SOUTHEASTERN GEOLOGICAL SOCIETY

GUIDEBOOK NO. 26

KARST HYDROGEOLOGY AND GEOMORPHOLOGY OF THE DOUGHERTY PLAIN, SOUTHWEST GEORGIA

BY

BARRY F. BECK & DANIEL D. ARDEN

DECEMBER 8, 1984

KARST HYDROGEOLOGY AND GEOMORPHOLOGY OF THE DOUGHERTY PLAIN, SOUTHWEST GEORGIA

BARRY F. BECK DANIEL D. ARDEN

A Guidebook for the Annual Fieldtrip of the
Southeastern Geological Society

December 8, 1984

Published by the
Southeastern Geological Society
P.O. Box 1634
Tallahassee, Florida 32303

In cooperation with:

FLORIDA SINKHOLE RESEARCH INSTITUTE

University of Central Florida
Orlando, Florida 32816

Acknowledgements

We would like to thank Marianne Korosy and the Florida State Cave Club for several maps and data. Most of the drafting was done by James Littlefield, Jr., with help from Tommy Sharp and other geology students at Georgia Southwestern College. The manuscript was artfully typed by Miss Debbie Standridge. The patient and painstaking work of these able assistants gave this guidebook a professional finish for which the authors thank them.

This fieldtrip was originally offered at the 1983 Southeastern Geological Society of America meeting held in Tallahassee, Florida. In preparation for the second fieldtrip, sponsored by the Southeastern Geological Society and the Florida Sinkhole Research Institute, the guidebook was revised, updated, and expanded slightly. Ms. Suzette Morris aided in typing the revisions.

Introduction

The Dougherty Plain is an extensive area of karst topography developed in Southwest Georgia. The land surface is generally flat and densely pock-marked by sinkholes. The Ocala Limestone which underlies this karst plain is normally mantled by 0-100 feet of unconsolidated sediment commonly termed "residuum." This area should therefore properly be described as a covered karst (Jennings, 1971, p. 3). Whether it is a "subsoil" or "mantled" karst (that is, whether it is covered by debris formed in place or by transported sediment) is undetermined because the exact origin of the entire thickness of residuum is undetermined and it is probably formed by both processes. An early description of the Dougherty Plain physiographic province is given by La Forge, and others (1925). The most recent discussion of the physiographic boundaries was by Clark and Zisa (1976), which the description below generally follows.

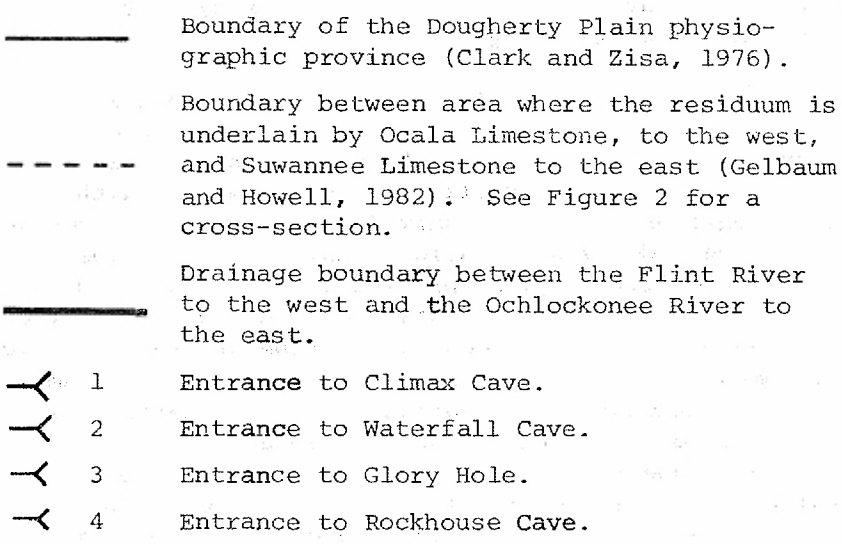
The Dougherty Plain is a generally northeast trending triangular area (Figure 1 and 2). It is bounded on the northwest, approximately along the 250 foot elevation contour, by the updip thinning of the eroded limestone wedge (Figure 2) where the Dougherty Plain gradually merges with the Fall Line Hills physiographic province, a highly dissected area of considerable relief developed on the underlying clastic rocks. To the southeast of the Dougherty Plain lies the Tifton Upland province, an area of normal fluvial drainage developed on the clastic sediments overlying the Ocala Limestone. The boundary between these two provinces is a northeast trending, highly insized cuesta escarpment known locally as the Curry Ridge or Pelham Escarpment. The northwestern and southeastern boundaries of the Dougherty Plain are not parallel and intersect to the northeast near Cordele in Crisp County. The western limit of the Dougherty Plain is normally considered to be the Chattahoochee River which marks the Georgia-Alabama state line. However, it is obvious that the geology does not abruptly change there, and similar topographic featues must extend into southeastern Alabama as noted by LaMoreaux (1948) and Stringfield, LeGrand, and LaMoreaux (1977).

A Brief Introduction to Karst for Students

The following general description of the processes forming a karst terrain is intended for undergraduate students with as yet a limited background. It is hoped that this will help make this field trip more meaningful to our student participants. The discussion is largely excerpted from Beck (1980), with pertinent additions.

Karst topography is a land surface characterized by bedrock which has been dissolved by chemical weathering rather than worn away by the abrasion of rivers. Karst surfaces are commonly marked by sinkholes, caves, and an absence of surface streams (Figure 3). Although karst topography may form on any type of soluble rock, it is most extensively and intensively developed on limestone. The solution of limestone is actually a reaction involving the gas, carbon dioxide (CO_2). Limestone is only sparingly soluble in pure water. However, as carbon dioxide is added to water a weak acid, carbonic acid, forms. This acid helps attack the limestone, and reacts with it to change it into soluble products. The more carbon dioxide there is, the more limestone can be dissolved.

Figure 1: (On facing page) The Dougherty Plain.



