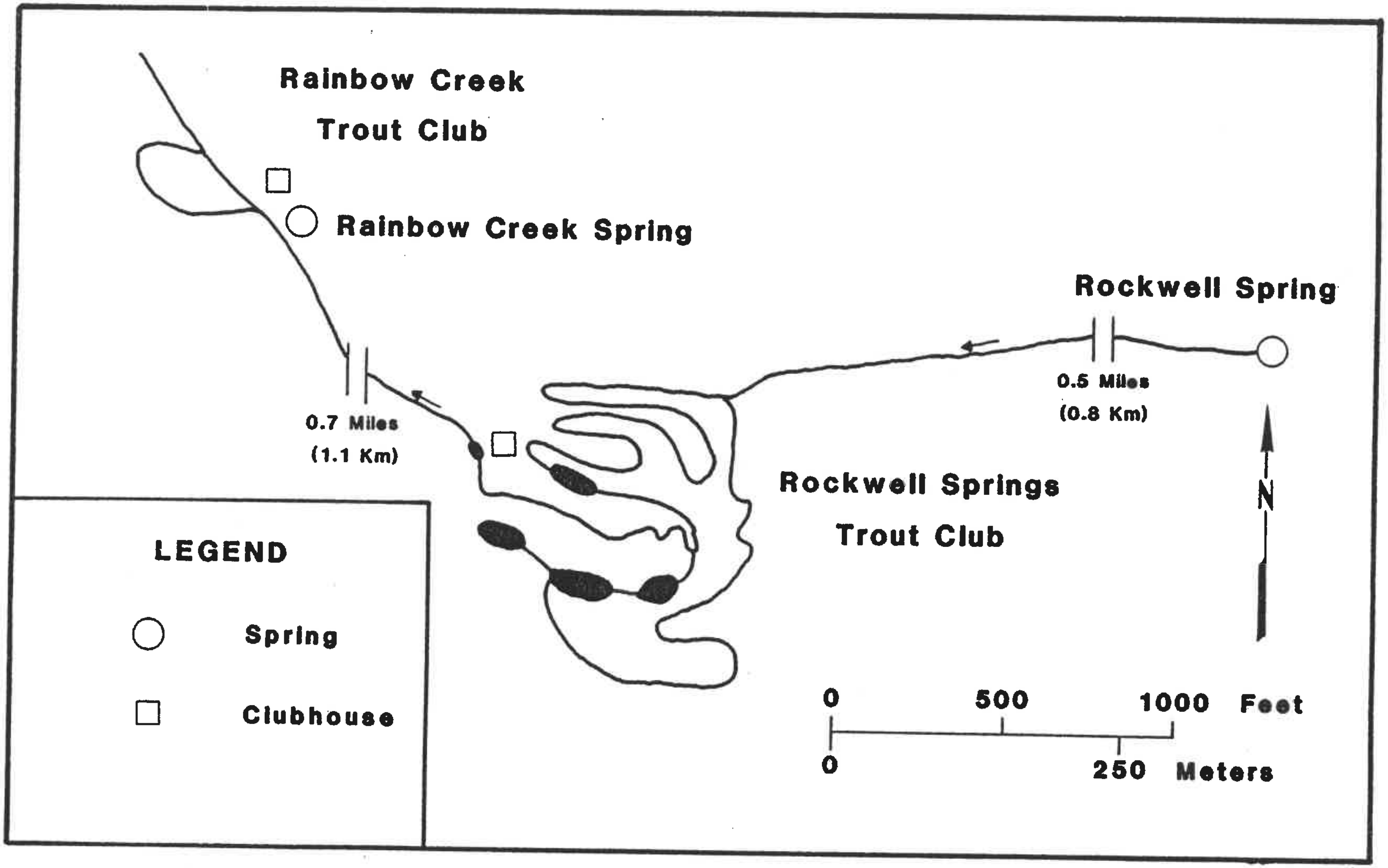


Figure 21. Map showing the location of Rockwell and Rainbow Creek springs and the channelization of Little Pickerel Creek.

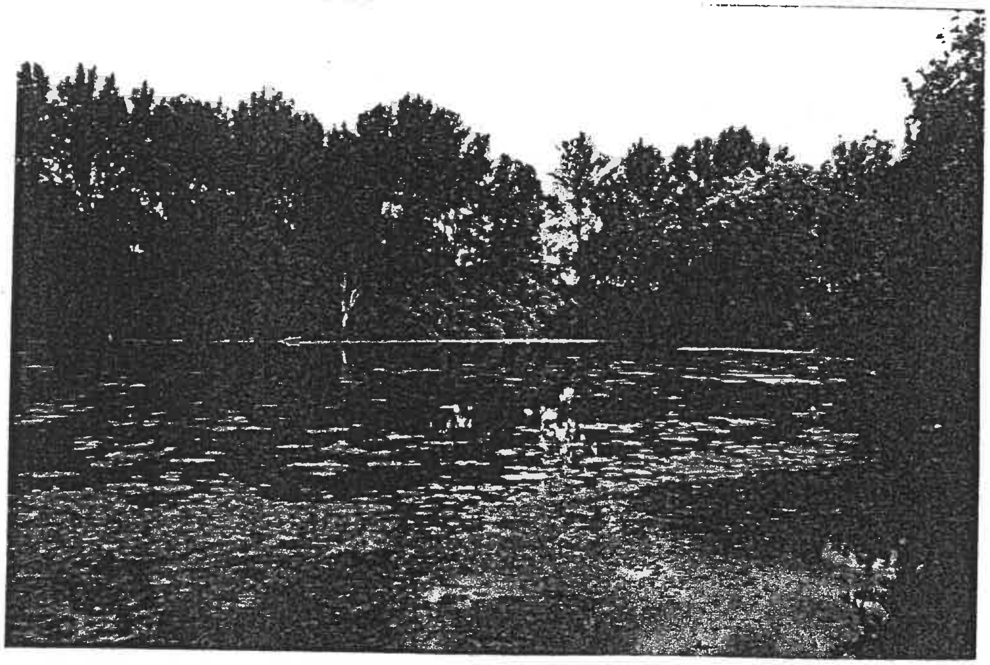


Figure 22. View looking north across Miller's Blue Hole. The marl spring mound is obscured by vegetation.

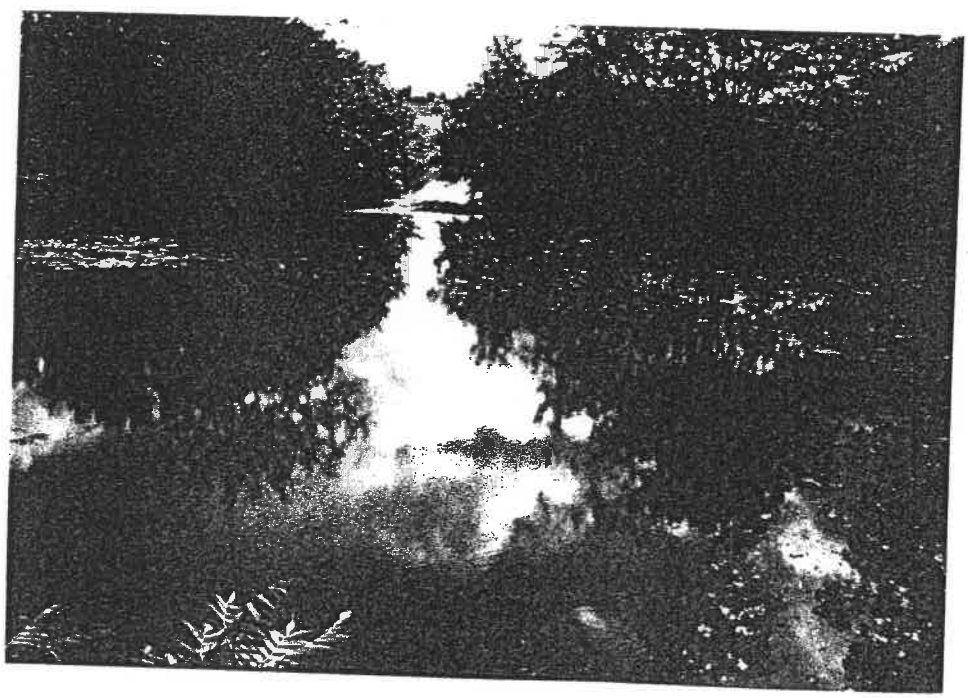


Figure 23. View looking east across Crane Spring. The aqua color is due to the presence of blue clay surrounding the spring.

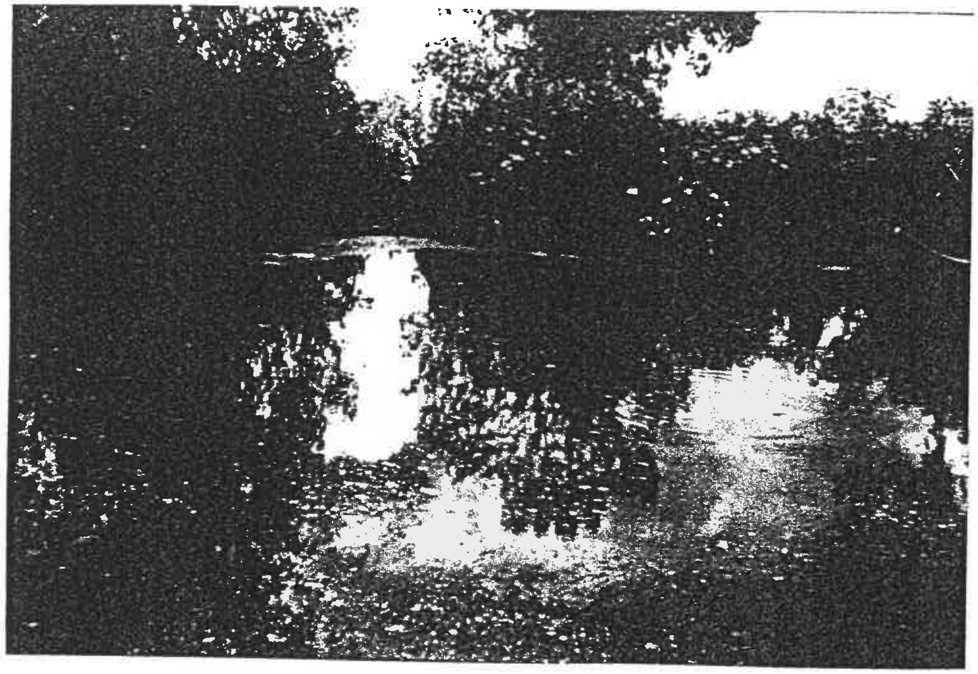


Figure 24. View looking south across Muddy Spring.

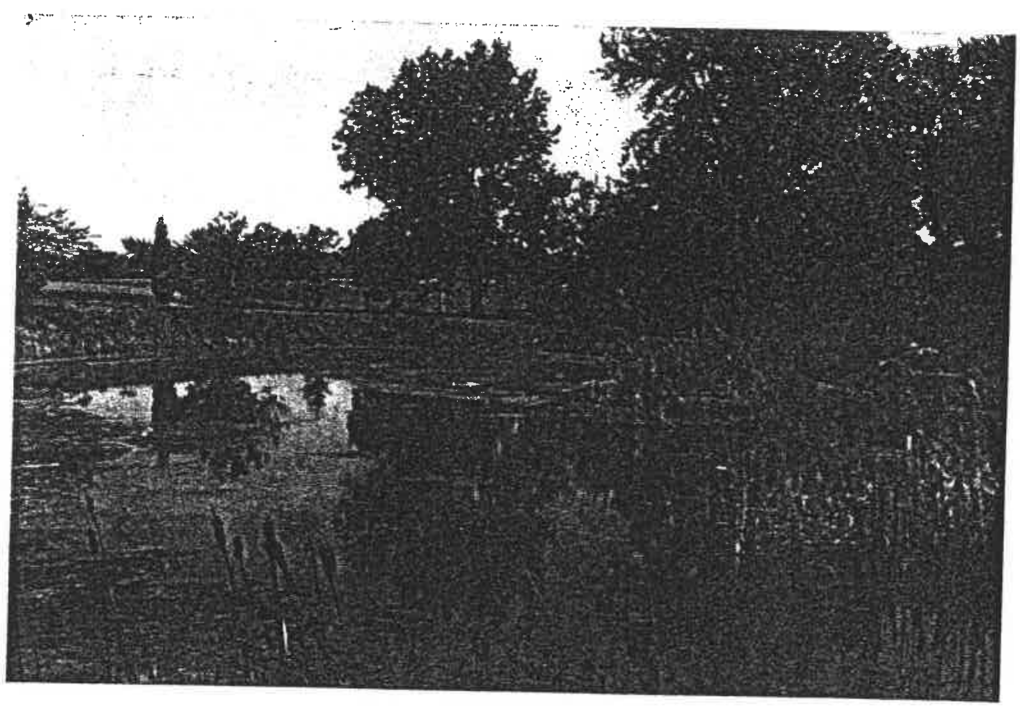


Figure 25. View looking southeast across Rainbow Creek Spring. The conduit is located in the far corner of the spring.

It is located approximately one mile (1.6 km) downstream from the Rockwell Springs Trout Club and is situated adjacent to Little Pickerel Creek (see Figure 21). A trout club, Rainbow Creek Trout Club, is currently under construction at this location and utilizes water from Rainbow Creek Spring and Little Pickerel Creek. The remaining fracture spring is Ransom Spring (Figure 26). It is located two miles (3.2 km) north of Castalia and discharges into Sandusky Bay at a rate of 552 to 660 gpm (0.03 to 0.04 m^3/s). Water from this spring outlet may be observed bubbling vigorously to the surface.

Near the study area, several subaqueous springs have been reported in the floor of Sandusky Bay (Norrocky, 1987). The location of these springs is most evident during the winter months when the lake waters are frozen except around the spring outlets, and early spring when clear pools of water are visible in the otherwise sediment-laden lake waters. During this study, no attempt was made to pinpoint the exact locations of these springs.

Spring Water Quality

Studies have been conducted by the ODNR, Division of Water (1968) and the USGS, Department of Interior (1972) to determine the physical and chemical properties of the spring waters in the study area. These studies, however, have focused exclusively on Castalia "Blue Hole" and Miller's

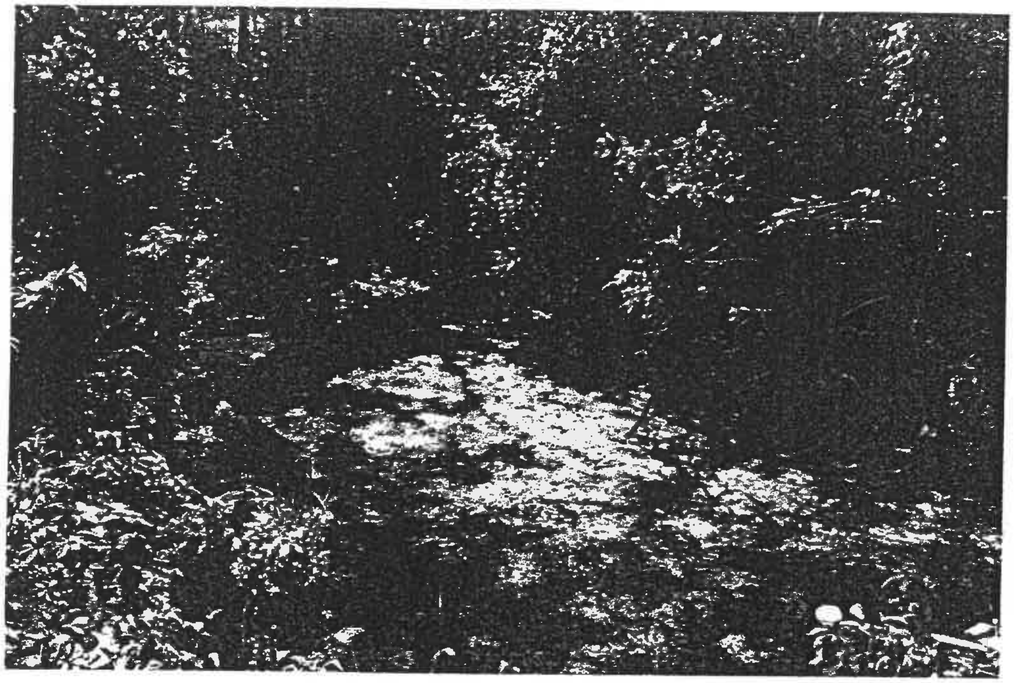


Figure 26. View looking west across Ransom Spring. Note bubbling water in the left center of the photograph.

Blue Hole, the results of which are found in Table 8. In addition to these tests, temperature and specific conductance measurements were periodically taken on all of the springs, and the results are in Table 9. These data were then examined to determine if the spring waters could be correlated with conduit or diffuse ground-water flow (which will be discussed in the next section) and, if the springs in and around Castalia are part of a different flow regime than those located farther west.

The most useful parameters from Table 8 for differentiating flow regimes are sulfate, total dissolved solids (TDS),

Table 8. Water chemistry of the Castalia "Blue Hole" and Miller's Blue Hole (after ODNR, Div. of Water, 1968 and USGS, Dept. of Interior, 1972).

Parameter	Castalia "Blue Hole" (mg/l)	Miller's Blue Hole (mg/l)
Silica (SiO_2)	8.8	9.3
Iron (Fe)	20	0.07
Manganese (Mn)	23	0.00
Calcium (Ca)	440	594
Magnesium (Mg)	35	36
Sodium (Na)	14	7.0
Potassium (K)	3.2	2.3
Bicarbonate (HCO_3)	295	328
Sulfate (SO_4)	**930**	**1,300**
Chloride (Cl)	24	16
Fluoride (F)	0.6	1.2
Nitrate (NO_3)	0.2	0.1
Total Dissolved Solids (calculated)	**1,600**	**2,130**
Hardness as $CaCO_3$	**1,240**	**1,630**
pH	7.4	7.6
Specific Conductance (micromhos at 20°C)	**1,850**	**2,160**

hardness as $CaCO_3$, and specific conductance. Unlike the other parameters listed, which are fairly consistent for both springs, these four parameters show a marked increase

Table 9. Specific conductance, TDS, and temperature values for springs in the Bellevue-Castalia area.

Spring Name	Specific Conductance (micromhos at 20°C)			Temperature (°F)	TDS (mg/l)
	Mean	SD	CV(%)		
Duck Pond Spring	1,558	53.1	3.41	47 - 50	1,640
Castalia "Blue Hole"	1,626	35.7	2.20	47 - 50	1,712
Railroad Spring	1,551	33.5	2.16	47 - 51	1,633
Castalia Trout Farm Spring	1,671	43.0	2.57	47 - 51	1,759
Rockwell Spring	1,424	49.0	3.44	48 - 50	1,499
Rainbow Creek Spring	2,209	8.2	0.37	47 - 51	2,324
Ransom Spring	2,195	33.9	1.54	47 - 51	2,309
Miller's Blue Hole	2,199	6.3	0.29	48 - 50	2,313
Crane Spring	2,171	33.8	1.57	46 - 51	2,284
Muddy Spring	2,163	34.5	1.60	47 - 51	2,275

in Miller's Blue Hole compared to Castalia "Blue Hole". Based on this information, along with specific conductance values collected from all the springs in the study area (see Table 9), it would appear that Rainbow Creek Spring, Ransom

Spring, Miller's Blue Hole, Crane Spring, and Muddy Spring are part of a flow system separate from that of Duck Pond Spring, Castalia "Blue Hole", Railroad Spring, Castalia Trout Farm Spring, and Rockwell Spring. Definitive dye tests, however, would be needed to substantiate this theory.

Conceptual Models for Flow in Carbonate Aquifers

Flow within a carbonate aquifer is determined primarily by the regional geology. Due to structural and stratigraphic complexities exhibited by many regions, two distinct flow systems -- diffuse flow and conduit flow -- have been identified (Figure 27). These flow systems may range from diffuse flow where flow occurs along bedding planes, joints, fractures, and other small interconnected openings, measured in centimeters, to conduit flow where flow is through an integrated network of solutionally-enlarged conduits which range in size from centimeters to meters in diameter (Shuster and White, 1971). Diffuse-flow is prevalent in regions where the carbonate rock has undergone very little solutional modification. Bedrock in these areas is normally composed of shaley limestones or coarsely crystalline dolomites (W.B. White, 1969). Flow in a diffuse system tends to be laminar and obeys, or nearly obeys, Darcy's law. A water table is generally very well established due to the high degree of interconnectivity of the secondary porosity. Where the aquifer is exposed at the ground surface, karst

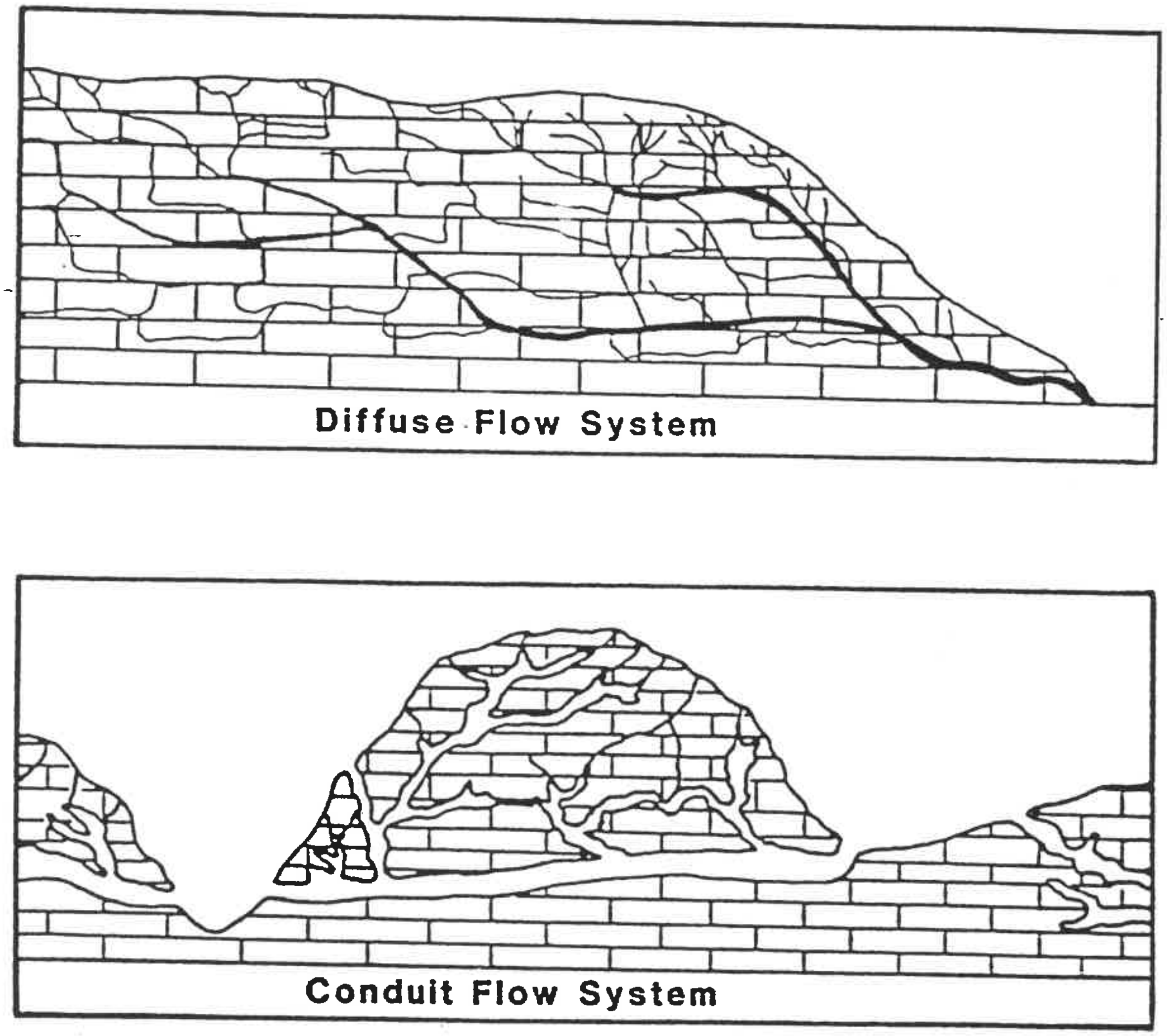


Figure 27. Flow models for carbonate aquifers (after Gaither, 1977).

landforms are subdued or absent altogether. The general relief of the region is fairly low as is the hydraulic gradient. Natural discharge from such a system normally takes place through a large number of smaller springs and seeps, or a few large springs that exist because of stratigraphic or structural features (Shuster and White, 1971). The discharge from these springs tends to be more constant than that from springs associated with conduit flow. Ground-water flow through the aquifer is relatively deep and is considered to be of the regional flow type as is described by Toth (1962 and 1963). Springs discharging from these systems have elevated TDS and hardness values along with fairly constant temperature and specific conductance readings.

Conduit-flow is common in regions where thick, massive limestone forms the bedrock. The ground-water flow paths in this system have been localized by solutional modification into well-integrated conduits (W.B. White, 1969). The major conduits often transport large volumes of water while the surrounding rocks carry only small volumes due to their very low hydraulic conductivities. The large volume flows may be on the order of meters per second and are normally in a turbulent state. There is no water table per se, but a network of solutional passages contributing flow to trunk conduits. The hydraulic gradient for the entire system is moderately high as is the topographic relief. Discharge is generally through a single spring and may vary over more than two orders of magnitude depending on precipitation. Response times are much less than those of diffuse flow systems, with a few hours being quite common. After storm events, springs discharging from conduits are often loaded with sediments which were transported through the system. Such springs have highly variable temperatures depending on the time of year and low total hardness values during periods of high flow. This is due to rapid flow-through times. During these high flow periods, water moves through the system so quickly that it does not have sufficient time to equilibrate with the surrounding rock. This is characteristic of Toth's (1962 and 1963) local flow system.

The carbonate aquifer in the Bellevue-Castalia area is primarily a diffuse flow system. The basic lithology of the aquifer, excluding the Columbus and Delaware formations, is a coarsely crystalline dolomite which, as a whole, has undergone little solutional modification. A definite water table is present, and the topographic relief and hydraulic gradient are fairly low. Discharge in the region takes place through numerous springs and small seeps. Several of the larger springs in the Castalia area are the result of a single stratigraphic feature -- the difference in permeabilities between the Columbus and Lucas formations. Groundwater flow through the aquifer is relatively deep and is considered to be of the regional flow-type as described by

Toth (1962 and 1963). This is illustrated by the springs' constant yearly discharge, temperature, and specific conductance values, as well as the elevated TDS and hardness readings (see Tables 7-9). In addition, the springs exhibit no signs of turbidity after storm events.

The only evidence of conduit flow in the Bellevue-Castalia area is the abundance of karst features which include dolines and sinking streams. These features are developed almost exclusively in the highly soluble Columbus Limestone and represent points of concentrated recharge of surface water into the ground-water system.

Fracture Trace Analysis

General

Ground-water flow in the Bellevue-Castalia area is assumed to occur through hundreds of joints or fractures. These features are the direct result of stress release by the rocks. Their abundance, distribution, and orientation is controlled primarily by the geologic structure of the area. The exact locations of these features may be determined through photogeologic investigations which entail the mapping and interpretation of natural linear features on aerial photographs. This practice has been widely used to locate such features which are not detectable in the field.

From such a study, it is hoped that the regional ground-water flow direction(s) in the study area could be determined.

In order to distinguish features recognized on aerial photographs, Lattman (1958) defined a photogeologic fracture or fracture trace, as:

> "A natural linear feature consisting of topographic (including straight stream segments), vegetation, or soil tonal alignments, visible primarily on aerial photographs, and expressed continuously for less than one mile. Only natural linear features not obviously related to outcrop pattern of tilted beds, lineation and foliation, and stratigraphic contacts are classified as fracture traces".

Lattman and Parizek (1964) further state that fracture traces may be revealed by aligned swallow holes, localized springs and diffuse seepage areas, and localized vegetational differences. A photogeologic lineament, or simply lineament, is the same as a fracture trace except that it is longer than one mile (1.6 km).

Fracture traces and lineaments are commonly straight in plan view and are unaffected by topography. They are thus considered to be the resultant of vertical to near-vertical zones of fracture concentration at the bedrock surface which reflect zones of increased permeability, weathering, and solution (Figure 28).

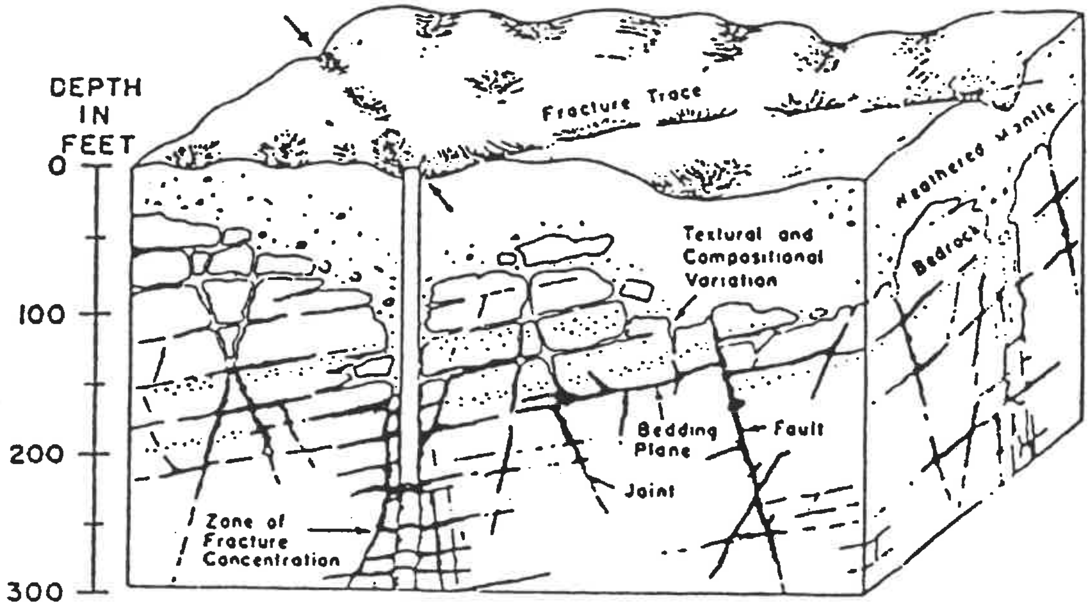


Figure 28. Block diagram showing factors influencing cavity distribution in carbonate rocks (after Lattman and Parizek, 1964).

Methods of analysis

Aerial photographs with a scale 1:24,000 were obtained from the Ohio Department of Natural Resources, Division of Soil and Water Conservation. These photographs were viewed with a 2 1/2 power lens stereoscope. The instrument was moved systematically over the photographs in a north-south direction. All observable fracture traces and lineaments were then marked on a clear overlay to preserve the photographs. Once this was completed, each photo was observed again without the use of the stereoscope. This was done to insure identification of linear features that escaped stereoscopic detection. All fracture traces and lineaments were then transferred from the photograph overlays to 7 1/2 minute quadrangle maps and ultimately reduced to the present scale seen on Plate 3.

An investigation of joint orientations also was undertaken to confirm the existence of the photolineaments. Four quarries -- Flat Rock, Bellevue, Parkertown, and Castalia -- were studied using a Brunton compass (Table 10). Twenty-five joints were randomly measured (Appendix B) at each of the four quarries, but fractures caused by blasting or other unnatural means were not included.

Fracture Trace Results

The locations of all fracture traces and lineaments are on Plate 3. A cumulative rose diagram of the photolineaments in the Bellevue-Castalia area is presented in Figure 29. A near orthoganol fracture pattern is apparent for the area, with the dominant trend being northeast-southwest and a less dominant trend northwest-southeast. This pattern is proposed to be the result of extensional stress generated in the Findlay Arch region as a result of differential subsidence of the adjoining Appalachian and Michigan Basins (Armstrong, 1976).

It is evident from Plate 3 that the greatest density of fracture traces and lineaments is found just southwest of

Table 10. Quarries utilized and formations encountered in the investigation of joint orientations.

Location	Owner and Present Status	Group or Formations Encountered
Flat Rock, Ohio	France Stone Co. (active)	Columbus Fm. and Detroit River Gp.
Bellevue, Ohio	France Stone Co. (inactive)	Columbus Fm. and Detroit River Gp.
Parkertown, Ohio	Sandusky Crushed Stone (active)	Delaware Fm., Columbus Fm. and Detroit River Gp.
Castalia, Ohio	Wagner Stone Co. Erie County Metropark (inactive)	Columbus Fm. and Detroit River Gp.

Castalia. The highest remaining portions of the Columbus cuesta are in this area. The soil cover is minimal which allows for the easier detection of fracture traces. In areas where the overburden thickens, namely west of the cuesta, the number of detectable traces decreases. The area just north of Castalia, however, is void of any traces because of the intense strip mining of marl in the past rather than due to the thickness of the overburden.

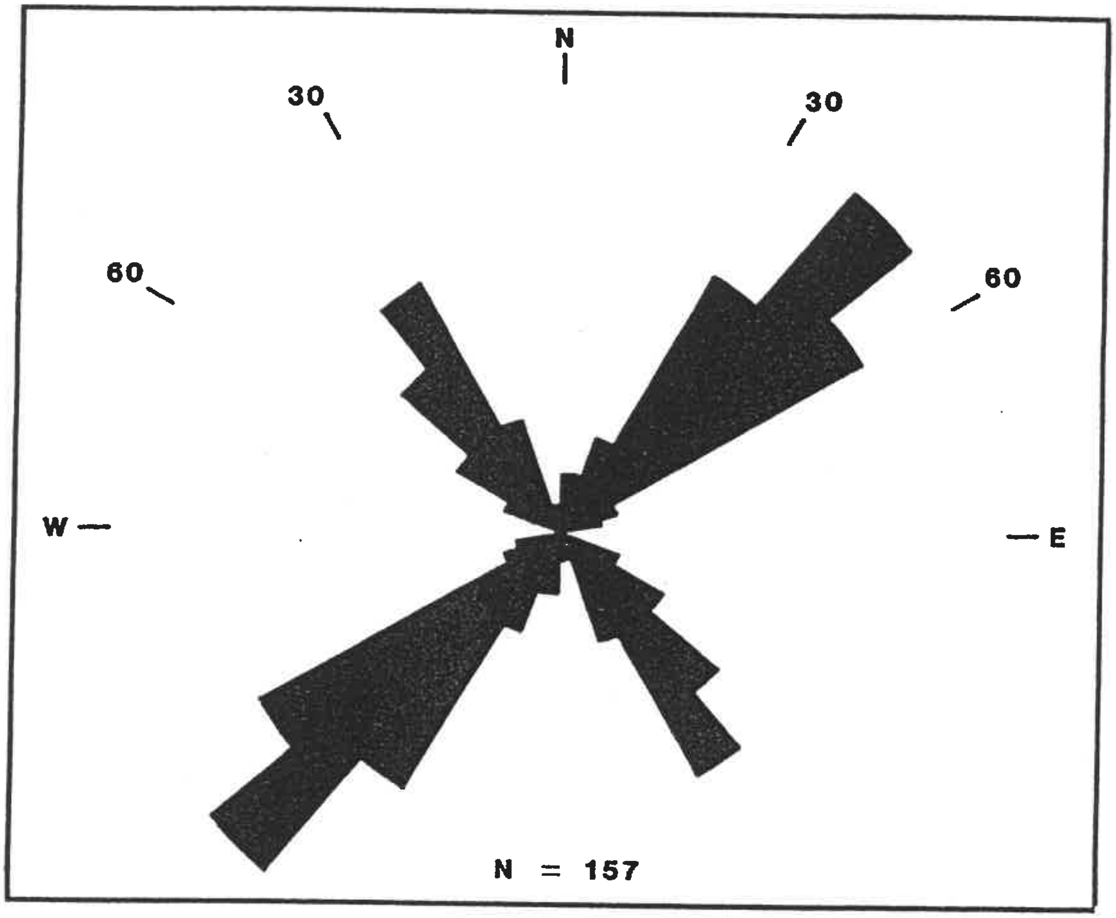


Figure 29. Rose diagram of fracture traces and lineaments in the Bellevue-Castalia area. N equals the number of readings.

Rose diagrams constructed from joint measurements in the quarries confirm this dominant northeast-southwest trend (Plate 3). It is also apparent from these diagrams that the nothwest-southeast direction becomes more noticeable further north in the study area.

Dye Trace Experiments

General

A dye trace refers to the tagging or labelling of a discrete quantity of water so that it can be identified at a later time. Dyes used in the tracing of karst ground water include Lycopodium spores, bacteriophage, radioactive substances, silica balls, and conservative ions. Fluorescent dyes, however, are the most commonly utilized dye group, of which Rhodamine WT and Sodium Fluorescein are the most widely used. The primary purposes of dye trace experiments in karst terranes are to determine flow-through times from an injection point to a recovery point and delineate ground-water basin divides. Injection points usually are sinking streams and dolines while the most common recovery points are springs.

Methodology and Results

Two fluorescent dyes were used to determine the direction and rate of ground-water flow through the carbonate aquifer in the study area. These included Rhodamine WT (CI Acid Red 388) and Sodium Fluorescein (CI Acid Yellow 73). These dyes were employed because of their high detectability, low sorptive tendency and low toxicity (Brown and Ford, 1971, Smart and Laidlaw, 1977, and Smart, 1984).

The initial dye trace commenced on May 5, 1987. Five pounds of Rhodamine WT were injected into a doline which was the sinking point of Speck Creek. This location was chosen in consultation with Dr. James F. Quinlan because of its close proximity to the recovery points (springs) and the presence of water to effectively carry the dye into the subsurface. Also, the ground-water flow direction, taken from the potentiometric surface map (see Plate 2), was to the north. The estimated discharge of the creek was between five and ten gallons per minute (3.1×10^{-4} to 6.3×10^{-4} m^3/s). The ten springs discussed previously, as well as the stream found within Crystal Rock Cave, were chosen as monitoring sites. Activated charcoal packets, mounted on a hanger and weight apparatus known as a gumdrop (Figure 30), were used to collect the dye. In addition, water samples were taken and analyzed on a Turner Model 111 filter fluorometer to measure the fluorescence of the sample. The activated charcoal packets were placed at the monitoring sites and water samples were taken one week prior to the beginning of the dye trace to check for background fluorescence. The charcoal packets were then changed twice a week once the dye trace began and were analyzed in accordance with the procedure described by Quinlan (1986). Water samples were also taken at the same time. After two months (May-June, 1987) of sampling the monitoring sites twice weekly, none of the dye was recovered. At this point, the

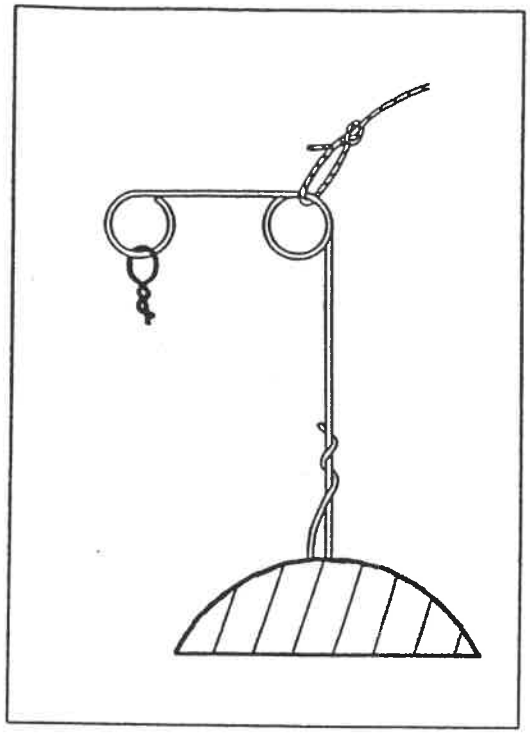


Figure 30. Gumdrop apparatus used to support the charcoal packet for dye retrieval.

sampling period was reduced to once a week for a period of another two months (July-August, 1987). No dye was subsequently found, however, and a second trace was proposed from the same injection point but with a greater concentration of a different dye.