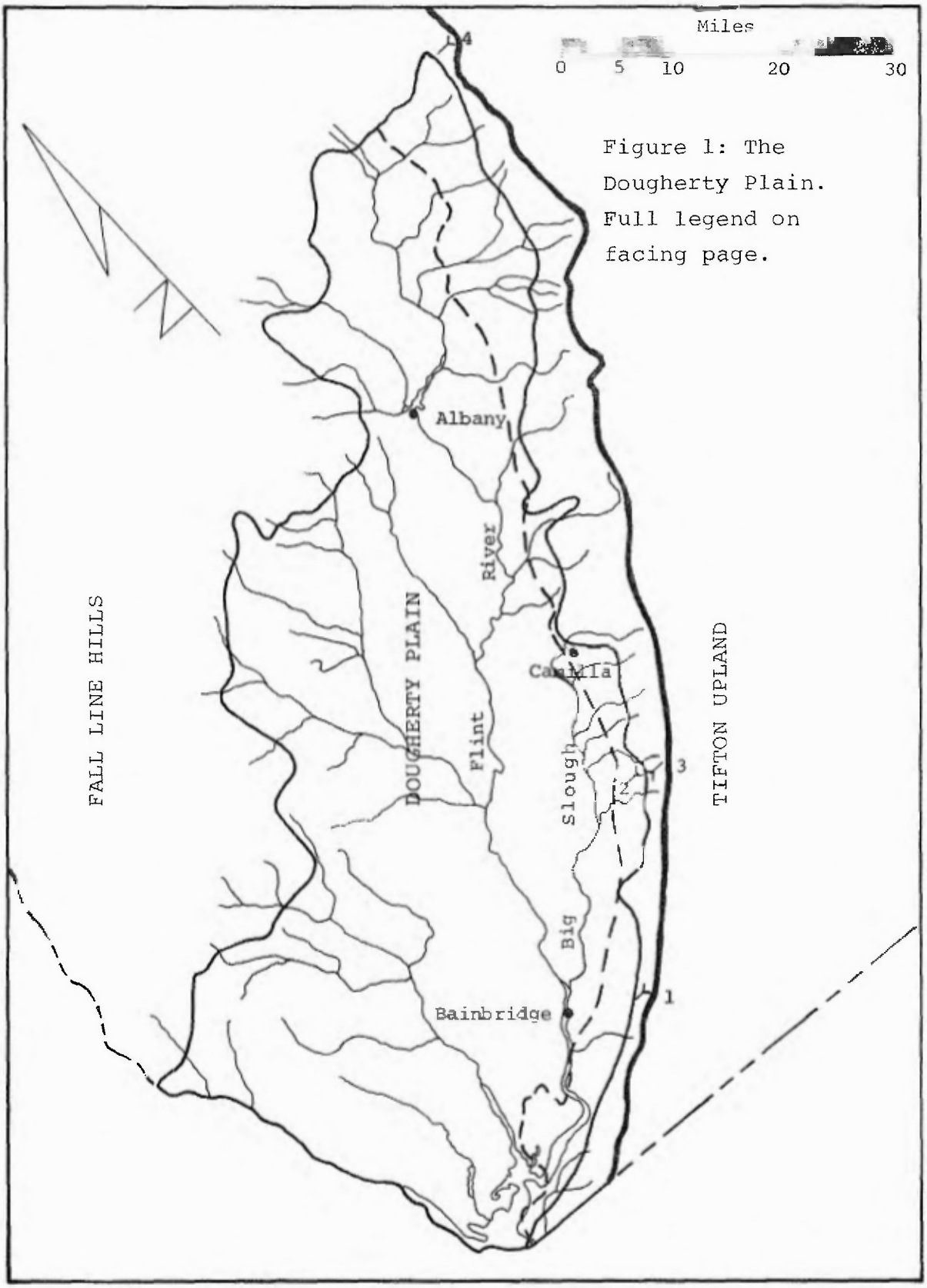


Figure 1: The Dougherty Plain. Full legend on facing page.

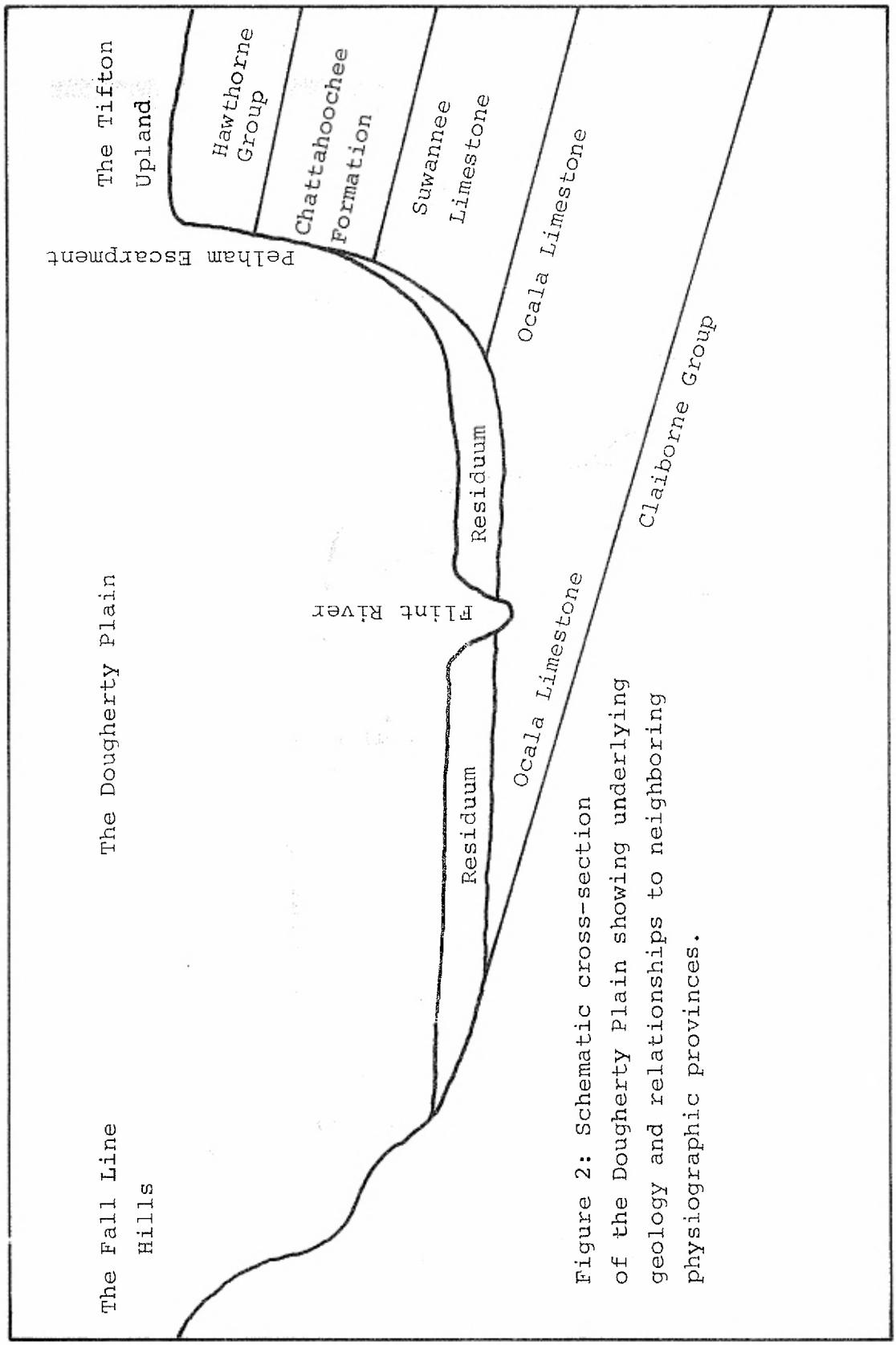


Figure 2: Schematic cross-section of the Dougherty Plain showing underlying geology and relationships to neighboring physiographic provinces.