During this phase of the dye tracing experiment, twenty pounds of Sodium Fluorescein were injected on August 25, 1987. Sodium Fluorescein was chosen because it could be analyzed on the fluorometer along with the possible presence of the Rhodamine WT dye from the previous test, with no threat of interference between these two dyes. Because of the lack of precipitation during the summer, Speck Creek had dried up. To remedy this situation, a pumper truck, provided by the Bellevue Fire Department, supplied the necessary water. A "primer" consisting of 700 gallons (2,652 l) of clean water was first pumped into the injection point. The dye was then mixed with 1,400 gallons (5,303 l) of water and injected at a rate of 75 to 100 gpm (4.7 x 10^{-3} to 6.3 x 10^{-3} m^3/s). This injection was followed by a "chaser" of 700 gallons (2,652 l) of water. At no time during the injection process was there any ponding of water, suggesting that the doline was very well developed and was capable of transmitting large amounts of water directly into the ground-water system.

The same monitoring schedule was employed as during the first test. Water samples were analyzed for the Sodium Fluorescein as well as for Rhodamine WT in the fluorometer. Weekly sampling was discontinued after four months with still no dye being recovered. Random samples were then taken monthly for five months (January-May, 1988) with no dye ever recovered.

The unexpected disappearence of the dyes may have been the result of one or more of the following:

(1) the amount of water present within the aquifer was underestimated, thus the dye was diluted beyond detection before it reached the monitoring sites;

(2) assuming the carbonate aquifer is of the diffuse flow type, it may be acting to store water rather than transmit it with any great velocity;

(3) the doline where the dye was injected may be feeding a local flow system operational at the till-bedrock interface, which in turn may not be discharging at any of the monitoring sites; and/or

(4) the dye may be moving along a deeper flow path and may be discharging at some other point(s), possibly through subaqueous springs located in the floor of Lake Erie.

SENECA CAVERNS

Location

Seneca Caverns is located in the extreme northwest portion of Thompson Township in Seneca County. It is situated approximately one mile (1.6 km) southwest of the village of Flat Rock and four miles (6.4 km) southwest of the city of Bellevue (N 41° 13' 35" latitude, W 82° 52' 30" longitude; see Figure 1).

History

Seneca Caverns was first discovered in 1872 by two area youths while hunting rabbits. The boys followed a rabbit into a brush pile and noticed a small fissure in the rock. They proceeded to squeeze through the opening and entered what is now the first level.

The upper three levels were operated as Good's Cave for three summers beginning in 1897, by Mr. Dunlop from Fremont, Ohio. These levels, however, were almost completely filled with clay that had washed in from the surface. Visitors who ruined their clothes while sliding from one level to another became discouraged and the venture failed.

In 1928, Don and Fannie Mae Bell took out a lease on the property and proceeded to clean out the cave. Eventually, a passageway was discovered leading down to an underground stream. On May 14, 1933 the cave was reopened under the name of Seneca Caverns. In 1964, Richard Bell purchased the Caverns from his parents and has commercially operated it to date.

Origin

The mode of origin of Seneca Caverns is to this day a mystery to many. What is evident, however, is that this is not a typical solution cavern as are those of the Mammoth Cave region in south-central Kentucky. There is no evidence of solution activity, such as scallops or anastomosies, inside the Caverns, nor does the ceiling display typical domepit features which are common in most solution caverns. Rather, Seneca Caverns are a collapse or breakdown-type cavern occurring along a major fracture in the Columbus and Lucas formations and possibly in the underlying strata. This mode of origin is illustrated by the displacement of beds which has been measured in numerous places at eight feet (2 m). The southern wall represents the downthrown block and is highly shattered and jumbled. There are places where the floor appears to fit into the ceiling much like a jigsaw puzzle. Slickensides, or polished surfaces that result from friction along a fault plane, show the movement to have been near vertical. The appearance of the rock, walls, and rubble support the fact that the Caverns were not formed by solution because these rocks are sharp and angular. However, some minor dripstone deposits have formed.

The probable processes leading to this collapse began with the conversion of anhydrite to gypsum in the underlying Bass Islands Group (see Table 1). This conversion process may entail a 33 to 62 percent increase in volume which would be instrumental in deforming stratified rock (G.W. White, 1926). The gypsum, being much more soluble than the surrounding bedrock, was preferentially dissolved by ground water moving downward and laterally through joints and bedding planes. This conversion process was used by Verber and Stansbery (1953) to describe the origin of the Put-in-Bay caves located on South Bass Island in Lake Erie approximately 14 miles (22 km) north of the study area. These caves are similar to Seneca Caverns.

Cavern Description

General

Seneca Caverns occurs within a large fracture in the Columbus and Lucas formations. From the photogeologic investigation discussed in Chapter 3, the fracture trends N68W for approximately 2 miles (3.2 km) and dips at an angle of 40NE (Figure 31). The total vertical extent of the frac-

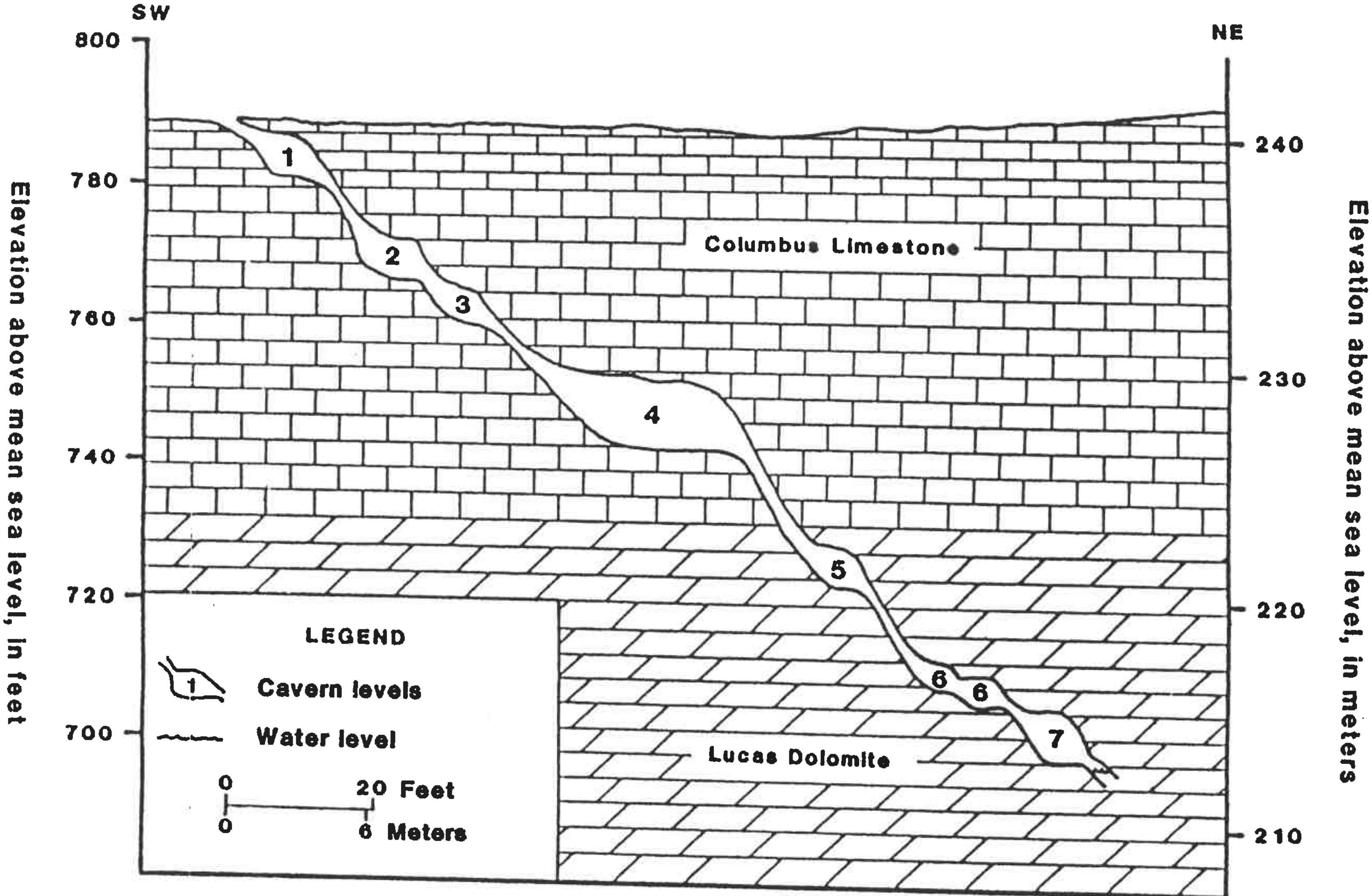


Figure 31. Cross-section trending NE-SW showing the dip of the fracture and the level arrangement of Seneca Caverns.

ture has yet to be determined because of the presence of water in the lower reaches of the Caverns. Seven distinct rooms or levels were visible during this study with water commonly found in the seventh level (see Figure 31). Because of the drought experienced by the midwestern states during the spring and summer of 1988, the water receded exposing a portion of the eighth level. The lowest water level, however, was observed in 1937. During this period, the water dropped an estimated 100 feet (30 m) below the seventh level exposing an additional five levels (Bell, 1986). The fracture, however, still appeared to continue downward below this point. The shape of the levels is generally rectangular in cross section. This is probably due to the joint pattern exhibited by the strata.

The first seven levels of Seneca Caverns were surveyed on September 18, 1985. Twenty-three survey stations were strategically located throughout the extent of these seven levels. These stations were subsequently marked with rock bolts and given numbers. A bench mark, representing a point of known elevation, was then located. From this point an alidade was employed to obtain an elevation at the cavern entrance. A Wild T1A theodolite was then utilized to measure the vertical and horizontal angles from station to station. This information was then computed to get an elevation for each station, as well as a straight line distance between each station. Where only one survey station was located, a tape and compass were used to obtain values for the length, width, and height of the smaller levels. In larger levels, where two or more stations were located, the method of triangulation was implimented as is described in Thompson and Taylor (1981). Data collected from these aforementioned techniques were then used to construct a cross-section of the seven levels (Figure 32). Thereafter, a detailed geologic investigation of each level was undertaken, with the results being outlined in the following sections.

Entrance and Level One

The primary entrance to Seneca Caverns is located about 30 feet (9 m) south of county road 178. It is not from a distinct sinkhole, but rather from a fissure which intersects the surface (G.W. White, 1926). Today, the entrance is approximately four feet (1 m) high and three feet (1 m) wide. A second entrance is located 47 feet (14 m) east of the primary entrance. This opening was originally very small and was subsequently enlarged to serve as an exit to cavern tours.

Level one is 10 to 12 (3 to 4 m) feet wide, 6 feet (2 m) high, and 115 feet (35 m) long (Figure 33). The south wall

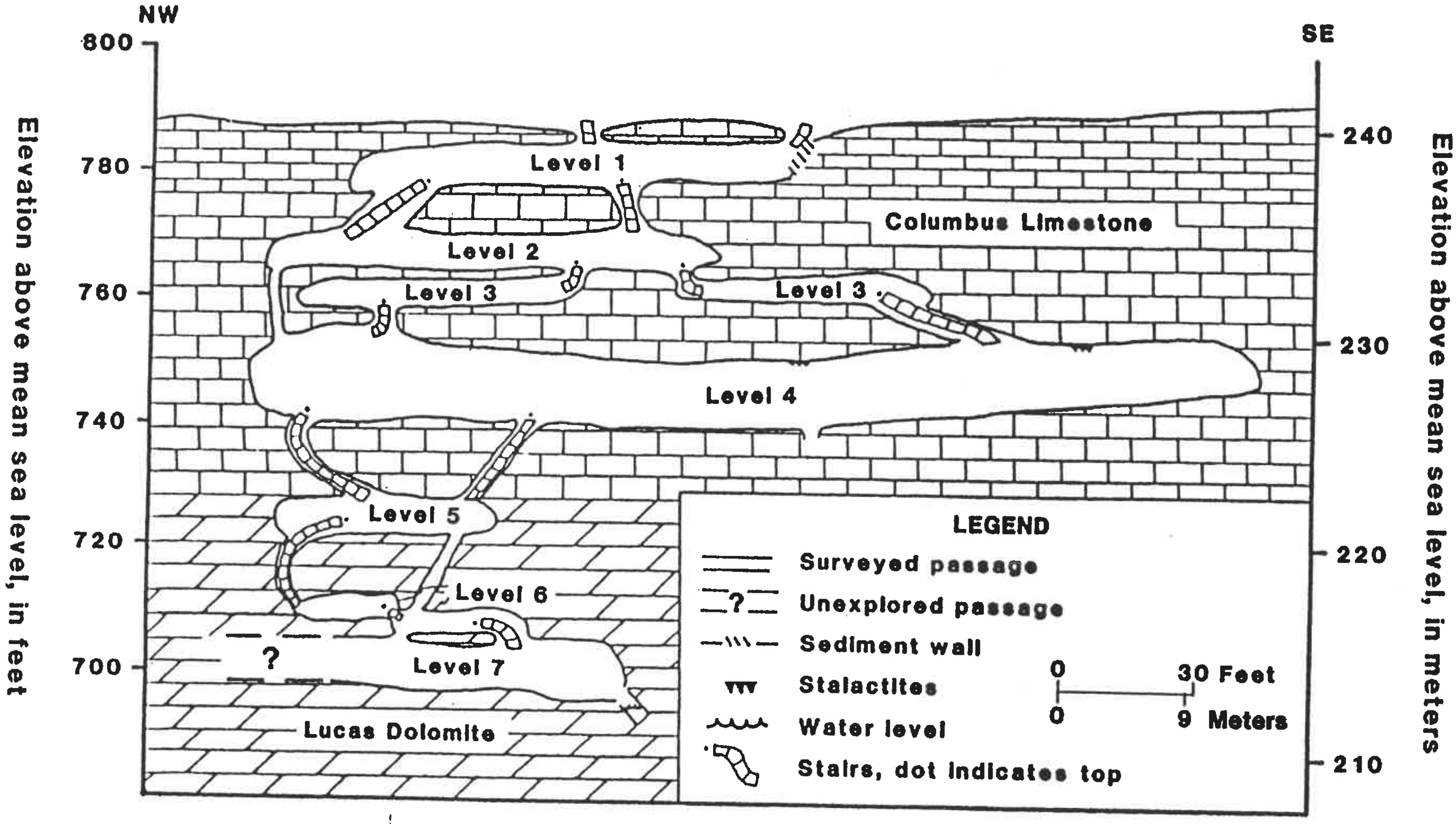
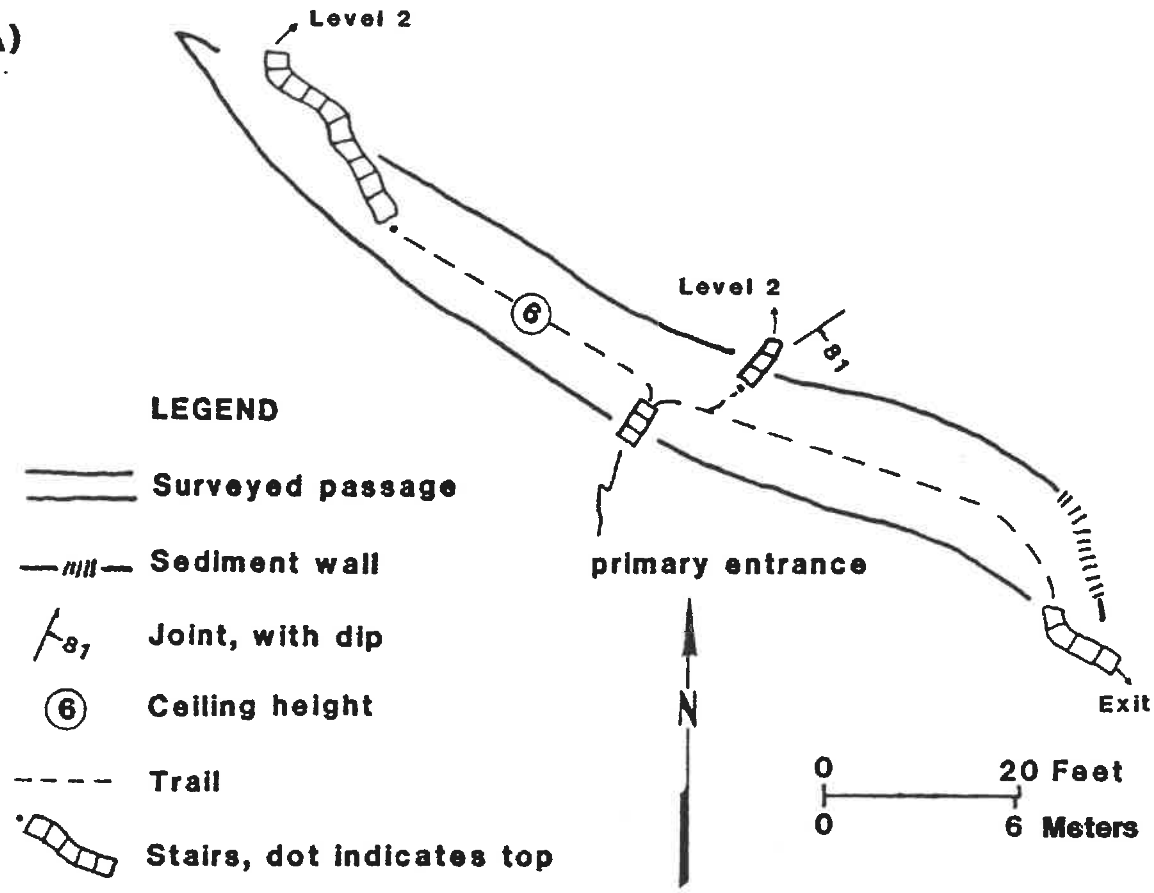


Figure 32. Cross-section trending NW-SE illustrating the seven surveyed levels of Seneca Caverns.

(A)



(B)

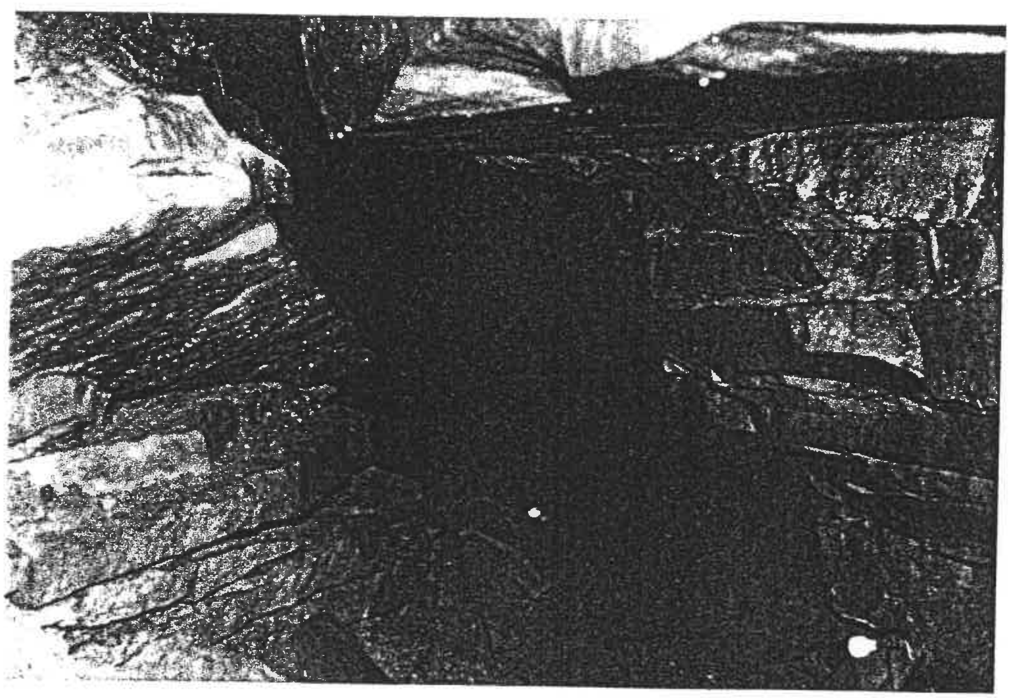


Figure 33. (A) Plan view of level one; (B) Photograph looking west from the primary entrance. Note jumbled appearance of strata to the left of the photograph.

and floor are covered with a thin layer of clay which was washed in from the surface. The east side of the first level, however, is filled almost entirely with clay. This area most closely reflects how the first level must have appeared at the time of discovery.