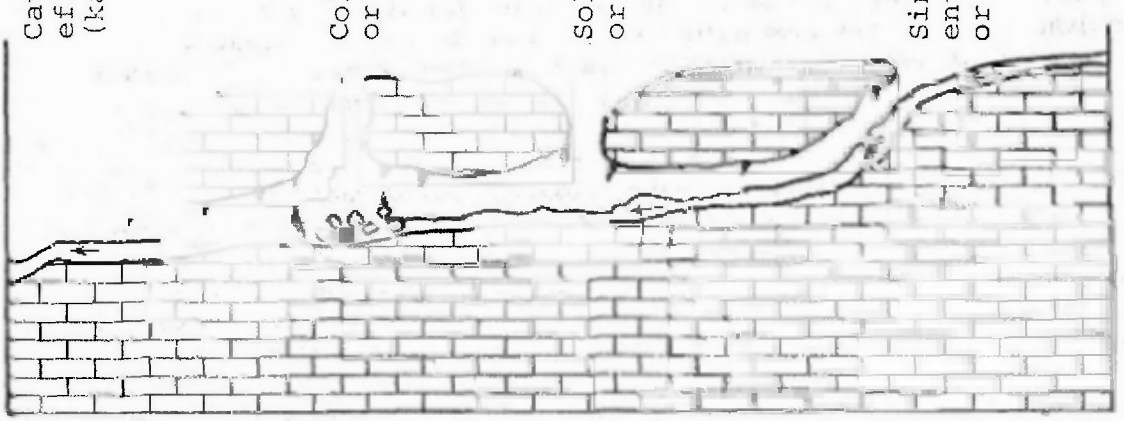


Figure 3: Generalized cross-section of some typical karst features (Beck, 1980).

We can write this reaction chemically as follows:

$$CO_2 + H_2O \rightarrow H_2CO_3$$

carbon dioxide — water — carbonic acid

$$H_2CO_3 + CaCO_3 \rightarrow Ca^{++} + 2(HCO_3)^-$$

carbonic acid — limestone — calcium ions — bicarbonate ions — in solution

When bare limestone is exposed at the earth's surface, it is subject to the attack of rainwater weakly acidified by the small amount of carbon dioxide in the atmosphere. However, when rainwater percolates through the soil, it absorbs more carbon dioxide, which is formed by the decay of plant material. Thus, if the limestone is soil-covered, the acid attack is much more intense.

Climate also plays a role in this solution process. Cold water dissolves more carbon dioxide than warm water does. This is easy to visualize when you think of a can of soda. Warm soda fizzes a lot when opened (loses carbon dioxide), but cold soda does not, because it can hold more carbon dioxide in solution. It appears, as it did to some early geologists, that karst formation would be more intense in cold climates. However, other factors counteract the basic temperature effect. In warm climates considerably more vegetation is present, and it grows all year around. Thus the soil is more carbon dioxide rich. Also, in tropical climates considerably more precipitation is available to dissolve the limestone. Further, most chemical reactions proceed faster at warmer temperatures. Thus, the combination of all these factors makes the process of limestone solution much more intense in warmer climates.

When precipitation falls on a porous rock layer, it soaks into the rock (infiltration) and seeps downward under the force of gravity. Most limestone is not porous, but is cut by numerous cracks called joints. Water seeps downward along these joints, eventually reaching a level where all fractures and openings are full of water. The upper limit of this water-filled zone is known as the water table. The water beneath the water table flows laterally and drains from the rock at low points, or springs. It is this lateral flow of ground water which forms caves.

When water first seeps into the limestone from the soil, it is very rich in carbon dioxide. It dissolves limestone rapidly, losing most, but not all, of its ability to dissolve. This rapid surface solution creates sinkholes and vertical shafts. However, when the descending water reaches the water table it can still dissolve more limestone. As it moves laterally in a thin zone just below the water table, it enlarges the cracks through which it is flowing. If one crack is somewhat larger, it transmits more water flow. The larger volume of water is capable of dissolving more limestone, so this crack grows faster than the others. As this process perpetuates itself, one crack eventually carries the vast majority of the flow and is enlarged into a cave.

The cave, then, is a water conduit. It collects water from one area through sinkholes or cracks in the limestone, and transmits it to another area through a naturally formed plumbing network, finally to rise again at the land surface as a spring. Most limestone caves form as a result of this water circulation. Their original hydrologic role frequently is obscured by the breakup of the cave into many isolated segments due to roof collapse, sediment buildup, or massive mineral deposition. Indeed, local valleys may deepen until the cave is left high above ground water level, and totally dry.

While the majority of karst areas in the world are formed on dense limestones where the water can only penetrate initially along joints, the young limestones of South Georgia and Florida are an exception. These limestones are commonly riddled with erratically shaped pores of various sizes ranging from fractions of a millimeter to tens of meters (caves). These pores are similar to the openings in a sponge. In these limestones, the water moves both along joints and through the numerous pores. Thus, there are some distinct differences between karst in South Georgia and Florida and the more classical karst areas such as Kentucky or Tennessee.

Stratigraphy of the Dougherty Plain

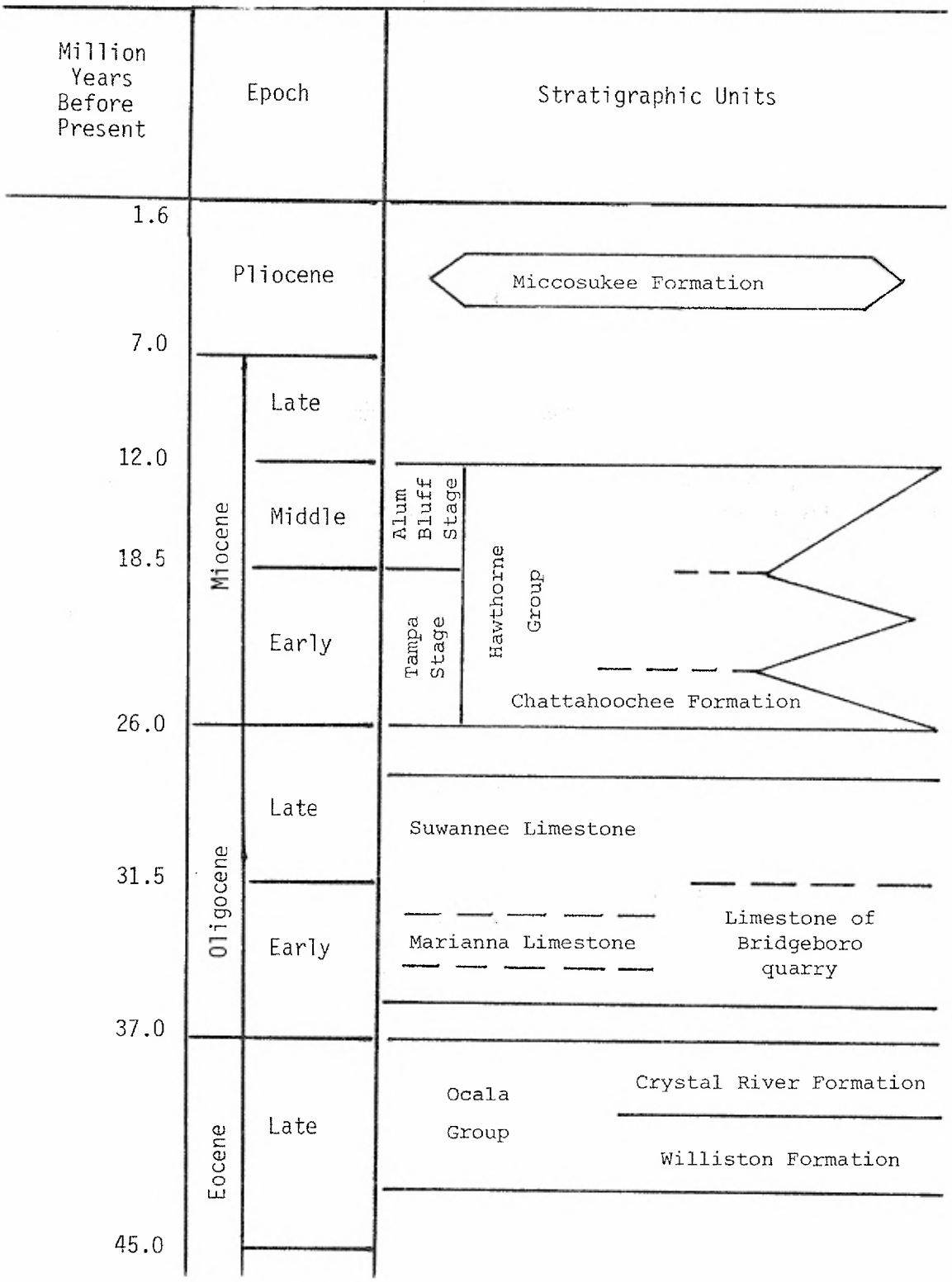
Outcropping formations of the Dougherty Plain region range in age from late Eocene to middle Eocene. Underlying the plain itself are rocks of the Ocala Group, while the eastern boundary, including the Pelham Escarpment, marks the contact with the overlying Suwannee Limestone (sensu lato), and extending eastward from the escarpment are deposits of the Hawthorne Group. Still farther east are outcrops of the Miccosukee Formation of Pliocene age (Georgia Geological Survey, 1976). For ages and correlations, see Table I. Tom Scott of the Florida Geological Survey has recently proposed revisions in the same section in Florida (Beck and others, 1984).