The first level is comprised of light gray, medium bedded, fossiliferous limestone. Brachiopods are the most common fossils and are represented by Leiorhynchus sp., Schizophoria propinqua, Chonetes sp., Mucrospirifer mucronatus, and Brevispirifer gregarius. In addition to the brachiopods, several specimens of Tentaculites scalariformis, which resemble a wood screw, are found scattered throughout the first level. Several stylolites averaging one inch (2.5 cm) in height are apparent on the north wall, two feet (0.6 m) above the steps in the western portion of level one. These features resemble sawteeth which are oriented at right angles to bedding and are the result of pressure solution in response to tectonic forces within the crust of the earth.

Evidence of collapse is apparent in Figure 33. The north wall (to the right in the photograph) is still intact, whereas the south wall has been severely broken up, most notably at the bedding planes. This is common throughout much of the Caverns. Since the north wall is in its original position and is relatively undisturbed, all level descriptions will be taken from this wall.

Level Two

Level two is 10 to 17 (3 to 5 m) feet wide, 6 feet (2 m) high, and 100 feet (30 m) long (Figure 34). Large amounts of clay are found blanketing the south wall between the two sets of steps descending from level one. Once again, this clay is not residual but was washed in through the entrance. The north wall is littered with large, angular blocks which have fallen from the ceiling. A shaft averaging four feet (1 m) in diameter is located at the extreme west end of level two and extends downward to level four.

The bedrock in this level is light gray, massively bedded limestone. It is highly fossiliferous with numerous large brachiopods and corals visible on the ceiling. *Megastrophia hemisphaerica*, *Brevispirifer gregarius*, *Chonetes* sp., and *Mucrospirifer mucronatus* comprise the brachiopod population, while *Heliophyllum halli* and *Zaphrentis corniculum* represent the horn corals.

Level Three

Level three consists of two separate chambers. The western chamber, known as "Earthquake Attic", is 8 feet (2 m) wide, 6 feet (2 m) high, and 60 feet (18 m) long (Figure 35). This chamber was once open to the public; however, numerous large blocks have fallen from the ceiling making it nearly impassible. The eastern chamber, known as "Tin Pan

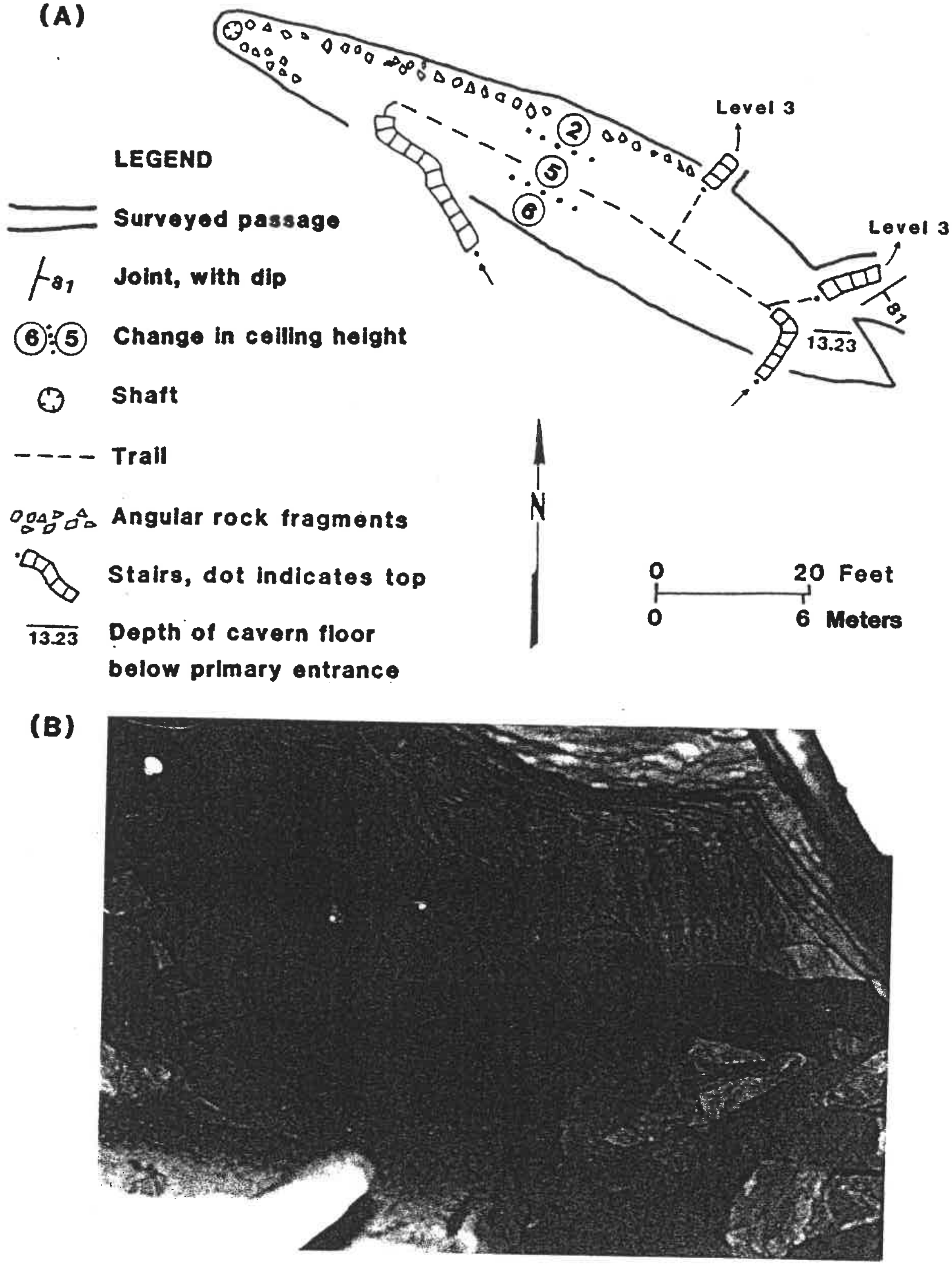
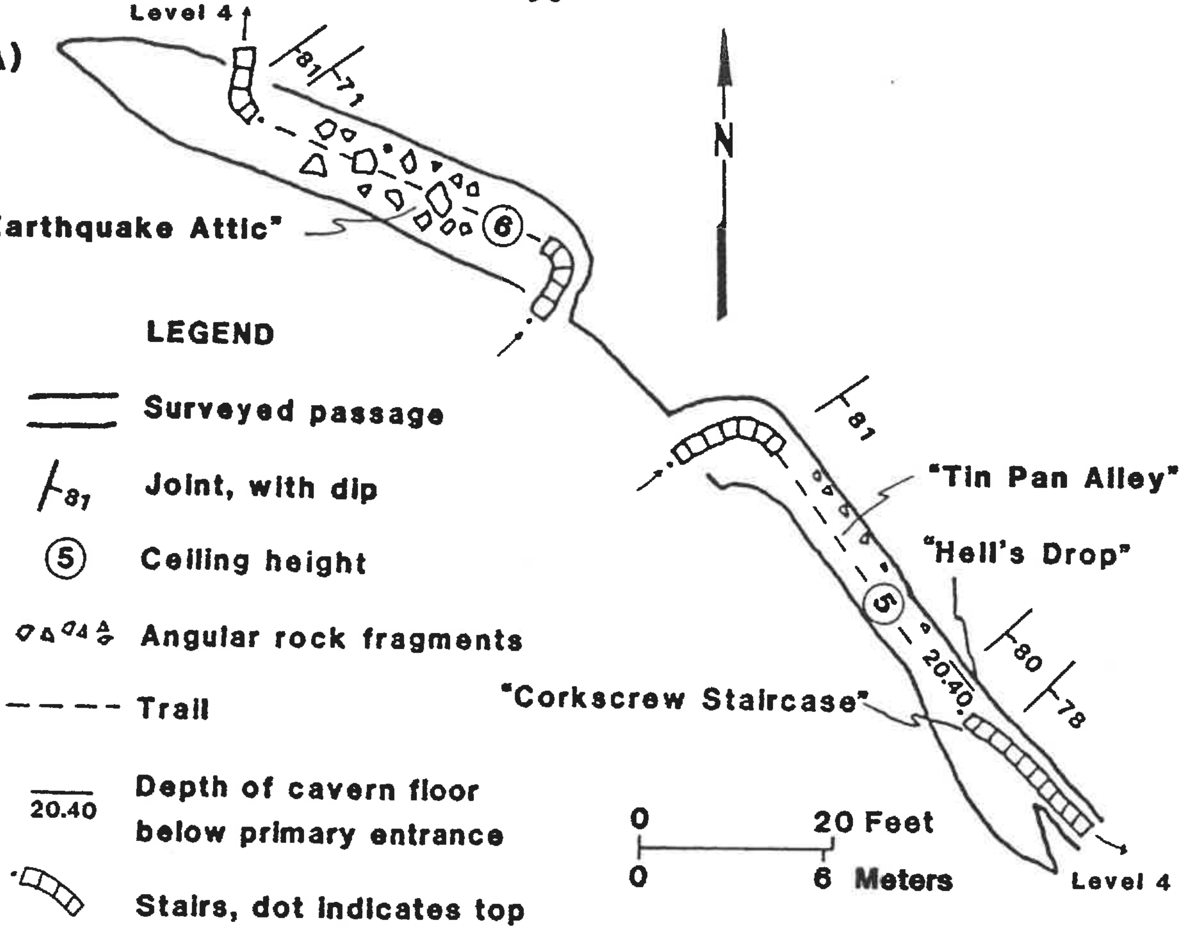


Figure 34. (A) Plan view of level two; (B) Photograph looking west from the east end of the room. Note clay deposits to the right of the photograph.

(A)



(B)

Figure 35. (A) Plan view of level three; (B) Photograph looking southeast through "Tin Pan Alley". "Hell's Drop" is located to the left of the railing.

Alley", is 5 to 10 feet (2 to 3 m) wide, 5 feet (2 m) high, and 60 feet (19 m) long (see Figure 35).

This level is composed of light gray, well jointed, massive limestone. Horn corals -- Zaphrentis corniculum -- and brachiopods -- Megastrophia hemisphaerica and Brevispirifer gregarius -- are less abundant than in the first two levels and are found in the upper portion of level three. Numerous chert nodules, ranging in diameter from two to six inches (5 to 15 cm), are most noticeable in "Earthquake Attic", whereas three chert layers averaging six inches (15 cm) in thickness are visible in "Tin Pan Alley". Also found approximately two feet (0.6 m) above the floor of level three is a continuous layer of brown, very thinly layered clay ranging in thickness from one to three inches (2.5 to 7.6 cm). This clay is highly plastic, meaning that it is capable of being molded into any form, which is then retained.

The steps descending to level four from "Tin Pan Alley" are called the "Corkscrew Staircase", while the ledge near the top of the Staircase is "Hell's Drop". "Hell's Drop" was formed when rocks fell away from a joint plane forming an overlook into level four.

Level Four

The fourth level, called the "Big Room", is the largest, being 225 feet (69 m) long, 20 to 25 feet (6 to 8 m) wide, and averaging 8 feet (2 m) in height (Figure 36). The level as a whole is arcuate in shape. The eastern portion of the room trends nearly north-south for approximately 80 feet (24 m) and then bends northwestward. A 40 foot (12 m) section located in the extreme eastern portion of level four has been given the name "The Gallery" (Figure 36). It is an area which stands about three feet (1 m) higher than the actual floor of the fourth level and was so named because the chert nodules in the fallen rocks along the south wall were once said to have looked like a group of paintings. A shaft, known as "Devil's Leap", is apparent on the northern wall (Figure 36). This feature extends vertically into the eighth level and was formed along the same joint plane as was "Hell's Drop". This joint is the most developed in the caverns and, thus, is termed the master joint. The bulk of the rain and meltwater is carried downward through this joint, as is evidenced by the high concentration of drip-stone deposits found on the ceiling surrounding it. These deposits are represented by bacon ribbing and numerous stalactites, the largest of which are only about one inch (2.5 cm) in length. Lesser deposits, however, are found throughout the fourth level wherever a joint is present. A large breakdown block is apparent on the north wall just before

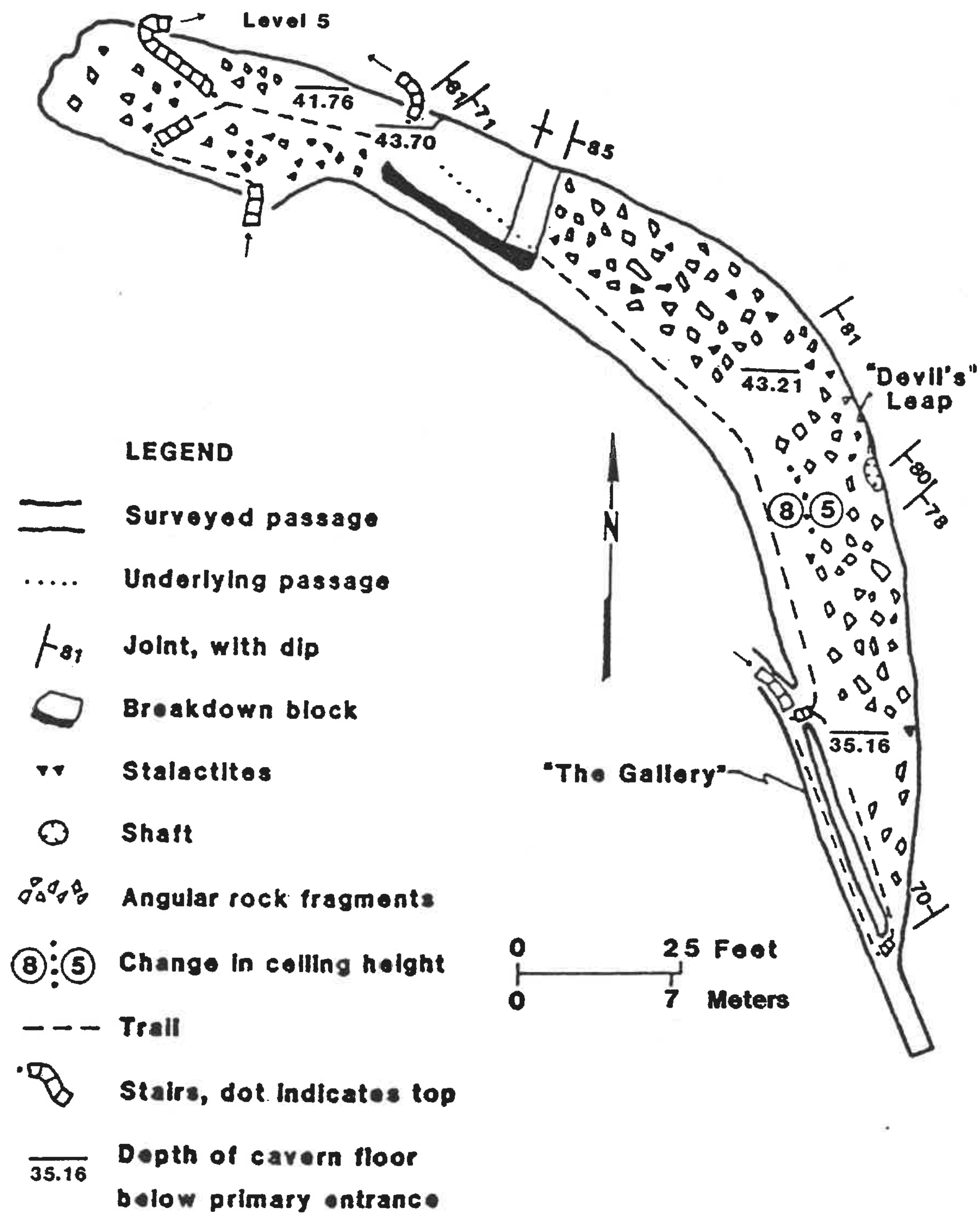


Figure 36. Plan view of level four.

the steps leading into the fifth level (Figure 37). The shape of this block is controlled by joints located on either side of it. A large amount of rubble is found throughout much of this level. This again may be due to the high degree of jointing which weakened the rock.

The bedrock exposed in the fourth level is light gray, massive dolomitic limestone. The level is relatively unfossiliferous except for a four foot (1 m) section near the top, where numerous specimens of _Zaphrentis_ _corniculum_ are found. It is finely laminated near the bottom of the level, which is most apparent in the western portion of the level near the steps descending into the fifth level.

The best evidence supporting the theory that the roof and floor of Seneca Caverns were at one time together may be seen in the eastern portion of this level (Figure 38). Here the floor is still intact and fits almost exactly against the ceiling. The amount of separation at this point is eight feet (2 m).

Level Five

Two sets of stairs lead downward from the western end of level four into a much smaller fifth level. This level is 55 feet (17 m) long, 6 feet (2 m) high, and 10 feet (3 m) wide (Figure 39). A shaft with a diameter of about six feet (2 m) is located at the eastern end of the level and extends downward through the sixth and to the seventh level.

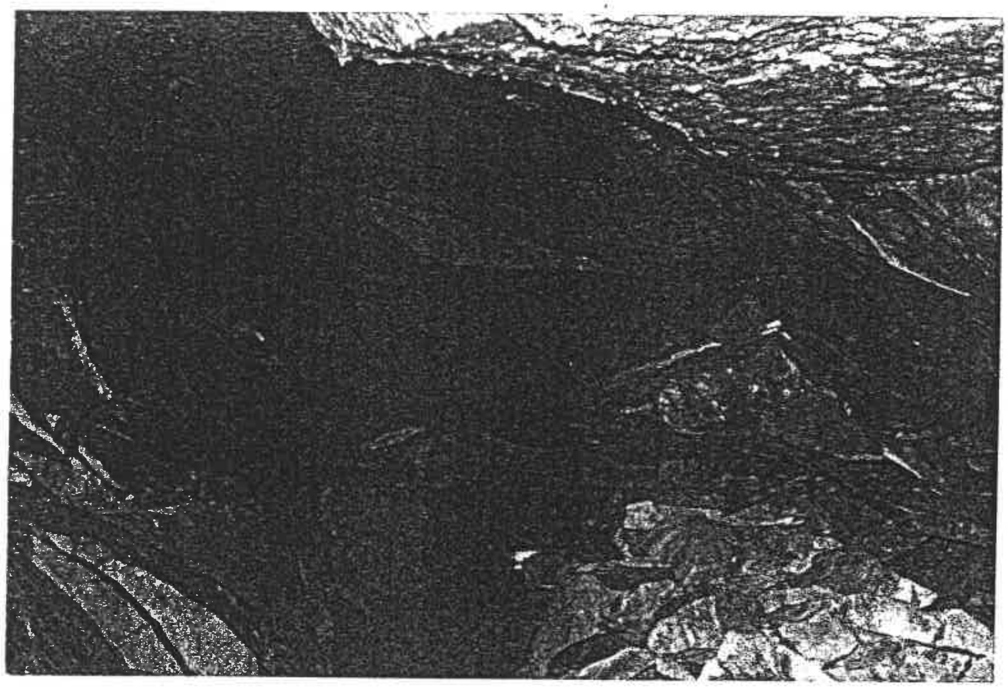


Figure 37. View looking west across the breakdown block in level four.

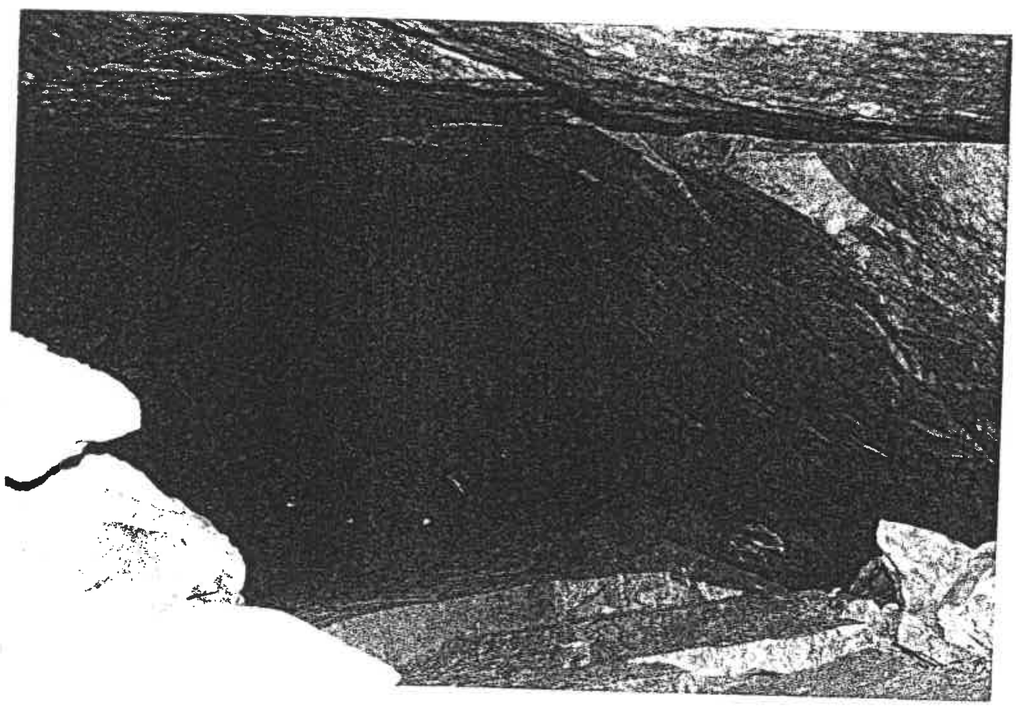


Figure 38. View looking north through the southern portion of level four. The large rock comprising the floor fits nicely into the ceiling rock.

(A)

(B)

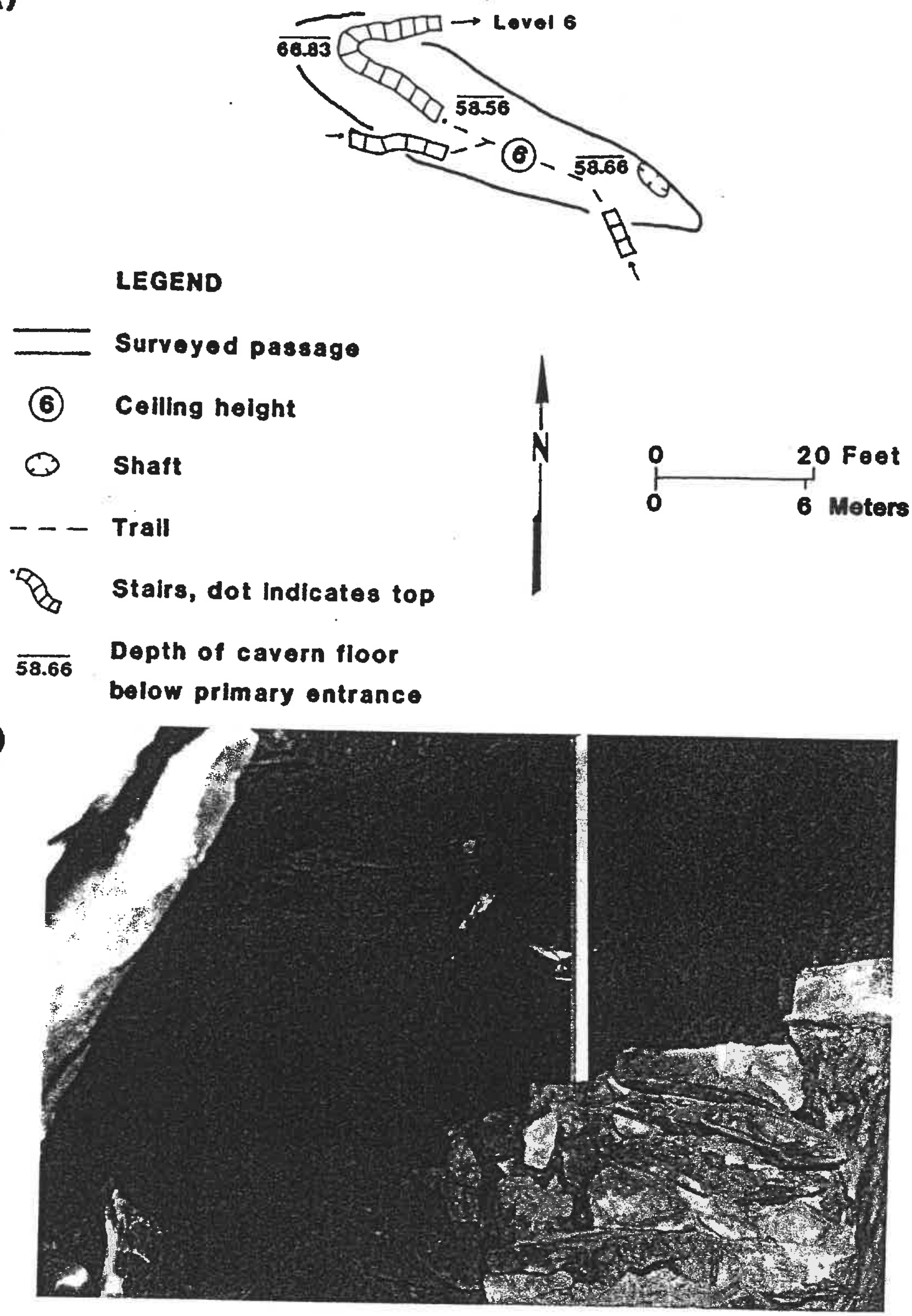


Figure 39. (A) Plan view of level five; (B) Photograph looking east from the west end of the room. Note railing for staircase from level four in center of photograph.

The bedrock in level five is light brown, finely laminated, massive and sugary textured dolostone. It is unfossiliferous and unjointed.

Level Six

The sixth level consists of two small chambers termed the upper and lower chambers (Figure 40). The upper chamber is 20 feet (8 m) long, 7 feet (2 m) high, and 4 feet (1 m) wide, while the lower is 28 feet (9 m) long and 5 feet (2 m) in width and height. Both chambers consist of a light gray, massive dolostone which is unjointed. The lower chamber, however, contains a three to six inch (7.6 to 15 cm) layer of chert located three feet (1 m) above the floor. Also apparent are several vugs or cavities which are lined with calcite crystals. Most of these vugs are the result of corals which have been preferentially dissolved by ground water.

Level Seven

The lowest surveyed level in Seneca Caverns is the seventh level or the "Water Room". It is about 65 feet (20 m) long, 8 feet (2 m) wide, and 8 to 10 feet (2 to 3 m) high (Figure 41). The floor is approximately 6 feet (2 m) wide and is found along the south wall. It is fairly level in the western one-half of the room but dips sharply downward

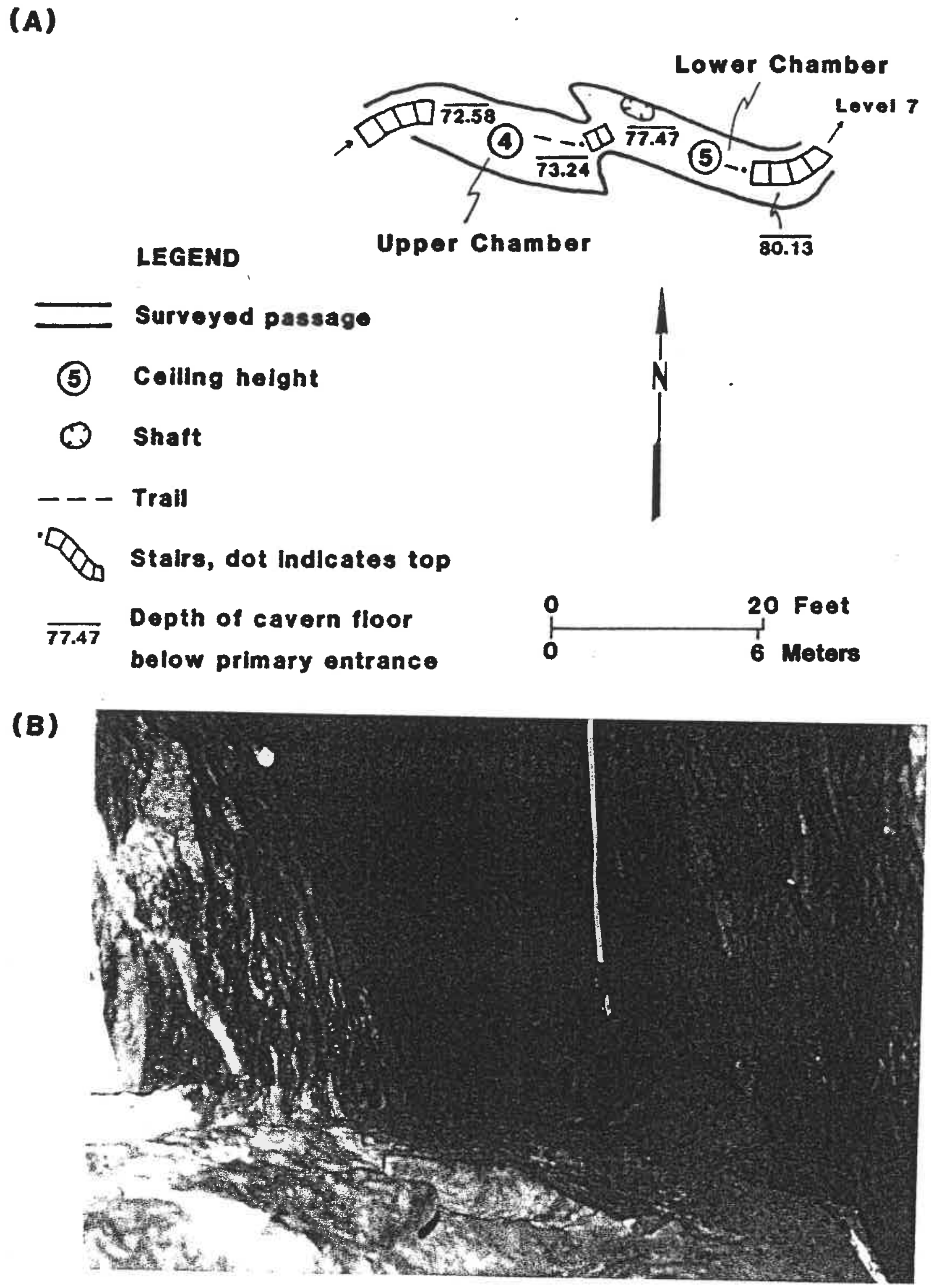


Figure 40. (A) Plan view of level six; (B) Photograph looking west through the lower chamber. Note water line running through shaft to "Old Mist'ry River" below.

(A)

(B)

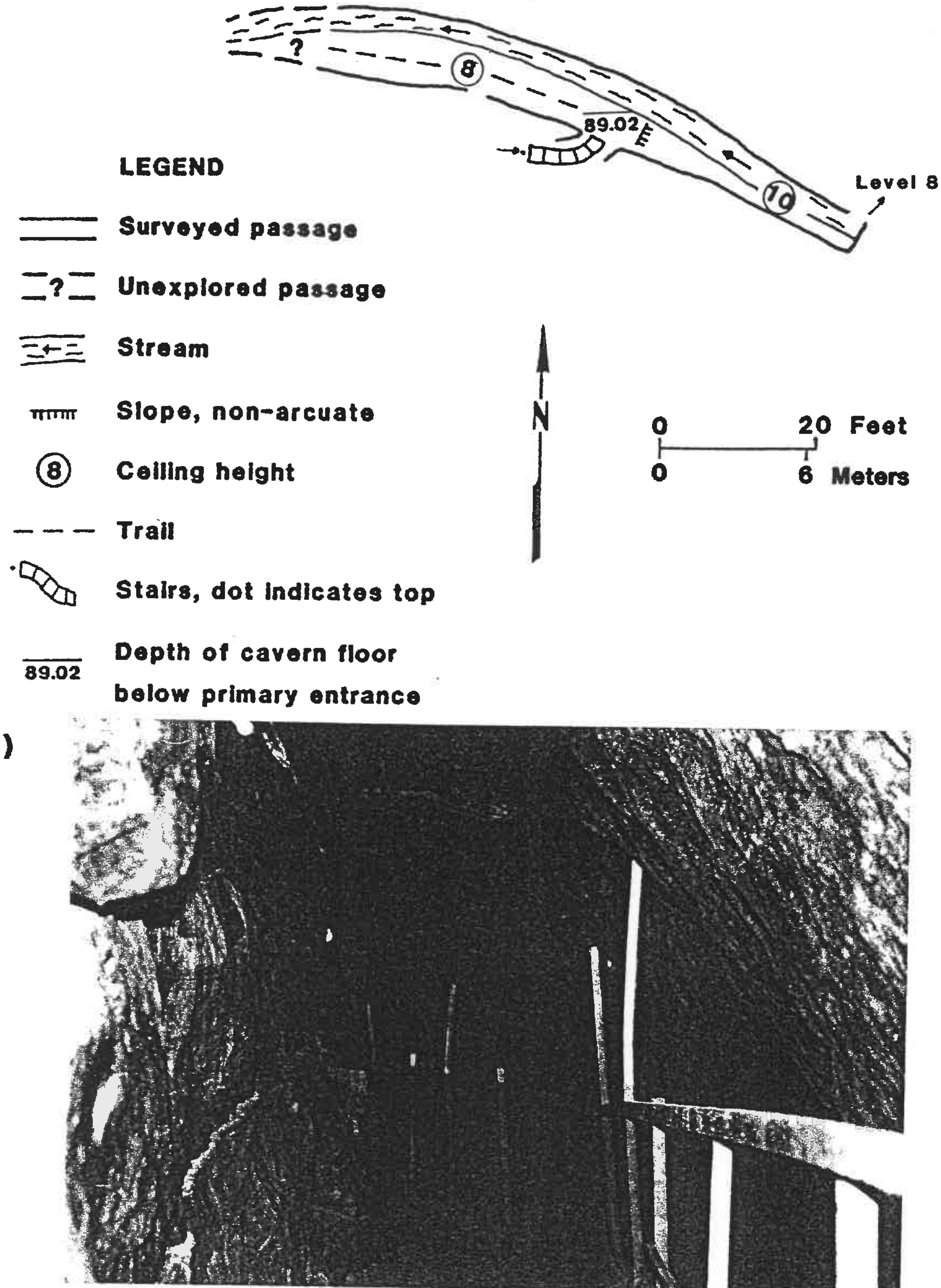


Figure 41. (A) Plan view of level seven; (B) Photograph looking west from the east end of the room. "Old Mist'ry River" is normally located in the crevice to the right of the railing.

east of the steps from the sixth level. A two foot (0.6 m) wide crevice is located between the floor and north wall. It is here that the water level is normally found. The crevice continues laterally from the western portion of the level for an unknown distance. The level becomes very narrow at this point which makes further surveying impossible. Access to the eighth level is gained through a hole in the eastern wall approximately eight feet (2 m) below the floor of the seventh level. Only during the summer of 1988 did the water level drop into the eighth level; however, the water obscured the view of much of this level.

The seventh level is comprised of light gray, massive dolostone. It contains numerous vugs which are filled with calcite crystals as well as portions of coral. A four-inch (10 cm) chert layer is visible near the ceiling while a one inch (2.5 cm) layer of carbonaceous shale is found almost even with the floor on the eastern portion of the level.

Stratigraphy

The area where Seneca Caverns is located is one of stratigraphic complexity with facies changes and the intertonguing of formations being common. This complexity is further confused by stratigraphic terminology of the members of the Columbus Formation. Thus, to better understand the area, the stratigraphy of north-central Ohio (Swartz, 1907) and central Ohio (Wells, 1947) will be discussed. The stratigraphy of Seneca Caverns is then compared to these regions.

Swartz (1907) recognized three subdivisions of the Columbus Formation in north-central Ohio. The upper member, known as the Venice member, is massive blue limestone with little chert and an intermittent fishbone bed at the top. Two fossil horizons below the fishbone bed were designated the "*Spirifer*" *duodenarius* horizon, and the upper *Paraspirifer acuminatus* horizon because of the abundance of these brachiopod species (Stewart, 1955). Wave markings are apparent at the base of this member.

The middle member, known as the Marblehead member, is gray fossiliferous limestone. Three spiriferid horizons, not all in one section, were distinguished by Swartz (1907): lower *Paraspirifer acuminatus* faunule (in north), *Brevispirifer gregarius* faunule, and "*Spirifer*" *macrothyris* faunule (in south). This member is equivalent to the lower two-thirds of the Delhi member in central Ohio (Stewart, 1955).

The lowermost member of the Columbus Formation, the Bellepoint member, is brown, dolomitic limestone. It is typified by many corals and has a one-inch (2.5 cm) sandy layer at the base. This member is only six feet (2 m) thick and rests on the Lucas Formation of the Detroit River Group. The total thickness of the Columbus Formation in north-central Ohio is 60 feet (18 m) (Stewart, 1955).

Wells (1947) performed a detailed analysis of the Columbus Formation in central Ohio and divided it into three parts or members. The uppermost member, known as the Delhi member, is a bluish-gray, relatively pure, often semicrystalline limestone with an average thickness of 56 feet (17 m). A fishbone bed, three to fifteen inches (7.6 to 38 cm) in thickness, is very well developed in the upper portion of the member, more so than in northern Ohio. Below this bone bed are three fossil horizons. These were designated the upper Paraspirifer acuminatus, Brevispirifer gregarius, and "Spirifer" macrothyris horizons, from top to bottom (Stewart, 1955).

The middle member, known as the Eversole member, is a rather pure limestone similar to the beds above, except for the gray to white chert occurring in somewhat alternate layers with the limestone. The chert is abundantly fossiliferous with the external markings of the original gastropod shells being beautifully preserved in the chalky shell replacement (Stewart, 1955). This layer is approximately six feet (2 m) in thickness in central Ohio but is absent in northern Ohio.

The lower member, known as the Bellepoint member, is a brown, dolomitic limestone approximately 43 feet (13 m) thick. This member is relatively unfossiliferous except for the upper portion which contains numerous corals and stromatoporoids. The bottom of the Bellepoint member is characterized by a one foot (0.3 m) layer of conglomerate which contains large and small water-worn pebbles of the Bass Islands (Silurian) rocks in a Columbus Limestone matrix (Stewart, 1955). It is important to note that the Detroit River Group (Devonian) is absent in central Ohio.