Ocala Group

The Ocala Limestone was first described by Dall in 1892 from exposures in the vicinity of Ocala, Marion County, Florida. The type locality was destroyed by quarrying operations, and a cotype locality for the redefined Ocala Group was designated by Puri (1957, p. 24) as the Zuber pit of the Cummer Lime and Manufacturing Company in SE¼ NW¼ sec. 11, T14S, R21E, Marion County, Florida, where 70 feet of limestone is exposed. Outcrops in the Dougherty Plain seldom expose more than 50 feet of Ocala Limestone, but drill holes in Dougherty and Decatur Counties encountered approximately 255 feet, and in Mitchell County, probably in a graben, 395 feet were penetrated.

The Applins (e.g., 1964, p. 120) recognized lower and upper members of the Ocala Limestone, probably equivalent respectively to the Williston Formation and Crystal River Formation as defined by Puri and Vernon (1964). The lower unit is generally a rather dense white limestone, sometimes nodular, and with well preserved fossils. These include abundant bryozoan fragments, echinoderms, and many small foraminifers. Some beds consist of a homogeneous microcoquina made up almost entirely of foraminiferal tests and fragments of bryozoa. The basal portion may contain beds of secondary dolomite up to 12 feet thick.

TABLE I: OUTCROPPING TERTIARY STRATIGRAPHIC UNITS OF THE DOUGHERTY PLAIN REGION



The upper unit consists of white to cream-colored detrital limestone, often porous and chalky, and generally with poorly preserved fossils, among which fragments of Lepidocyclina are common. In the subsurface, the top of the unit may consist of a few feet of tan to light brown dolomitic limestone.

"Suwannee Limestone"

The term Suwannee Limestone has, at times, been applied to all limestones of Oligocene age in the Georgia-Florida area. However, it is probable that it would be best to restrict the name to strata of Chickasawhayan (late Oligocene) age and recognize an unconformity between early Oligocene Vicksburgian beds and the late Oligocene limestones. Thickness of the sequence is 100 to 200 feet along the Pelham Escarpment and may exceed 400 feet in a graben in Grady County, Georgia.

The name Suwannee Limestone was introduced by Cooke and Mansfield in 1936 and described more fully by Mansfield in 1937. Their type section was along the Suwannee River between Ellaville and White Springs, Hamilton and Columbia Counties, Florida. Clearly, Cooke and Mansfield (1936) intended that the Suwannee Limestone represent beds of Chickasawhayan age. They applied it to "yellowish limestone typically exposed along the Suwannee River in Florida, from Ellaville, where it unconformably overlies white limestone containing Vicksburg (Oligocene) fossils, almost to White Springs, near which it lies unconformably below the Miocene Hawthorn formation." Cooke correlated the Suwannee with the lower Chickasawhay Marl. Whether this unit, as here restricted, occurs in the Dougherty-Mitchell County area is probable but not documented in publication. It seems to be the equivalent of Cooke's (1943) Flint River Formation, a term applied to in situ cherts that contain a Chickasawhayan fauna. Possibly it is present as a bedded deposit only in the subsurface.

The Marianna Limestone, which has a limited distribution and outcrops only in Jackson and Holmes Counties, Florida, is a cream to white, granular, massive limestone. It is Vicksburgian in age and is characterized by an abundance of the zone fossil Lepidocyclina mantelli.

The Marianna Limestone forms the roof of the caverns at Florida Caverns State Park, where it overlies the Crystal River Formation, within which the caverns are developed.

A limestone of quite different facies is found in the Dixie Lime and Stone Company quarry 6.5 miles southwest of Bridgeboro, Mitchell County, Georgia, but is probably correlative in age. This is dominantly an algal reef, a feature not common in the Southeastern Coastal Plain.

Glawe (1974, p. 18) gives the following section in the quarry:

Suwannee Limestone	Thickness (feet)
5. Clay, red sandy, with silicified mollusks	5.0+
4. Limestone, white, weathers brown bioclastic; contains Chlamys duncanensis, Clypeaster cotteaui, and abundant molds of mollusks	2.5

Suwannee Limestone	Thickness (feet)
3. Limestone, same as bed above but with algal balls and Cassidulus gouldi	2.0
2. Limestone, white, nodular, bioclastic, with green clay pockets; contains Sondylus sp., Lepidocyclina sp., oysters, Chlamys duncanensis (rare), and algal balls	16.5
1. Limestone, cream, crumbly, bioclastic, with numerous small pockets of green clay; contains Lepidocyclina sp., Clypeaster cotteaui, Chlamys duncanensis and abundant algal balls	54.0+
	80.0+

Hawthorn Group

In southwest Georgia, the exposed Miocene formations are updip facies of shallow marine, paralic, and terrestrial environments. Presently, there is no consensus regarding details of age and correlation throughout the region, and a number of formation names have been proposed but not universally accepted. For example, "Tampa Limestone" or "Tampa Formation" are no longer in general use, although the Florida Geological Survey uses "Tampa Stage" for beds of early Miocene age (Puri and Vernon, 1964).

In the Lake Seminole area, the Tampa Stage is represented by the Chattahoochee Formation, which consists of sandy and clayey limestones and dolomites interbedded with clayey sands and sandy clays. The clays include montmorillonite and attapulgite, which provides the fullers earth of the Attapulgus, Georgia, region. The type locality for the formation is the Jim Woodruff Dam section in the SW¼ sec. 29, T4N, R6W, on the access road to the dam directly below the U. S. Engineers office on the east side of the Apalachicola River (Puri and Vernon, 1964, p. 118).

The term "Hawthorne beds" was introduced by Dall in 1892, without designation of a type locality. Pirkle (1956, p. 209) described a cotype locality at Brooks Sink, Bradford County, Florida. These beds are generally considered to be middle Miocene in age and essentially constitute the Alum Bluff Stage in the region. In outcrop the formation is a mottled sandy clay, while under cover it is phosphatic and contains tongues of arkosic, phosphatic sand.

Hydrogeology

The hydrogeology of the Dougherty Plain is intimately related to the geomorphology, being both cause and effect thereof, and accordingly, many facets of the hydrogeology will be discussed below under the topic of geomorphology. However, it is also pertinent to understand the broader regional setting in which the Dougherty Plain occurs and the detailed characteristics of the Ocala Limestone as an aquifer.

The Dougherty Plain is developed on the exposed updip portion of the Ocala and Suwannee Limestones (Figure 2). Downdip, to the south and southeast, where these formations are confined by the overlying strata, they are the principal stratigraphic units included in one of the nations most prolific aquifers (LaMoreaux, 1982) which is known as the principal artesian aquifer in Georgia and the Floridan aquifer in Florida. Also downdip, the underlying Claiborne Group changes facies to a limestone and it too is generally included with the principal artesian aquifer. In coastal Georgia, the principal artesian aquifer is the major water source for almost all uses.

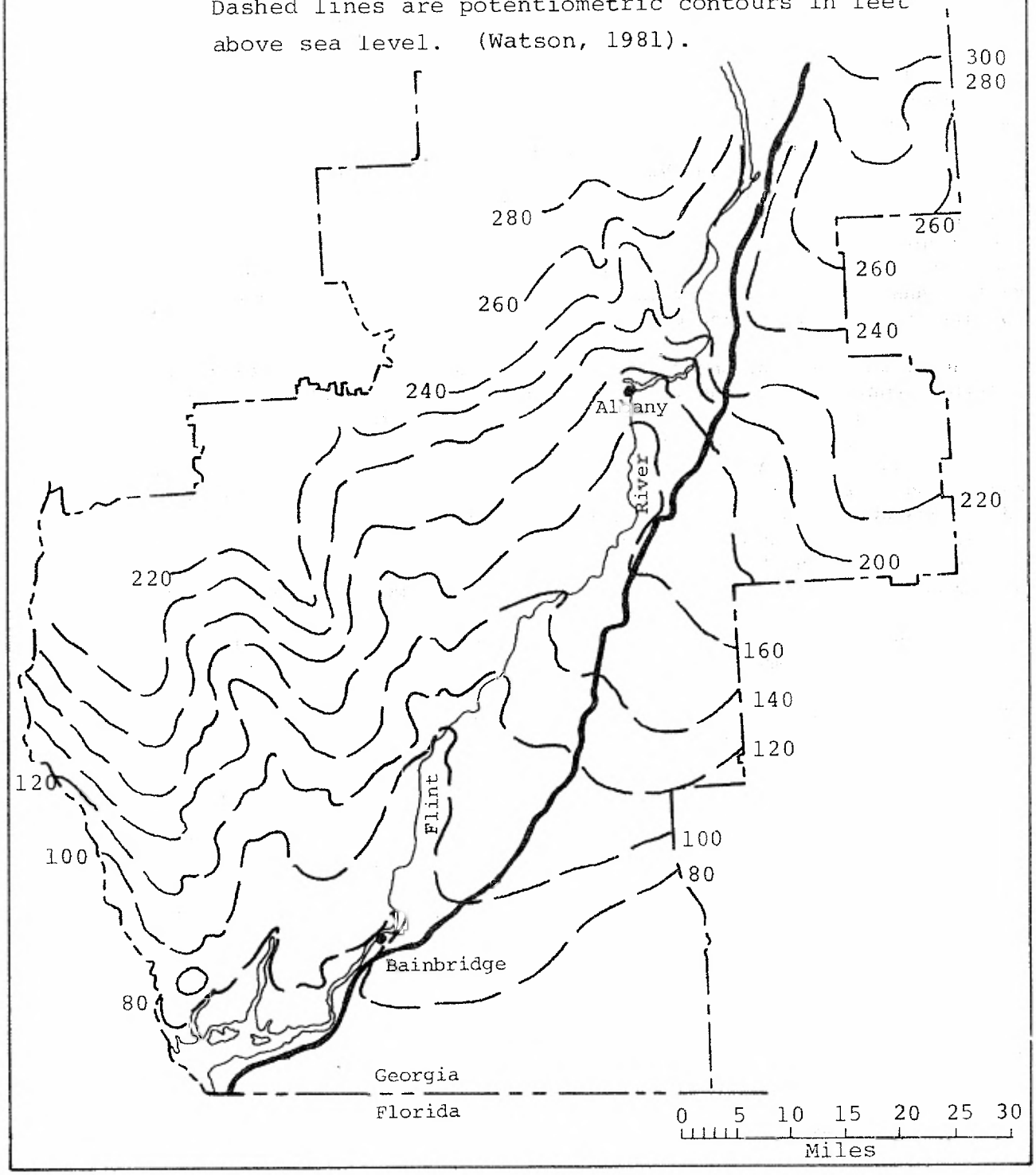
It is a common misconception that the Dougherty Plain is a recharge area for the principal artesian aquifer, either in coastal Georgia or to the south in the Florida Panhandle. The potentiometric surface (see Figure 4) indicates that there is a major groundwater divide occurring roughly beneath the Pelham Escarpment; this divide also continues into the Florida Panhandle (Wagner, 1981). To the southeast of this divide, beneath the Tifton upland, groundwater in the principal artesian aquifer moves downdip generally towards the southeast. However, in the Dougherty Plain, groundwater in the Suwannee and Ocala Limestones moves toward the Flint River from both sides. This trend also continues along the Apalachicola River, which is the southerly extension of the Flint River in Florida (Wagner, 1981). Thus, the outflow for the hydrologic cycle in the Dougherty Plain is basically the surface drainage of the Flint River.

The Ocala Limestone underlies the major area of the Dougherty Plain and the overlying Suwannee Limestone is only present in a narrow belt along the base of the Pelham Escarpment (Figure 1). Both limestones have been subjected to repeated cycles of sea level fluctuation. During some regressions they were subaerially exposed producing unconformities marked by paleokarst surfaces (Randazzo, 1982; Schmidt and Scott, 1984; Popenoe and others, 1984). During the transgressions and regressions the saltwater/freshwater mixing zone also oscillated. This zone has been related to various diagenetic changes. "A major environment of regional dolomitization is in the mixing zone (zone of dispersion) where profound changes in mineralogy and redistribution of porosity and permeability occur from the time of early emergence and continuing through the time when the rocks are well-developed aquifers. The reactions and processes, in response to mixing waters of differing chemical composition, include dissolution and precipitation of carbonate minerals in addition to dolomitization" (Hanshaw and Back, 1979, p. 287). It has been noted that the more permeable zones of the Floridan Aquifer are dolomitic (e.g., Lichtler and others, 1968).

On the other hand, Randazzo (1982, p.9) identifies the same processes -- "dissolution and the creation of cavernous porosity, infilling of pore spaces by sparry calcite cements, calcitization and dolomitization" -- but ascribes their occurrence to "interaction with a phreatic meteoric water environment." Whatever the exact mechanisms, various zones within the Ocala and Suwannee Limestones have been modified by these, and possibly other, processes to produce a spongework of large pores with dimensions measured in centimeters. This macroporosity can be seen in the banks of the Rise Pool at Radium Springs and where the Radium Springs Run joins the Flint River.