The stratigraphy of Seneca Caverns must now be compared to the stratigraphy of north-central and central Ohio. The upper member of the Columbus Limestone visible within Seneca Caverns is light gray, medium to massively bedded limestone averaging approximately 30 feet (9 m) in thickness. Levels one, two, and the upper portion of level three are composed of this member. A distinct horizon of the brachiopod *Brevispirifer gregarius* and *Mucrospirifer mucronatus* is found in the ceiling of level one and two, respectively. Other fossils located in this member include *Leiorhynchus* sp., *Schizophoria propinpua*, *Chonetes* sp., *Heliophyllum halli*, *Zaphrentis corniculum*, and *Tentaculites scalariformis*. This member correlates with the Marblehead member in north-central Ohio and the lower two-thirds of the Delhi member in central Ohio.

The middle member is analagous to the Eversole member of central Ohio. It is massive, light gray limestone which contains gray to white layers and nodules of chert. This member is 6 feet (2 m) thick and is visible in the lower half of level three and in the "Corkscrew Staircase".

The lowermost member, the Bellepoint member, is light brown, massive dolomitic limestone which contains two systems of well-developed joints trending N10E and N45E. It is relatively unfossiliferous, except for the upper few feet, which contain the coral *Zaphrentis corniculum*. The thickness of this member is approximately 25 feet (8 m) and is apparent in level four and the upper portion of level five. It correlates with the Bellepoint members of north-central and central Ohio. The sandy layer which marks the contact between the Columbus and Lucas formations as described by Swartz (1907), could not be located in the Caverns. The bedrock, however, abruptly changed colors from light brown to light gray and was determined to be highly dolomitic. The sixth and seventh levels contain appreciable amounts of chert and corals which are indicative of the Detroit River Group. The total thickness of the Columbus Limestone visible within Seneca Caverns is approximately 61 feet (19 m) which is almost identical to that given by Stewart (1955) for north-central Ohio.

Hydrogeology

Under normal circumstances, the behavior of ground water cannot be readily observed. Streams, however, often exist in the cave or cavern environment which are analogous to the water table. For this reason, caves and caverns are often used to directly study the behavior of ground water. Seneca

Caverns is just one such place where this phenomenon may be observed.

An underground stream, known as "Old Mist'ry River", is apparent in Seneca Caverns (Figure 42). It was found during much of the study in the seventh level approximately 100 feet (34 m) below the surface. The level of the stream is controlled primarily by the amount of runoff, mainly via dolines, reaching the stream. This was most evident on July 4, 1969, when the area received approximately 9 inches (23 cm) of rain in 4 hours. In two days, the water level rose to the Cavern entrance. It subsequently took four months for the water to return to its original level (Short, 1970). For comparison, in 1934 during a period of extended drought, the stream fell to a point approximately 100 feet (30 m) below its normal position, exposing levels 8, 9, 10, 11, and 12 (Bell, 1986).

To further define the effects of precipitation on the water level, an Omnidata International Easy Logger was installed in Seneca Caverns from July 10, 1986 to May 12, 1987. This instrument was equipped with a pressure transducer to measure the water level, a water temperature probe, and a tilt-bucket rain gauge which was placed in an open area near the cavern entrance. Readings were taken from each of these functions every two minutes. The water elevation and temperature readings were subsequently averaged

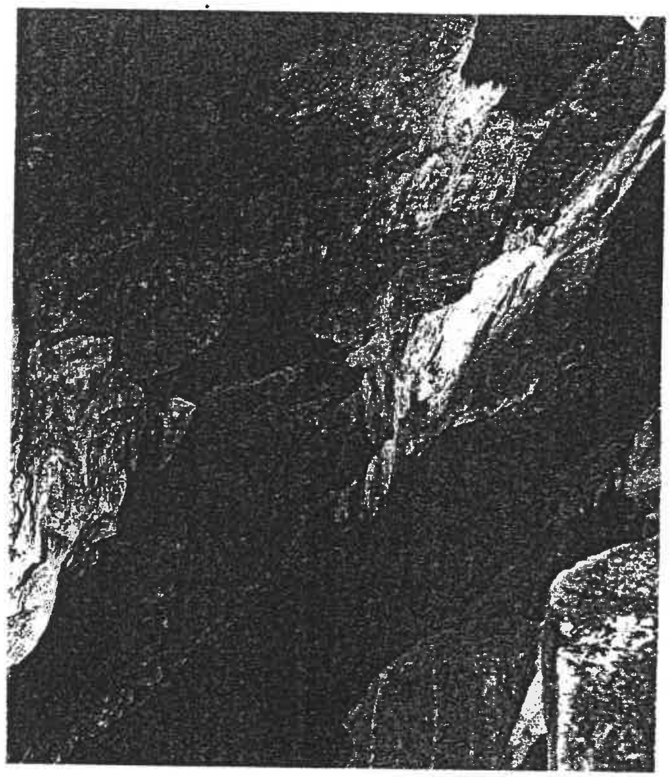


Figure 42. View of "Old Mist'ry River" in level seven of Seneca Caverns.

over a twenty minute period with the value for each being stored in a datapack. The same process was implemented for the precipitation except that the values were summed rather than averaged.

The resulting fluctuations in the cavern stream in response to precipitation events are graphically shown on a hydrograph (Figure 43). A typical hydrograph resulting from an isolated period of rainfall can be divided into three parts: the rising limb, crest segment, and falling limb or recession. Each respective segment may vary in shape in response to several factors. The rising limb begins at the

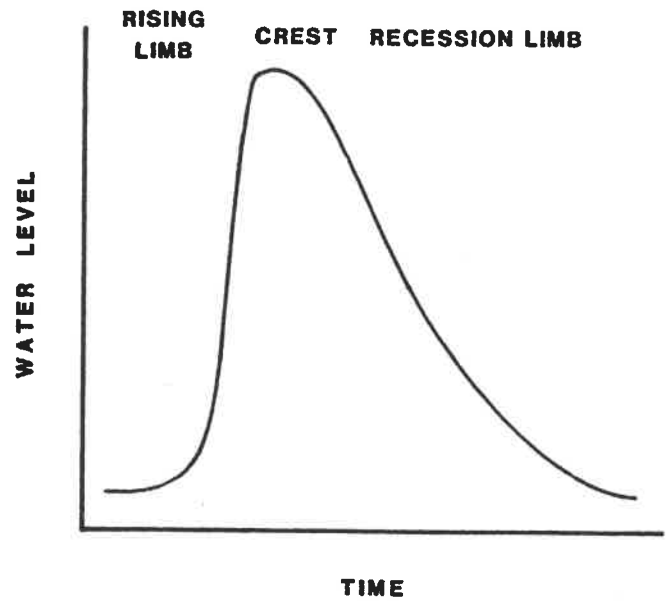


Figure 43. A typical stream hydrograph.

first recorded increase in water level and continues to the first inflection point. It represents an increase in storage within the aquifer. Its shape is usually convex upward and is influenced primarily by the travel time of water from different areas of the basin in response to storm events. In general, the travel time and the slope of the rising limb are inversely proportional. The crest segment of the hydrograph is located between the inflection on the rising limb and the equivalent on the recession limb. The crest represents the peak water level for the particular precipitation event. Oscillating rainfall intensity may cause two or more peaks from a single rainfall event. The recession limb begins at the first point of inflection on the falling limb and continues until the next resurgence. It is commonly assumed to mark the time at which surface inflow to the aquifer ceases and, thereafter, represents the withdrawal of water from storage within the aquifer. The recession limb is largely dependent on the physical features of the flow paths and flow system rather than the characteristics of the storm causing the rise. If the aquifer system is primarily of the diffuse-flow type, the recession limb will be fairly long and more gently sloping than if the system were of the conduit-flow type.

The response or lag time is a measure of the length of time required for the first storm pulse to reach the recording station from the midpoint, or centroid, of the precipitation event. As the definition implies, it is inherent in the determination of lag times that sharp pulses of water be produced. These are best provided by intense storms of short duration followed by periods of little or no rain through the recovery period. Such storm pulses should then produce a sharp ascension in the rising limb of the hydrograph. This, in turn, should be accompanied by a corresponding rise in the water temperature which is produced by a mixing of the warmer rainwater with the cool ground water.

Storm events from July, 1986 to May, 1987 which provided useful records are labeled A through E on Figure 44, while the data used for the construction of these graphs are in

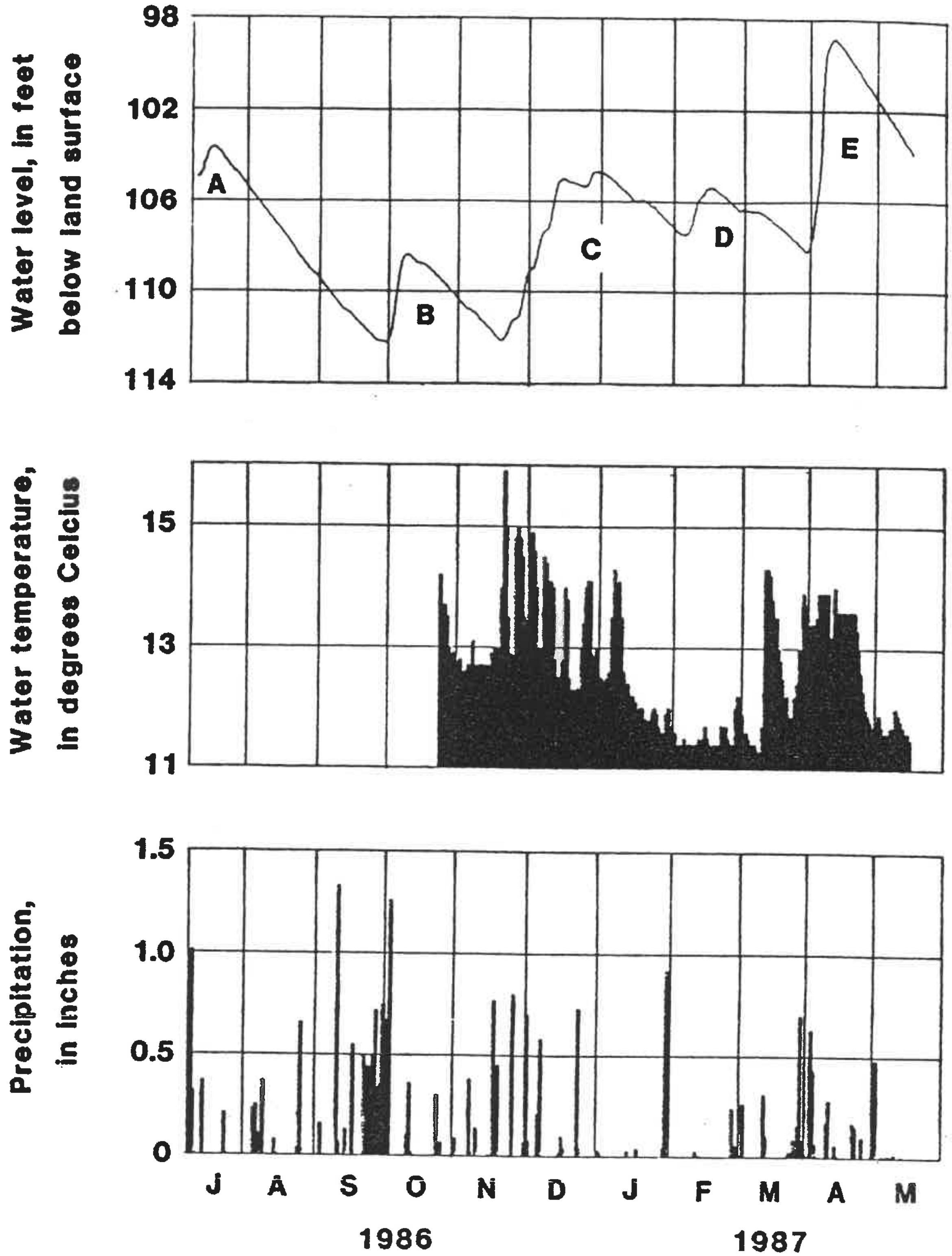


Figure 44. Fluctuations in water level and temperature of "Old Mist'ry River" in response to storm events at Seneca Caverns from July, 1986 to May, 1987.

Appendix C. The first record of rainfall from event A was obtained by the rain gauge at 1:50 A.M. on July 11, 1986, whereas the last record of precipitation was at 7:10 A.M. of the same day. The centroid for this event was at 4:10 A.M. The total amount of rainfall between these times was 0.88 inches (2.24 cm). The resulting rise in the Cavern stream began at 9:10 A.M. on July 11, 1986 and reached the crest at 6:10 P.M. on July 16, 1986, with the total increase in water level being 1.33 feet (0.41 m). This produced a lag time of 5 hours. The recession limb began at 8:10 P.M. on July 17, 1986 and continued downward at a rate of approximately 1.40 in/day (3.56×10^{-2} m/s) until 1:30 P.M. on August 30, 1986, with a total drop in water level of 9.30 feet (2.84 m). Several small deviations from the normal slope may be seen in the recession limb of event A. These are due to separate storm pulses. The water level, however, was not significantly affected because of the lengthy dry spells incurred between these aforementioned storm pulses. Thus, the bulk of the precipitation was utilized to satisfy the previously low soil moisture content.

Accurate precipitation records for event B were lost due to a malfunction of the rain gauge. Daily rainfall records were, however, obtained from the Bellevue Water Treatment Plant. The rainfall occurred sporadically over a period of one week prior to the initial rise in the cavern stream, which began at 1:51 P.M. on September 30, 1986. The total precipitation for this period was 2.52 inches (6.4 cm). The crest was attained at 5:11 P.M. on November 8, 1986, and began to recede at 7:11 A.M. the next day. The total rise in the cavern stream for event B was 3.99 feet (1.22 m). The lag time could not be accurately calculated because of the absence of detailed precipitation records from Seneca Caverns, along with the fact that the rise in water level could not be attributed to any one single rainfall event.

Event C is an example of a hydrograph with two crests. This was due to two separate storm events, the latter occurring during the recession limb produced by the former. Rainfall for the first portion of event C occurred between 5:16 A.M. on November 18, 1986, and 6:56 P.M. of the same day, with the total rainfall for the event being 0.77 inches (1.96 cm). The centroid for this event occurred at 2:46 P.M. The initial increase in the water level was recorded at 12:16 P.M. on November 19, 1986, while the crest was reached at 12:02 P.M. on December 14, 1986. This produced a lag time of 21 hours and 20 minutes, and a total water level increase of 7.16 feet (2.18 m). From Figure 44, it is apparent that the slope of the rising limb is irregular. This was in reaction to three storm pulses which caused an increase in the slope of the rising limb. This was also evident in the water temperature record which showed peaks in water temperature shortly after the rainfall events. The recession limb began at 1:42 A.M. on December 15, 1986, and continued until 6:42 A.M. on December 25, 1986. The water level began to rise at this point in response to 0.75 inches (1.91 cm) of rain which occurred between 3:02 P.M. on December 24, 1986 and 12:42 A.M. on December 25, 1986. The centroid for this event was at 9:02 P.M. The crest was achieved at 12:02 A.M. on December 29, 1986, and the recession limb began at 1:42 A.M. on December 30, 1986. The total water level increase was 0.68 feet (0.21 m), and the lag time was 9 hours and 40 minutes. The lag times for event C were longer than usual because the precipitation was distributed over a longer period of time.

The precipitation for event D was in the form of a wet snow. The first record of this snow was at 12:02 A.M. on January 30, 1987, and persisted until 1:02 P.M. the same day. The total precipitation from this event was 0.91 inches (2.31 cm) with the centroid being at 6:12 P.M. The Cavern stream began to rise at 9:22 P.M. on February 4, 1987, and peaked at 12:48 P.M. on February 14, 1987, thus yielding a total water level increase of 2.10 feet (0.64 m). The recession limb began at 7:28 P.M. on February 16, 1987. The lag time for event D was an unusually long 5 days, 15 hours, and 10 minutes. This was due, once again, to the fact that the snow fell gradually over a 13 hour period. Thus, most of the snowmelt probably percolated slowly downward through the soil rather than entering the ground-water system rapidly via dolines. In addition, the lack of appreciable precipitation prior to this may explain the minimal rise in the water level.

Event E represented the final recorded storm pulse. The first record of rainfall was at 9:48 P.M. on March 29, 1987, and lasted until 7:08 A.M. the next day. The total amount of rainfall in this time was 0.84 inches (2.13 cm) with the centroid occurring at 2:48 A.M. The resulting rise in water level began at 2:48 P.M. on March 30, 1987. The level rose a total of 9.43 feet (2.88 m) at a rate of approximately 1.00 ft/day (0.30 m/s). The crest was attained at 4:48 A.M. on April 11, 1987, whereas the recession limb began at 2:28 P.M. the same day. The lag time for this event was 12 hours. The relatively large water level increase for event E was the result of snowmelt which was triggered by an increase in air temperature and the previously mentioned rainstorm. In addition, several storm pulses occurred during the rising limb which contributed supplemental amounts of runoff. This is well documented by the peaks in the water temperature record. The lag time was fairly short because the ground was still frozen, thus promoting quicker runoff.

The elevation of the stream level over the course of the observation period showed a good seasonal correlation. The water level began to recede at a fairly steady rate during the summer months, although several storm events did cause slight rises in the elevation (event A).. The lowest water elevations were recorded in late September and mid-November. The water level would normally have been much lower in November and December, but the precipitation was well above normal for September and November, which led to a dramatic rise in water elevation (events B and C). The level of the cavern stream then remained fairly constant during the winter months, being affected only slightly during event D. The water level, however, reached its maximum in early spring after the period of snowmelt (event E). This provided a maximum seasonal variation in the stream level of 13.43 feet (4.09 m).

The cavern stream hydrograph was very useful in determining the flow types operating within the carbonate aquifer. The relatively short lag times and the steepness of the rising limbs indicate that the surface runoff was quickly transmitted into the ground-water regime, which in this case was through dolines. Thus, conduit-type flow recharges the aquifer. The low to moderate slopes and the length of the recession limbs, however, illustrate that the aquifer acts to store water more readily than transmit it. This is indicative of the diffuse flow nature of the aquifer. This relates well with what was discussed previously about the dual nature of flow within the carbonate aquifer.

The geographic location of a storm exerts a great influence on where and to what degree the water table will be affected. This is best illustrated by the floods of 1937 and 1969. In 1937, the bulk of the precipitation was received well south of the Seneca Caverns, whereas with the

1969 flood, rainfall was heaviest at the Caverns and in Bellevue. The degree of flooding at these two locations, however, was much more severe in 1937 than in 1969. During the flood of 1937, the water level within Seneca Caverns began to rise sometime after the storm event had ended. Water completely filled the Caverns and issued forth onto the surface as a large stream. Fields south of Bellevue were soon turned into large lakes. A short time after this took place, four foot (1 m) fountains sprung forth from the storm sewers in Bellevue, causing mass flooding of the town (Bell, 1988). Similar events were noticed during the 1969 flood, but they were observed to the north of Bellevue and in Castalia. These two contrasting flood events illustrate the importance of the location of the storms and that in flood conditions, the aquifer acts to transmit large amounts of water in waves or pulses.

A detailed geochemical analysis of the cavern stream was performed by the USGS, Department of Interior (1972), and the results are in Table 11. These results were very similar to those obtained for the Castalia Blue Hole (see Table 8), which may indicate a connection between the two. Numerous stories exist stating that bottles placed in Seneca Caverns appeared years later in the Castalia Blue Hole (Bell, 1986). These reports, however, could not be verified. Thus, to prove this theory, a dye trace originating from Seneca Caverns had been planned for the spring or sum-

Table 11. Geochemical parameters of "Old Mist'ry River", Seneca Caverns (after USGS, Dept. of Interior, 1972).

Parameter	Reading (mg/l)
Sodium (Na)	8.5
Bicarbonate (HCO_3)	260
Sulfate (SO_4)	990
Chloride (Cl)	16
Total Dissolved Solids (calculated)	1,620
Hardness as $CaCO_3$	1,300
Specific Conductance (micromhos at 20°C)	1,850
pH	7.2

mer of 1988. The optimal time for such an experiment is after a major precipitation event when ground-water levels are high or just beginning to recede. Low ground-water levels and the lack of significant rainfall events during the drought of 1988, however, precluded such an experiment.