Zones of unusually high macroporosity are developed at unconformities at the contact of the Ocala Limestone and the Suwannee Limestone and at the contact of the Suwannee Limestone and the Hawthorne Formation (Burnson, 1981). The Ocala-Suwannee contact has been shown to be a locus of karst development and groundwater circulation at Peacock Springs Cave in Florida (Fisk and Exley, 1977).

Figure 4: Potentiometric surface of the principal artesian aquifer, May 1980, in the Dougherty Plain. Heavy line shows approximate location of Pelham Escarpment. Dashed lines are potentiometric contours in feet above sea level. (Watson, 1981).



The transmissivity of the Ocala Limestone is highly variable due to the complex development of secondary porosity, as discussed above. However, all measured transmissivities for the Ocala and Suwannee Limestones in the Dougherty Plain area are high, except of course on the updip edge where the limestones pinch out. Mitchell (1981, p. 33) reports transmissivities calculated from pumping tests in the Dougherty Plain ranging from 21 to 112 thousand feet squared per day. The storage coefficients ranged from 0.002 to 0.02 (ibid.). Kwader and Wagner (1980) estimated T values from the specific capacities of high yield production wells and calculated an average T for the Dougherty Plain of 87,000 feet squared per day. For the southern portion of the Dougherty Plain (south of Dougherty County), where the aquifer is thicker, the average T was approximately 125,000 feet squared per day (ibid.). Sever (1965) conducted pumping tests in the southern Dougherty Plain and calculated T's ranging from 53,000 feet squared per day at Donaldsonville to 334,000 feet squared per day near Bainbridge. Storage coefficients calculated from the same tests averaged 0.001 and showed little variation (ibid.). In contrast, northwest of Albany where the aquifer is thinner, Hicks, Krause, and Clarke (1981) report that the transmissivity of the Ocala Limestone can be as low as 2,000 feet squared per day.

Within the Principle Artesian Aquifer, then, lateral flow of groundwater is somewhat diffuse and homogeneous, although the greatest volume of flow occurs within zones of megaporosity. Hickey (1984) confirmed that flow in this aquifer near St. Petersburg, Florida, is diffuse (Darcian), rather than conduit (a la Shuster and White, 1971) based on aquifer test data.

Dispite the high overall permeability, preferential solution of the limestone along joints has produced large, cavernous solution channels, or conduits, where a concentration of groundwater flow occurs. The joint-controlled nature of these solution features was demonstrated by an analysis of passage trends in caves underlying the Pelham Escarpment (Ellwood, 1973) as well as by the "Sinkhole Subsidence Susceptability Model" developed by Brook and Allison (1981).

Beck and others (1984) have proposed a conceptual framework for the karst hydrologic cycle in northern Florida. With minor modifications these principles apply equally well to the Dougherty Plain. They are summarized below.

1. The Tertiary limestones of the Principle Artesian Aquifer are generally highly permeable.
2. Within the Principle Artesian Aquifer there are laterally persistent zones of megaporosity due to diagenesis and paleokarst.
3. Over millions of years joints, fractures, and bedding planes in the limestone have been enlarged by karstic dissolution. However, dissolution does not generally cause perceptible changes in man's time frame.
4. Although dissolution may have created anisotropy in the aquifer, groundwater flow is generally diffuse.
5. Recharge to the aquifer is diffuse except where sinkholes, or structural or erosional defects, breach the confining layer; where the limestone is at, or near, the land surface sinkholes may still facilitate rapid recharge.
6. Modern sinkhole collapse is generally caused by subsidence of the unconsolidated (or semiconsolidated) overburden into preexisting solution pipes or other karstic voids.
7. Karstic conduits or cavities have formed where groundwater flow is concentrated.
   A. Beneath individual sinkholes or clusters of sinkholes;
   B. Beneath vertical shafts or ponors (sinking streams) along the edge of

the confining layer (the Pelham Escarpment) where greater volumes of aggressive water enter the limestone; conduit enlargement is greatest where flow is greatest and the resurgences of major rivers may show true conduit flow. (Several major rivers sink and resurge in Florida, but none occur on the Dougherty Plain.)

C. At artesian springs where a large volume of diffuse flow converges radially to a relatively small outlet.

Irrigation in the Dougherty Plain

Any description of water use in Southwest Georgia is basically a discussion of irrigation. Much of the Dougherty Plain is level and underlain by well-drained, sandy soils developed on the limestone "residuum". The natural precipitation ranges from 46" per year near Cordele to 52" per year near Bainbridge. The frost-free season generally extends from late February to October and is suitable for double-cropping. In the Southern Dougherty Plain (south of Albany) the Ocala Limestone is thickest and properly constructed wells drilled almost anywhere will yield sufficient water to supply a major irrigation system. The Dougherty Plain is obviously prime agricultural land.

Natural precipitation is sufficient to sustain crop growth in most years in the Dougherty Plain. However, (1) natural precipitation may not occur when needed, such as during seed germination; (2) droughts may periodically occur with accompanying financial hardships for the farmer; and (3) recent research has demonstrated that even in years when natural precipitation is adequate, judiciously applied irrigation water may increase crop yields, and profits, substantially. Irrigation in the Dougherty Plain first began to be significant in the late 1950's, following a severe drought in 1954 (Kwader and Wagner, 1981). However, as recently as 1970 the use of water for irrigation was still only minor compared to the present consumption. During the 1970's additional drought periods and modern automated irrigation technology led to dramatic increases in the total number of acres irrigated in Georgia (most of the increase being in Southwest Georgia). Figure 5 shows both the total number of irrigated acres in Georgia and the average yearly increase in irrigated acreage (after U.S.D.A. data: Harrison, 1981).

There are three basic methods of irrigation which account for 90 percent of the irrigated acreage in Georgia. Portable pipe systems consist of a mainline and one or several laterals with small to medium rotating sprinklers on the laterals. When the required quantity of water has been applied to one area the lateral is dismantled and moved to the next location by hand or by tractor, where it is reconnected. This process is repeated until the necessary acreage is covered. Figure 6 shows a possible layout for this type of system. Obviously, this type of irrigation is labor intensive.

A somewhat more automated system is the large, single gun, self-propelled system, either cable-tow or hose-reel (Plate 1). A high pressure sprinkler gun, capable of covering 1-6 acres, is towed down the center of the field, irrigating from side to side as it moves. The wheel mounted gun may be pulled by a steel cable which is stretched across the field and is rewound by water power as the gun moves, or by the hose itself which may be rewound similarly. One self-propelled sprinkler may be used to irrigate several different fields. Water consumption ranges from 300-600 gallons per minute.

The most automated and least labor consumptive of the new irrigation systems is the center-pivot. A radial main line is supported some 8' above the ground by a series of A-shaped towers, each with wheels (Plate 2). Small to medium sprinklers are mounted along the main line and the entire unit pivots around one fixed end irrigating a circular area. The sprinklers are adjusted for an even application of water. A high pressure gun may be used at the end of the

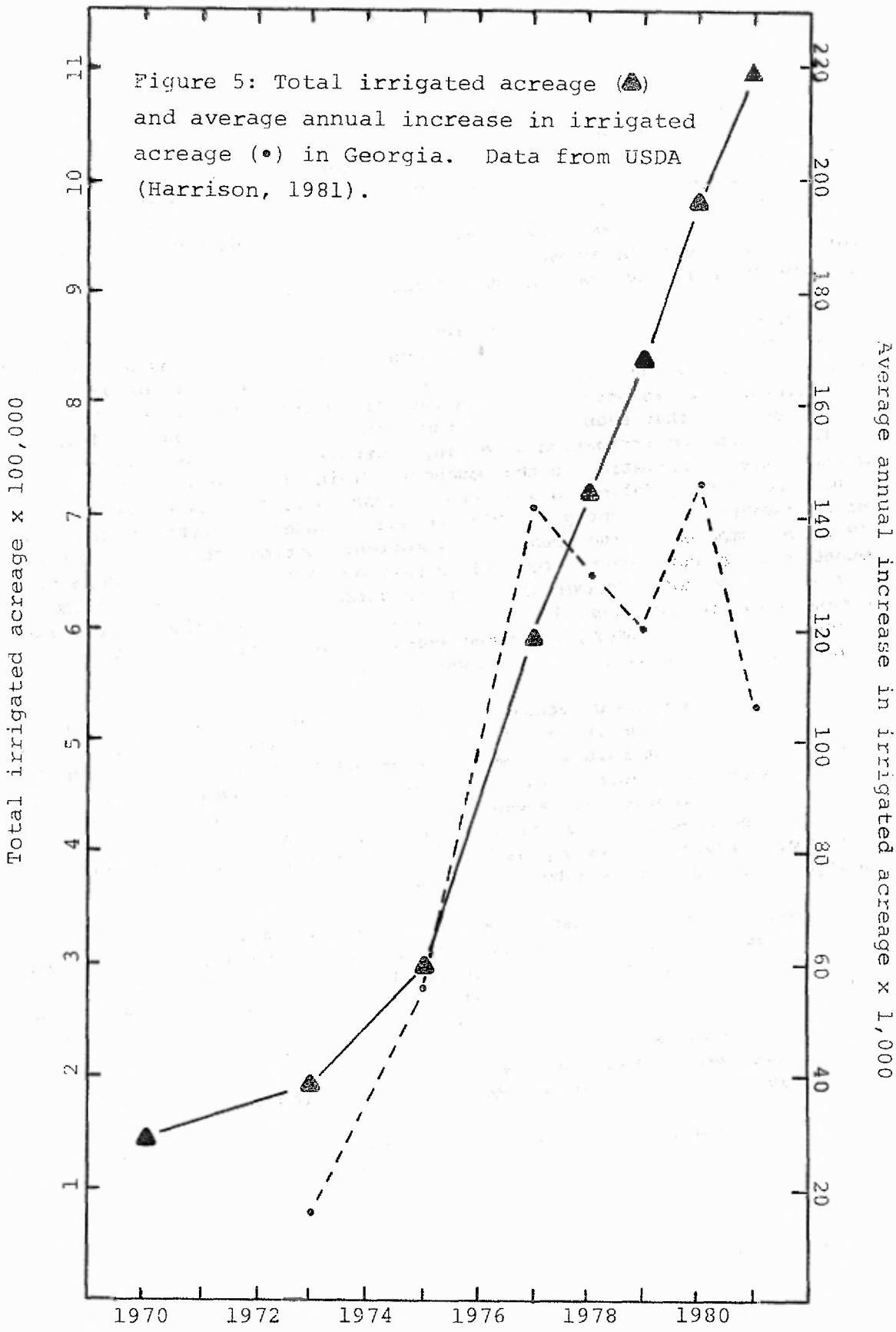


Figure 5: Total irrigated acreage (▲) and average annual increase in irrigated acreage (•) in Georgia. Data from USDA (Harrison, 1981).

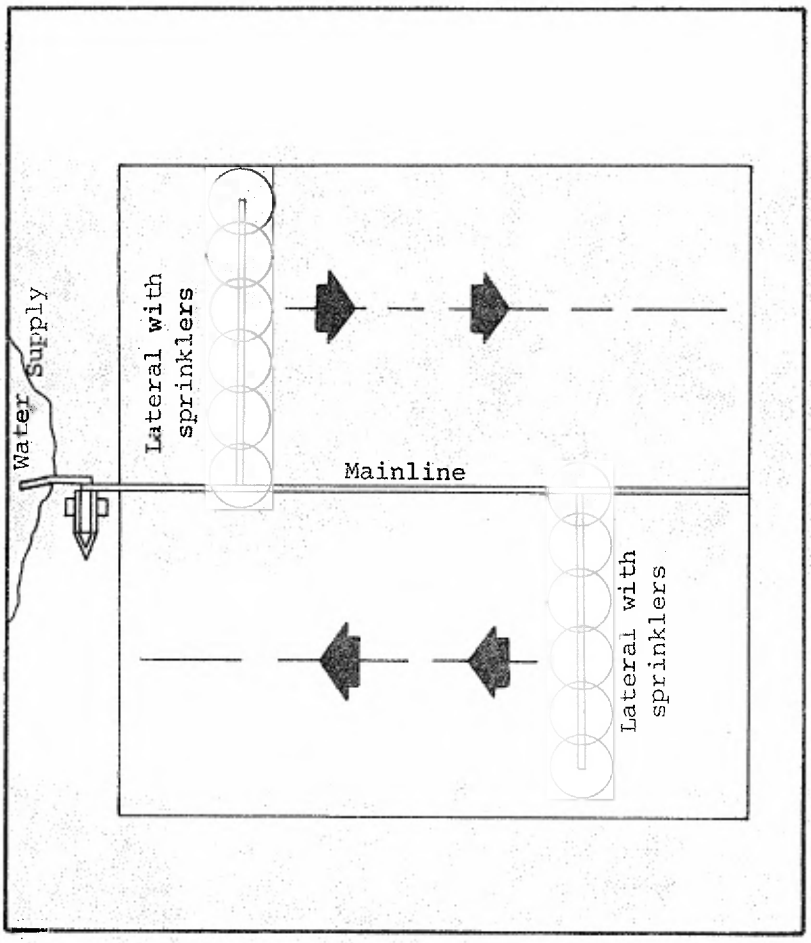


Figure 6: Schematic diagram of a field layout for a hand-moved irrigation system with two laterals and multiple sprinklers. (after Turner and Anderson, 1980).