Water from the cavern stream is quite pure except for the high degree of mineralization. After periods of heavy rain, however, the Cavern stream becomes cloudy, bubbles appear at the surface, and the air possesses a pungent odor similar to that of rotting eggs. The cloudiness of the water is the result of sediments washed in from the surface, whereas the bubbles are from a chemical reaction between surface runoff and the ground water. The smell is caused by hydrogen sulfide given off as a product of this reaction.

There is only one known group of organisms known to inhabit the cavern stream. They are called Amphipods and are found mostly in clear cool waters of springs, ponds, lakes, and pools. Many species are, however, restricted to subterranean waters such as the Cavern stream ("Old Mist'ry River"). The Amphipods found here are of the genus Crangonyx sp. and are approximately one-half inch (1.3 cm) in length, are blind and exhibit a clear body (Page, 1968). They most commonly feed on detritus and algae washed in from the land, as well as bacterial scum floating on the water surface.

SUMMARY AND CONCLUSIONS

Karst development in Ohio is fairly widespread throughout the western half of the state, as is evidenced by the location of numerous caves and caverns in the region (G.W. White, 1926 and Hobbs, in progress). Detailed karst hydrogeologic studies of this region are, however, seriously lacking. To remedy this situation, an investigation was initiated to determine the relationships between the karst geomorphology and hydrogeologic regime of the Bellevue-Castalia area in north-central Ohio (see Figure 1). The geologic framework of Seneca Caverns was also examined along with the hydrogeology of the Cavern stream ("Old Mist'ry River").

The areas major karst landforms -- dolines (sinkholes), sinking streams, ponors, caves and caverns, and springs -- were located and classified based on their origin and physical properties. The fundamental component of karst topography, the doline, is a near-circular or elongate, funnel-shaped depression resulting from the solution or collapse of underlying strata. Five major classes of dolines are recognized in the literature (see Figure 8), of which four were mapped in the Bellevue-Castalia area (see Plate 1). The most common type is the subsidence doline. This type was formed by the subsidence of the surficial deposits (glacio-lacustrine tills) into solutionally-widened joints in the carbonate bedrock beneath. A typical subsidence doline is shown in Figure 9. The alluvial streamsink doline is generally analagous to a ponor, which represents the point where a stream sinks through alluvium into the underlying karst bedrock. Such dolines in the Bellevue-Castalia area were found in the channels of Speck Creek and Schneider's Creek (see Figures 12 and 13), both of which are sinking streams. The two most contrasting types are the funnel-shaped solution doline and the steep or cliffed collapse doline. The solution doline is located where surface solution is concentrated, which is often where two or more joints intersect. The collapse doline is normally caused by the collapse of the roof of a cave formed by underground solution. Only one example of each, however, was identified in the study area (see Figures 10 and 11).

Dolines in the Bellevue-Castalia area were confined almost exclusively to the surficial extent of the Columbus Formation and were arranged in a fairly straight line trending N45E due to joint control. These dolines also form because the purity of the limestone (94 percent calcium carbonate) allows for easier solution by percolating waters.

Aside from Seneca Caverns, which will be discussed later, only a few small caves are found in the study area. Crystal

Rock and Brewery Caves are located just south of the town of Crystal Rock (see Plate 1) in the Put-in-Bay Formation (Silurian). Several small caves have also been found on the northeastern rim of the collapse doline mentioned previously.

The 700 foot (213 m) Silurian-Devonian carbonate sequence represents the regional aquifer system of the Bellevue-Castalia area. This aquifer exhibits characteristics of a confined, semi-confined, and unconfined aquifer. In Townsend Township and west-central Margaretta Township, where the confining layer of glacio-lacustrine till is thicker than 60 feet (18 m) (see Figure 6), the aquifer is confined. The location of this area is analagous to the zone of flowing artesian wells (see Plate 2). In the southern two-thirds of the study area where the confining layer is very thin or absent altogether, the aquifer is unconfined. The aquifer is thought to be semi-confined for some distance between the confined and unconfined areas.

Plate 2 represents the carbonate aquifer potentiometric surface for the area. Over 100 bedrock wells were measured in October 1988 with a Roctest Model CPR6 water level indicator. Ground-water flow direction in the study area is controlled primarily by the bedrock topography of the region. Numerous grooves or troughs have been developed in the bedrock (see Figure 4) possibly as a result of erosion by the preglacial Erigan River system described by Spencer (1894). The large bedrock trough visible just west of Flat Rock and Bellevue and extending northeast and then northwest toward Sandusky Bay coincides nicely with the ground-water low seen on Plate 2. This ground-water low undoubtedly represents the trunk of the system, while smaller deviations in the water table are the result of bedrock lows produced by tributaries of the main river. The overall hydraulic gradient for the study area is 10 feet per mile (1.9 m/km).

In the study area, two types of recharge to the carbonate aquifer are prevalent depending on the precipitation rate. During moderate to heavy precipitation, when runoff rates are high, recharge takes place directly through dolines and sinking streams. This is known as concentrated autogenic recharge (Gunn, 1985). When, however, the precipitation rate is low, recharge occurs directly where bare limestone is exposed at the surface or indirectly through the cover of surficial deposits. This type is known as diffuse autogenic recharge.

The bulk of ground-water discharge in karst terranes takes place through springs. Ten major springs were located in the study area (see Plate 1). In addition, several springs located in the floor of Sandusky Bay were noted. Each spring was subsequently classified as a contact, sinkhole, fracture, or subaqueous-type spring based on its location and physical attributes (see Figure 14).

A contact spring, as the name implies, is found along the lithologic contact between an overlying permeable rock unit and one of less permeable rock. Springs of this type are Duck Pond Spring, Castalia "Blue Hole", Railroad Spring, Castalia Trout Farm Spring, and Rockwell Spring (see Figures 15 to 21). These are all located at or very near the contact between the highly permeable Columbus Formation and the underlying, less permeable Detroit River Group. Sinkhole springs are found where a cavern is connected to a shaft that rises to the surface. A specialized type of sinkhole spring was identified exclusively in collapse-type dolines produced by the solution of gypsum by artesian waters in the underlying Bass Islands Group. Springs of this type are Miller's Blue Hole, Crane Spring, and Muddy Spring (see Figures 22 to 24). Fracture springs exist where jointed zones present in less permeable rock intersect the land surface at low elevations. Representative springs are the Rainbow Creek Spring and Ransom Spring (see Figures 25 and 26). Subaqueous springs are found where the spring outlet is located below the water level of a lake or sea. Several of these springs have been reported in the floor of Sandusky Bay (Norrocky, 1987). However, no attempt was made to pinpoint the exact locations of these springs.

Discharge measurements were taken for each spring during high and low flow periods occurring in June and October, respectively (see Table 7). The total discharge of all the springs in June was 22,183 gpm (1.26 m^3/d) as compared to 19,971 gpm (1.40 m^3/d) for October. This yielded a coefficient of variation (CV) of 7.03 percent, indicating a low discharge variability.

Chemical analyses of Castalia "Blue Hole" and Miller's Blue Hole were performed by the USGS, Department of the Interior in 1968 and 1972, respectively (see Table 8). Sulfate concentration, TDS, hardness as $CaCO_3$, and specific conductance are much greater at Miller's Blue Hole than at Castalia "Blue Hole". Based on this information, along with specific conductance values collected by the author from all the springs (see Table 9), it would appear that Rainbow Creek Spring, Ransom Spring, Miller's Blue Hole, Crane Spring, and Muddy Spring are part of a flow system separate from Duck Pond Spring, Castalia "Blue Hole", Railroad Spring, Castalia Trout Farm Spring, and Rockwell Spring. Definitive dye results, however, need to be obtained to substantiate this theory.

Ground-water flow within a carbonate aquifer may be classified as either diffuse flow, conduit flow, or any combination thereof. Diffuse flow in aquifers takes place along joints, fractures, and bedding planes, while that of conduit aquifers occurs through an integrated network of solutionally-enlarged conduits (see Figure 27).

The carbonate aquifer underlying the Bellevue-Castalia area was determined to be primarily of the diffuse-flow type. The basic lithology of the aquifer, excluding the Columbus and Delaware formations, is a coarsely crystalline dolomite that has undergone very little solutional modification. A definite water table is present (see Plate 2), and the topographic relief and hydraulic gradients are relatively low. Discharge in the region takes place through numerous springs and small seeps. Several of the larger springs in the Castalia area -- Duck Pond Spring, Castalia "Blue Hole", Castalia Trout Farm Spring, and Rockwell Spring -- are the result of a single stratigraphic feature -- that being the difference in permeabilities of the Columbus and Lucas formations. Ground- water flow through the carbonate aquifer is relatively deep and is considered to be of the regional flow-type (Toth 1962 and 1963). This is illustrated by the springs' constant yearly discharge (see Table 7), the elevated TDS and hardness values (see Table 8), and the consistent temperature and specific conductance readings (see Table 9). In addition, the springs exhibited no signs of turbidity after storm events.

The only evidence for conduit flow in the Bellevue-Castalia area is the abundance of dolines and sinking streams. These features are developed almost exclusively in the highly soluble Columbus Limestone and represent points of concentrated recharge to the ground-water regime, which is typical of conduit flow.

Diffuse ground-water flow in the Bellevue-Castalia area also occurs through hundreds of tiny openings referred to as joints or fractures. Their abundance, distribution, and orientation is controlled primarily by the geologic structure of the area. The exact locations of such features, however, may be determined through photogeologic investigations, which entail the mapping and interpretation of natural linear features on aerial photographs. A fracture trace and lineament are natural linear features consisting of topographic, vegetation, or soil tonal alignments visible on aerial photographs and expressed continuously for one mile (1.6 km) and greater than one mile (1.6 km), respectively.

Low altitude (scale 1:24,000), black and white, aerial photographs were used to locate all linear features in the study area in an effort to determine regional ground-water flow direction(s). The end result was a near orthoganal fracture pattern with the dominant trend being northeast-southwest and a less dominant northwest-southeast trend (see Figure 29). Joint orientations from four quarries -- Flat Rock, Belleve, Parkertown, and Castalia -- in the study area (see Table 10) also yielded this same pattern (see Plate 3). This configuration was the result of extensional stress generated in the Findlay Arch region as a result of differential subsidence of the adjoining Appalachian and Michigan basins (Armstrong, 1976).

Two dye tests were conducted during the spring and summer of 1987 to further define the direction and determine the rate of ground-water flow through the carbonate aquifer. The initial dye trace commenced on May 5, 1987. Five pounds of Rhodamine WT (CI Acid Red 388) were injected into a doline which was the sinking point of Speck Creek. This location was chosen in consultation with Dr. James F. Quinlan because of its close proximity to the recovery points (springs) and the presence of water to effectively carry the dye into the subsurface. Also, the ground-water flow direction taken from the potentiometric map (see Plate 2) was to the north. The estimated discharge of the creek was between five and ten gallons per minute (3.1×10^{-4} to 6.3×10^{-4} m^3/s). The ten springs discussed previously, as well as the stream found within Crystal Rock Cave were chosen as monitoring sites (see Plate 1). Activated charcoal packets were used to collect the dye and were analyzed in accordance with the procedure described by Quinlan (1986). Water samples were also taken and analyzed on a Turner Model 111 filter fluorometer. After two months (May-June, 1986) of sampling the sites twice weekly, none of the dye was recovered. At this point, the sampling period was reduced to once a week for a period of another two months (July-August, 1986). No dye was ever recovered, so a second trace was proposed for the same injection point with a larger concentration of a different dye.

Twenty pounds of Sodium Fluorescein (CI Acid Yellow 73) were injected on August 25, 1987. Sodium Fluorescein was chosen because it could be analyzed on the fluorometer along with the possible presence of the Rhodamine WT dye from the previous test, with no threat of interence between these two dyes. An artificial water supply had to be used for the injection because the flow of Speck Creek had ceased. To remedy this situation, a pumper truck, provided by the Bellevue Fire Department, supplied the necessary water. A "primer" consisting of 700 gallons (2,652 l) of clean water was pumped into the doline. The dye was then mixed with 1,400 gallons (5,303 l) of water and injected at a rate of 75 to 100 gpm (4.7×10^{-3} to 6.3×10^{-3} m^3/s). This injection was followed by a "chaser" of 700 gallons (2,652 l) of water. The same monitoring schedule was employed as during the first test. After nine months of sampling, however, none of the dye was discovered.

The unexpected disappearance of the dyes may have been the result of one or more of the following:

(1) the amount of water present within the aquifer was underestimated, thus the dye was diluted beyond detection before it reached the monitoring sites;

(2) assuming the carbonate aquifer was of the diffuse flow type, it may be acting to store water rather than transmit it with any great velocity;

(3) the doline where the dye was injected may be feeding a local flow system operational at the till-bedrock interface, which in turn may not be discharging at any of the monitoring sites; and/or

(4) the dye may be moving along a deeper flow path and may be discharging at some other point(s), possibly through subaqueous springs located in the floor of Lake Erie.

Seneca Caverns was first discovered in 1872 by two area youths. In 1872, the upper levels were opened under the name Good's Cave, however, this venture lasted only three years. In 1928, Don and Fannie Mae Bell acquired a lease on the property and proceeded to clean out the cave. Eventually a passageway was discovered leading down to an underground stream. On May 14, 1933, the cave was reopened under the name Seneca Caverns. Richard Bell purchased the land from his parents in 1964 and continues to commercially operate the Cavern.

Seneca Caverns is a collapse or breakdown-type cavern occurring along a major fracture in the Columbus and Lucas formations and possibly the underlying strata. The southern wall or block has been downthrown approximately eight feet (2 m) and is highly shattered and jumbled. There are places where the floor appears to fit into the ceiling much like a jigsaw puzzle (see Figure 35). The collapse was caused initially by the conversion of anhydrite to gypsum in the underlying Bass Islands Group. This may have led to a 33 to 62 percent increase in volume, which would be instrumental in deforming stratified rock. The gypsum was then preferentially dissolved by ground water moving vertically through joints and fractures, and laterally through bedding planes.

The fracture containing Seneca Caverns trends N68W for at least 2 miles (3.2 km) and has an average overall dip of 40NE (see Figure 31). The total vertical extent of the fracture has yet to be determined because of the presence of water in the lower reaches of the Caverns. Seven distinct rooms or levels were visible during this study with water commonly encountered in the seventh level (see Figure 31). Because of the drought experienced by the midwestern states during the spring and summer of 1988, the water receded exposing a portion of the eighth level. The lowest water level, however, was observed in 1937. During this period, the water dropped an estimated 100 feet (30 m) below the seventh level exposing an additional five levels (Bell, 1986). The fracture, however, still appeared to continue downward below this point. The shape of the levels is generally rectangular in shape due to the joint pattern exhibited by the strata.

The seven levels of Seneca Caverns were surveyed on September 18, 1985 with a Wild T1A theodolite. This yielded the dimensions and elevations of each level from which a cross-section could be constructed (see Figure 31). Subsequent maps were drawn illustrating the unique features of each level (see Figures 33 to 36 and 39 to 41).

The stratigraphy of Seneca Caverns consists of a 61 foot (19 m) section of Columbus Formation underlain by an undetermined thickness of Lucas Formation. The Columbus Formation was subdivided into three members based on analyses of the formation in north-central Ohio (Swartz, 1907) and central Ohio (Wells, 1947). The upper member is medium to massively bedded limestone averaging 30 feet (9 m) in thickness. The first, second, and upper portion of level three are composed of this member. A distinct horizon of the brachiopod Brevispirifer gregarius and Mucrospirifer mucronatus is visible in the ceiling of level one and two, respectively. Other fossils include Leiorhynchus sp., Schizophoria propinqua, Chonetes sp., Heliophyllum halli, Zaphrentis corniculum, and Tentaculites scalariformis. This member correlates with the Marblehead member in north-central Ohio and the lower two-thirds of the Delhi member in central Ohio. The middle member is massive, light gray limestone with gray to white layers and nodules of chert. It is 6 feet (2 m) thick and is visible in the lower half of level three and in the "Corkscrew Staircase". This member is analgous to the Eversole member of central Ohio which is absent in north-central Ohio. The lowermost member of the Columbus Formation, the Bellepoint member, is brown, dolomitic limestone which contains two systems of well-developed joints trending N10E and N45E. It is relatively unfossiliferous except for the upper few feet which contain the coral Zaphrentis corniculum. The thickness of this member is approximately 25 feet (8 m) and is apparent in level four and the upper portion of level five. It correlates with the Bellepoint members of north-central and central Ohio. The Lucas Formation is visible in levels five, six, and seven. The bedrock is light gray dolostone with appreciable amounts of chert and corals apparent in the sixth and seventh levels.

An underground stream, known as "Old Mist'ry River", is found for much of the year in the seventh level (see Figure 42). The level of the stream is controlled primarily by the amount of runoff, mainly via dolines, reaching the stream. This was most evident on July 4, 1969, when the area received approximately 9 inches (23 cm) of rain in 4 hours. In two days, the water level rose to the Cavern entrance. It subsequently took four months for the water to return to its original level (Short, 1970). For comparison, in 1934 during a period of extended drought, the stream fell to a point approximately 100 feet (30 m) below its normal position, exposing levels 8, 9, 10, 11, and 12 (Bell, 1986). An Omnidata International Easy Logger was installed in Seneca Caverns from July 10, 1986 to May 12, 1987 to determine the lag time from precipitation events to resulting rises in the water level. Five storm events were chosen for the calculation of lag times during this period (see Figure 44). These storm events yielded lag times ranging from five hours to five days, fifteen hours, and ten minutes. Shorter lag times were recorded when the storm duration was shorter and the soil was moderately saturated, and vice versa.

The cavern stream hydrograph (see Figure 44) was very useful in determining the flow types operating within the carbonate aquifer (see Figure 44). The short lag times and the steepness of the rising limbs indicate that the surface runoff was quickly transmitted to the ground-water regime through dolines and sinking streams. Thus, conduit flow is the dominant factor for recharge to the aquifer. The low to moderate slope and length of the recession limbs, however, illustrates that the aquifer acts to store water more readily than transmit it. This is indicative of the diffuse nature of the aquifer.

A detailed geochemical analysis of "Old Mist'ry River" was performed by the USGS, Deptartment of Interior (1972) (see Table 11). These results were very similar to those obtained for Castalia "Blue Hole" (see Table 8), which may indicate a connection between the two. To prove this theory, a dye trace originating from Seneca Caverns had been planned for the spring or summer of 1988. Low ground-water levels and the lack of significant rainfall events during the drought of 1988, however, precluded such an experiment.

RECOMMENDATIONS FOR FURTHER STUDY

The primary purpose of this study was to stimulate interest in a relatively untouched area in terms of hydrogeologic research. From this more regional study, numerous topics could be further developed. These include:

1. a study of the relation of spring discharge behavior to the hydrologic properties of the carbonate aquifer. This would be accomplished by the installation of discharge recorders in one or two of the larger springs in the area for one year. The shape of the hydrograph is instrumental in the determination of an aquifers' hydrologic properties and the regional hydrogeologic setting. The recession limbs of transient discharge events may be used to characterize the geometry and spatial distribution of porosity in the aquifer. This type of study is described in detail by Gaither (1977).

2. a study of the seasonal fluctuations in the chemistry of selected springs in the study area. Such chemical parameters as specific conductivity, temperature, total dissolved solids, and hardness are most commonly used for characterizing carbonate aquifers (Shuster and White, 1971, and Jacobson and Langmuir, 1974).

3. numerous dye traces from select dolines and/or sinking streams, and also from Seneca Caverns, during high and low flow conditions. Data from such tests could prove useful in delineating ground-water basin boundaries and determining flow-through times. within the carbonate aquifer.

4. a geochemical analysis of select springs located in the area. Such a study could be directed primarily to the seasonal changes in concentrations of nitrates, chloride, and sulfate as is described by Kastrinos and White (1986). Other parameters could also be tested for as may be deemed necessary.

5. a geophysical investigation of the Erigan River valley. Numerous gravity and seismic refraction surveys could be conducted in the study area to further define the dimensions and location of this buried valley. These surveys may also be used to locate some structure, possibly a major fracture or fault, that is responsible for the location and orientation of this valley.

It is hoped that this investigation will aid the public in better understanding the geology and hydrogeology of the area. The carbonate aquifer system may represent an almost inexhaustible supply of fresh water if it is managed correctly. Contamination of the aquifer during the mid 1900's, however, is just one example of mismanagement of this system. This was caused primarily by a lack of appreciable knowledge of the geology and hydrogeology of the area. It is the sincere hope of the author that this study will aid in the better understanding of the karst hydrogeology of the area so that such a situation will never happen again.

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Appendix A

WELLS DATA USED FOR THE CONSTRUCTION OF A POTENTIOMETRIC SURFACE MAP FOR THE STUDY AREA.

SENECA COUNTY

Thompson Township

LOCATION	SURFACE ELEVATION	STATIC WATER LEVEL
Section 1,T3N,R17W	779	677
Section 2	780	660
Section 2	770	650
Section 3	775	664
Section 4	800	739
Section 9	805	723
Section 12	800	700
Section 12	806	797
Section 13	810	788
Section 13	809	750
Section 15	802	716
Section 15	794	665
Section 16	806	722
Section 17	800	778
Section 17	802	756
Section 18	819	791
Section 18	811	795
Section 19	830	790
Section 21	807	723
Section 21	810	738
Section 22	815	741
Section 23	805	722
Section 24	814	777
Section 24	808	777
Section 25	829	738
Section 27	825	742
Section 27	836	761
Section 28	838	763
Section 29	825	785
Section 30	833	799
Section 30	839	796
Section 30	835	796

HURON COUNTY

Lyme Township

LOCATION	SURFACE ELEVATION	STATIC WATER LEVEL
Section 3, T4N,R24W	762	720
Section 3	760	713
Section 3	738	703
Section 3	740	652
Section 3	771	749
Section 4	807	797
Section 4	800	789
Section 4	789	747
Section 4	782	760

SANDUSKY COUNTY

York Township

LOCATION	SURFACE ELEVATION	STATIC WATER LEVEL
Section 1,T4N, R17E	736	645
Section 2	699	640
Section 4	673	620
Section 4	701	646
Section 4	685	648
Section 5	663	644
Section 10	730	682
Section 11	734	643
Section 11	727	644
Section 13	736	652
Section 15	739	657
Section 16	738	685
Section 16	730	678
Section 17	735	684
Section 19	735	693
Section 20	747	730
Section 21	754	711
Section 23	749	655
Section 24	737	656
Section 25	760	655
Section 26	765	657
Section 27	778	732
Section 28	779	730
Section 28	770	714
Section 29	782	734
Section 30	750	712
Section 31	782	744
Section 32	784	762

York Township

LOCATION	SURFACE ELEVATION	STATIC WATER LEVEL
Section 32	784	762
Section 33	791	735
Section 34	780	724

Townsend Township

Section 22	619	616
Section 24	662	640
Section 25	700	638
Section 26	644	626
Section 26	675	626
Section 27	645	622
Section 29	628	619
Section 34	679	624

ERIE COUNTY

Groton Township

Section 2,T5N,R24W	723	706
Section 2	716	699
Section 2	723	703
Section 3	711	644
Section 3	706	637
Section 3	707	677
Section 3	722	696
Section 4	721	644
Section 4	731	647
Section 4	723	647
Section 4	713	646
Section 4	726	691

Margaretta Township

Section 1,T6N,R24W	700	655
Section 1	697	669
Section 1	701	680
Section 2	627	610
Section 2	613	600
Section 2	595	585
Section 2	605	590
Section 2	627	590
Section 2	578	573
Section 2	576	573
Section 4	716	633
Section 4	685	634
Section 4	714	630

Appendix B

JOINT ORIENTATIONS FROM SELECT QUARRIES IN THE STUDY AREA.

France Stone Company - Flat Rock Quarry

Joint Number	Strike	Dip
1	N 9 E	Vertical
2	N 25 E	85 SE
3	N 49 E	80 NW
4	N 43 E	Vertical
5	N 1 W	Vertical
6	N 2 E	Vertical
7	N 49 E	Vertical
8	N 40 E	Vertical
9	N 11 E	Vertical
10	N 12 E	Vertical
11	N 19 E	Vertical
12	N 17 E	Vertical
13	N 22 E	Vertical
14	N 9 E	Vertical
15	N 49 W	Vertical
16	N 20 W	Vertical
17	N 41 E	83 NW
18	N 60 E	Vertical
19	N 42 E	85 SE
20	N 32 E	Vertical
21	N 10 W	Vertical
22	N 52 E	Vertical
23	N 52 W	85 SW
24	N 70 E	Vertical
25	N 55 W	Vertical

France Stone Company - Bellevue Quarry

Joint Number	Strike	Dip
1	N 1 W	Vertical
2	N 48 E	Vertical
3	N 32 E	Vertical
4	N 55 E	80 SE
5	N 69 W	Vertical
6	N 6 W	Vertical
7	N 12 E	Vertical
8	N 41 W	32 NE
9	N 19 E	Vertical
10	N 49 E	Vertical
11	N 16 E	Vertical
12	N 20 E	Vertical
13	N 35 E	82 SE
14	N 34 W	Vertical
15	N 40 W	Vertical
16	N 31 W	68 SW
17	N 48 E	84 SE
18	N 33 W	69 NE
19	N 17 E	71 SE
20	N 14 W	67 NE
21	N 15 E	78 SE
22	N 46 E	76 NW
23	N 40 E	76 NW
24	N 47 E	80 NW
25	N 46 E	86 NW

Sandusky Crushed Stone Company - Parkertown Quarry

Joint Number	Strike	Dip
1	N 38 E	65 NW
2	N 38 W	88 SW
3	N 26 W	76 NW
4	N 35 E	Vertical
5	N 33 W	89 SW
6	N 47 E	69 NW
7	N 30 E	Vertical
8	N 41 E	Vertical
9	N 34 E	63 NW
10	N 39 E	89 NW
11	N 30 W	Vertical
12	N 38 E	81 NW
13	N 59 E	56 NW
14	N 45 E	63 NW
15	N 54 E	68 NW
16	N 54 W	88 SW
17	N 59 E	87 NW
18	N 44 E	Vertical
19	N 30 E	80 NW
20	N 55 E	88 NW
21	N 54 E	81 NW
22	N 63 W	85 NE
23	N 77 W	86 NE
24	N 71 E	72 NW
25	N 68 E	77 NW

Erie County Metroparks (Wagner Stone Company) - Castalia Quarry

Joint Number	Strike	Dip
1	N 42 E	82 NW
2	N 52 W	Vertical
3	N 57 E	Vertical
4	N 26 W	Vertical
5	N 64 E	Vertical
6	N 58 W	79 SW
7	N 40 E	Vertical
8	N 55 E	Vertical
9	N 1 E	Vertical
10	N 11 E	83 NW
11	N 58 W	Vertical
12	N 42 W	Vertical
13	N 35 E	Vertical
14	N 49 W	Vertical
15	N 62 E	82 NW
16	N 41 E	Vertical
17	N 5 E	Vertical
18	N 17 E	75 NW
19	N 58 W	80 SW
20	N 38 W	Vertical
21	N 39 E	Vertical
22	N 43 E	Vertical
23	N 49 E	86 NW
24	N 32 E	Vertical
25	N 61 W	Vertical

Appendix C

DATA USED FOR THE CONSTRUCTION OF WATER ELEVATION, TEMPERATURE, AND PRECIPITATION GRAPHS FOR SENECA CAVERNS.

DATE	WATER LEVEL FEET	PRECIPITATION INCHES	WATER TEMPERATURE DEG C
7/10	102.35	0	N/A
7/11	102.31	1.01	N/A
7/12	102.39	0.32	N/A
7/13	102.62	0.01	N/A
7/14	102.96	0	N/A
7/15	103.29	0	N/A
7/16	103.55	0.37	N/A
7/17	103.64	0	N/A
7/18	103.60	0	N/A
7/19	103.52	0	N/A
7/20	103.42	0	N/A
7/21	103.30	0	N/A
7/22	103.15	0	N/A
7/23	103.00	0	N/A
7/24	102.85	0	N/A
7/25	102.72	0.21	N/A
7/26	102.58	0.01	N/A
7/27	102.55	0	N/A
7/28	102.42	0	N/A
7/29	102.28	0	N/A
7/30	102.13	0	N/A
7/31	101.99	0	N/A
8/1	101.85	0	N/A
8/2	101.71	0	N/A
8/3	101.58	0	N/A
8/4	101.44	0	N/A
8/5	101.30	0	N/A
8/6	101.15	0.23	N/A
8/7	101.01	0.25	N/A
8/8	100.88	0.11	N/A
8/9	100.76	0	N/A
8/10	100.62	0.78	N/A
8/11	100.51	0	N/A
8/12	100.37	0	N/A

DATE	WATER LEVEL FEET	PRECIPITATION INCHES	WATER TEMPERATURE DEG C
8/13	100.25	0	N/A
8/14	100.12	0	N/A
8/15	100.00	0.08	N/A
8/16	99.86	0.01	N/A
8/17	99.74	0	N/A
8/18	99.61	0	N/A
8/19	99.45	0	N/A
8/20	99.31	0	N/A
8/21	99.16	0	N/A
8/22	99.01	0	N/A
8/23	98.87	0	N/A
8/24	98.75	0	N/A
8/25	98.62	0.04	N/A
8/26	98.48	0.66	N/A
8/27	98.35	0.03	N/A
8/28	98.24	0	N/A
8/29	98.12	0	N/A
8/30	98.06	0	N/A
8/31	97.92	0	N/A
9/1	97.81	0	N/A
9/2	97.70	0	N/A
9/3	97.57	0	N/A
9/4	97.44	0.16	N/A
9/5	97.32	0	N/A
9/6	97.19	0	N/A
9/7	97.07	0	N/A
9/8	96.95	0	N/A
9/9	96.82	0	N/A
9/10	96.69	0	N/A
9/11	96.57	1.33	N/A
9/12	96.48	0.07	N/A
9/13	96.44	0	N/A
9/14	96.37	0	N/A
9/15	96.29	0.13	N/A
9/16	96.18	0	N/A
9/17	96.06	0	N/A
9/18	95.96	0.55	N/A
9/19	95.86	0	N/A
9/20	95.76	0	N/A
9/21	95.66	0	N/A
9/22	95.56	0	N/A
9/23	95.46	0.50	N/A
9/24	95.37	0.02	N/A
9/25	95.28	0.44	N/A
9/26	95.19	0	N/A
9/27	95.09	0.50	N/A
9/28	95.09	0.72	N/A
9/29	95.09	0	N/A

DATE	WATER LEVEL FEET	PRECIPITATION INCHES	WATER TEMPERATURE DEG C
9/30	95.07	0.34	N/A
10/1	95.01	0.75	N/A
10/2	95.29	0.67	N/A
10/3	95.65	0.17	N/A
10/4	96.38	1.26	N/A
10/5	97.21	0	N/A
10/6	97.97	0	N/A
10/7	98.52	0	N/A
10/8	98.85	0	N/A
10/9	98.93	0	N/A
10/10	98.89	0	N/A
10/11	98.83	0	N/A
10/12	98.74	0.11	N/A
10/13	98.64	0.36	N/A
10/14	98.54	0.02	N/A
10/15	98.50	0	N/A
10/16	98.49	0	N/A
10/17	98.45	0	N/A
10/18	98.38	0	N/A
10/19	98.30	0	N/A
10/20	98.22	0	N/A
10/21	98.14	0	N/A
10/22	98.04	0	N/A
10/23	97.94	0	N/A
10/24	97.84	0	N/A
10/25	97.73	0.30	14.2
10/26	97.63	0.05	13.5
10/27	97.54	0.07	13.7
10/28	97.42	0	13.5
10/29	97.31	0	13.0
10/30	97.20	0	12.7
10/31	97.07	0	12.9
11/1	96.95	0	12.7
11/2	96.84	0.09	12.7
11/3	96.74	0	12.8
11/4	96.63	0	12.6
11/5	96.51	0	12.6
11/6	96.48	0	12.7
11/7	96.41	0	12.7
11/8	96.36	0.38	13.1
11/9	96.27	0.02	12.7
11/10	96.14	0	12.6
11/11	96.02	0.14	12.7
11/12	95.91	0	12.7
11/13	95.82	0	12.7
11/14	95.71	0	12.7
11/15	95.59	0	12.7
11/16	95.47	0.01	12.9

DATE	WATER LEVEL FEET	PRECIPITATION INCHES	WATER TEMPERATURE DEG C
1/4	102.32	0	12.5
1/5	102.24	0	12.6
1/6	102.14	0	13.5
1/7	102.07	0	14.3
1/8	101.93	0	13.8
1/9	101.83	0	14.1
1/10	101.75	0.22	13.5
1/11	101.67	0	12.6
1/12	101.56	0	12.4
1/13	101.43	0	12.4
1/14	101.32	0.03	12.2
1/15	101.24	0	12.2
1/16	101.26	0	12.1
1/17	101.30	0	11.9
1/18	101.30	0.04	12.0
1/19	101.25	0	12.0
1/20	101.19	0	11.8
1/21	101.10	0	11.8
1/22	101.02	0	11.8
1/23	100.97	0	11.9
1/24	100.85	0	12.0
1/25	100.74	0	11.9
1/26	100.64	0	11.7
1/27	100.53	0	11.6
1/28	100.43	0	11.6
1/29	100.31	0.05	11.9
1/30	100.23	0.92	12.0
1/31	100.12	0.01	11.7
2/1	100.01	0	11.4
2/2	99.93	0.01	11.6
2/3	99.85	0	11.4
2/4	99.80	0	11.4
2/5	99.80	0	11.4
2/6	99.86	0	11.5
2/7	99.99	0	11.4
2/8	100.29	0	11.4
2/9	100.71	0	11.4
2/10	101.14	0	11.4
2/11	101.42	0	11.4
2/12	101.54	0.03	11.5
2/13	101.64	0	11.4
2/14	101.83	0.01	11.5
2/15	101.90	0	11.7
2/16	101.90	0	11.5
2/17	101.89	0	11.4
2/18	101.84	0	11.4
2/19	101.76	0	11.4
2/20	101.67	0	11.4

DATE	WATER LEVEL FEET	PRECIPITATION INCHES	WATER TEMPERATURE DEG C
11/17	95.35	0	13.0
11/18	95.23	0.77	13.0
11/19	95.12	0	13.0
11/20	95.17	0.45	14.0
11/21	95.33	0	15.9
11/22	95.49	0	15.0
11/23	95.82	0	13.5
11/24	96.02	0	13.2
11/25	96.10	0.01	12.9
11/26	96.13	0.78	14.8
11/27	96.28	0	15.0
11/28	96.77	0	14.8
11/29	97.42	0	14.5
11/30	97.96	0	13.5
12/1	98.22	0.07	13.0
12/2	98.30	0.70	13.4
12/3	98.36	0.08	14.9
12/4	98.75	0	14.6
12/5	99.24	0	14.0
12/6	99.69	0	13.0
12/7	99.91	0.21	12.8
12/8	99.98	0.58	14.5
12/9	100.03	0	14.4
12/10	100.39	0	14.1
12/11	100.97	0	14.1
12/12	101.57	0	14.0
12/13	102.00	0	12.8
12/14	102.21	0	12.5
12/15	102.27	0	12.4
12/16	102.22	0.02	12.8
12/17	102.15	0.10	14.0
12/18	102.10	0.04	13.8
12/19	102.07	0	12.5
12/20	102.05	0	12.3
12/21	102.02	0	12.3
12/22	101.98	0	12.3
12/23	101.94	0	12.3
12/24	101.90	0.73	13.0
12/25	101.88	0.04	13.4
12/26	101.97	0	14.0
12/27	102.24	0	14.1
12/28	102.50	0	14.1
12/29	102.57	0	12.9
12/30	102.57	0	12.8
12/31	102.50	0	13.0
1/1	102.48	0	12.5
1/2	102.48	0.03	12.5
1/3	102.41	0.02	12.4

DATE	WATER LEVEL FEET	PRECIPITATION INCHES	WATER TEMPERATURE DEG C
2/21	101.60	0	11.4
2/22	101.53	0	11.7
2/23	101.47	0	11.7
2/24	101.35	0	11.7
2/25	101.25	0	11.5
2/26	101.15	0	11.4
2/27	101.06	0	11.4
2/28	100.95	0.24	12.0
3/1	100.87	0.04	12.2
3/2	100.78	0.06	11.9
3/3	100.75	0.01	11.9
3/4	100.71	0	11.5
3/5	100.70	0.26	11.6
3/6	100.68	0	11.5
3/7	100.67	0	11.4
3/8	100.69	0	11.4
3/9	100.70	0	11.4
3/10	100.64	0	11.3
3/11	100.57	0	11.3
3/12	100.51	0	11.7
3/13	100.42	0	14.3
3/14	100.36	0.30	14.3
3/15	100.26	0.11	14.2
3/16	100.16	0	13.8
3/17	100.08	0	13.5
3/18	100.02	0	13.5
3/19	99.95	0	13.0
3/20	99.87	0	12.8
3/21	99.79	0	12.5
3/22	99.70	0	12.2
3/23	99.60	0	12.2
3/24	99.51	0	11.9
3/25	99.42	0.03	11.8
3/26	99.32	0	12.2
3/27	99.21	0.09	12.4
3/28	99.11	0.01	13.0
3/29	99.00	0.16	13.0
3/30	98.90	0.70	13.9
3/31	98.91	0	13.7
4/1	99.16	0.09	13.5
4/2	99.54	0.01	13.3
4/3	100.23	0	13.4
4/4	101.04	0.63	13.4
4/5	101.76	0.43	13.5
4/6	102.98	0.07	13.9
4/7	106.10	0.01	13.9
4/8	107.53	0	13.9
4/9	107.88	0	13.9

DATE	WATER LEVEL FEET	PRECIPITATION INCHES	WATER TEMPERATURE DEG C
4/10	108.10	0	13.9
4/11	108.27	0.17	13.0
4/12	108.28	0.28	13.2
4/13	108.15	0	14.0
4/14	108.02	0	13.6
4/15	107.92	0.06	13.6
4/16	107.79	0.01	13.6
4/17	107.65	0	13.6
4/18	107.52	0	13.6
4/19	107.38	0	13.5
4/20	107.23	0	13.6
4/21	107.10	0	13.6
4/22	106.96	0	13.6
4/23	106.85	0.17	13.3
4/24	106.69	0.15	13.0
4/25	105.51	0	12.6
4/26	106.35	0	12.3
4/27	106.21	0.10	12.0
4/28	106.08	0	11.9
4/29	105.93	0	11.8
4/30	105.80	0	11.7
5/1	105.60	0.01	11.7
5/2	105.45	0.33	11.8
5/3	105.28	0.48	11.9
5/4	105.10	0	11.7
5/5	104.94	0	11.6
5/6	104.80	0.01	11.6
5/7	104.67	0	11.6
5/8	104.51	0.01	11.7
5/9	104.35	0.01	11.7
5/10	104.18	0	12.0
5/11	103.99	0.02	11.9
5/12	103.84	0	11.8
5/13	103.65	0	11.7
5/14	103.51	0.01	11.6
5/15	103.36	0	11.6
5/16	103.16	0	11.5

27 out of 50 TURBINES SITED AT SINKHOLES

REPUBLIC WIND LLC

The red points are field confirmed sinkholes, orange are suspect field visited, yellow are suspect not visited, and blue are springs

The large plus signs are FAA Wind Turbine Coordinate Layout