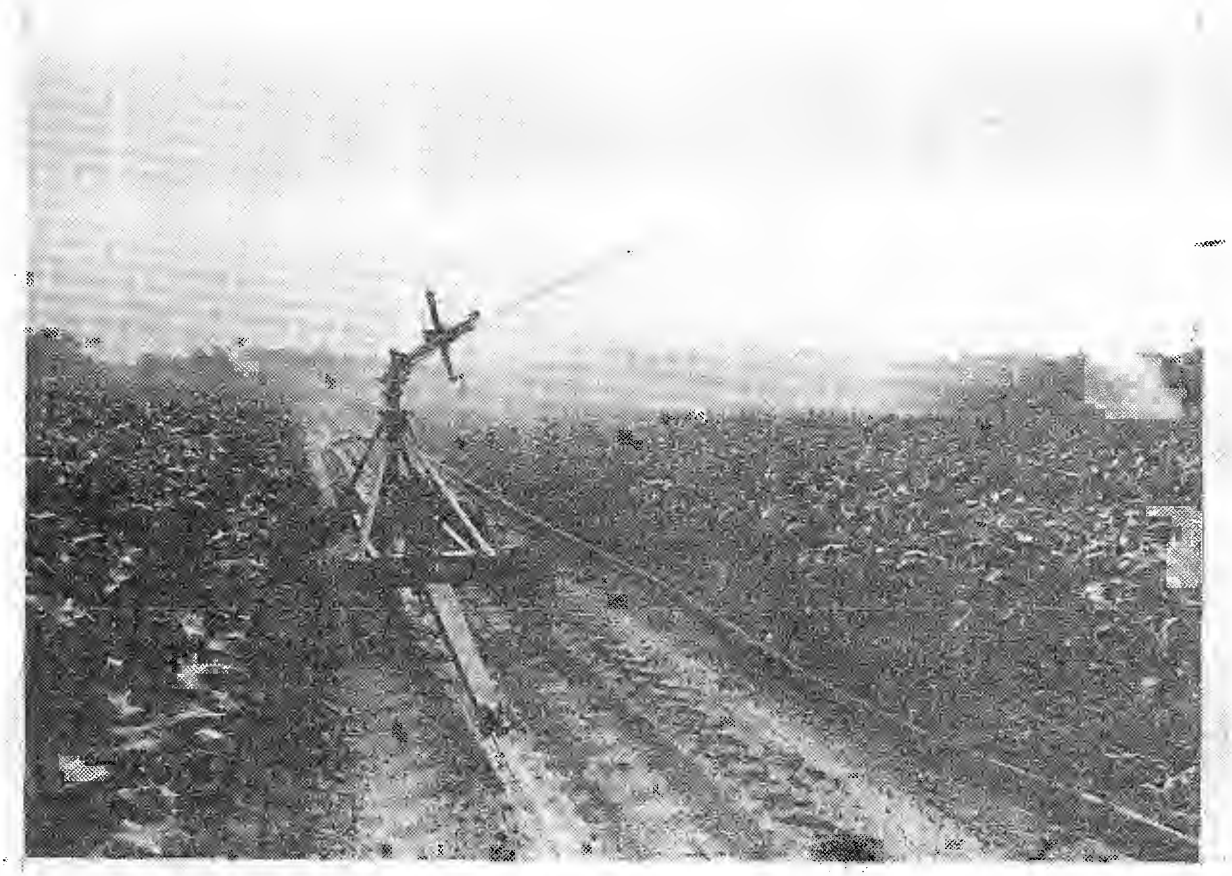
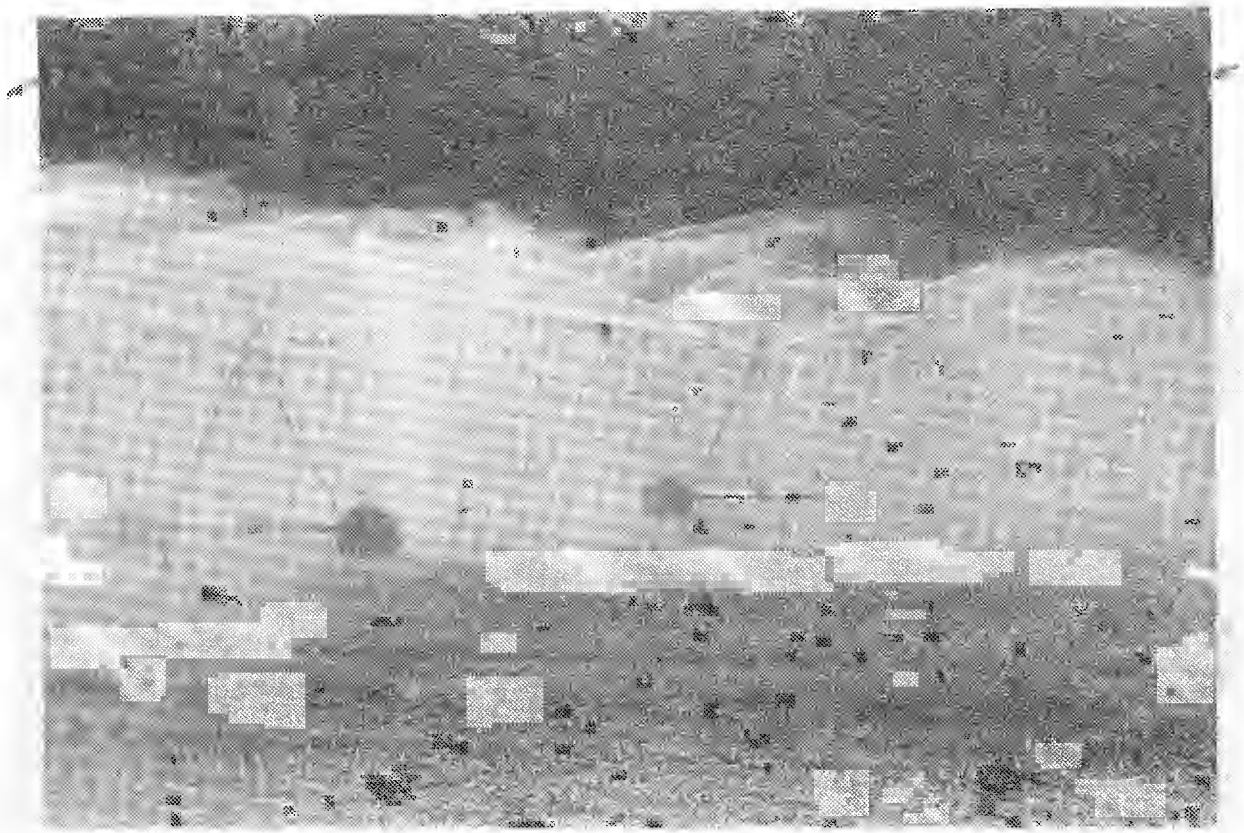


Plate 1: Cable-tow irrigation. Photo courtesy of Kerry Harrison.



main line to reach into the corners of the field. The main line has flexible couplings at the towers so that it may adjust to undulating terrain. Each tower is self-propelled by electric or hydraulic motors and the towers are kept in pace by a simple control system: if one tower gets ahead of its neighbors, the flexing of the coupling activates a switch which turns the power to that tower off; when it gets behind, the opposite flexing turns it back on. Center pivot systems are generally permanently mounted at one location, although they can be towed between several fields. A single system may irrigate up to 500 acres and consume up to 2,000 gpm (roughly 3 million gallons per 24-hour day). Most are somewhat smaller than this.

Center pivot systems have several advantages over other methods of irrigation. They are virtually labor free. They may also be used to apply chemicals to the crops, a process now known as chemigation. Fertilizers, herbicides, fungicides, and insecticides are all applied by center pivot systems. While they might also be applied by other irrigation methods, the uniformity of application is distinctly higher using center pivot systems (Threadgill, 1981). Chemigation itself has many advantages over conventional (airborne or tractor) application of chemicals.

1. It provides excellent uniformity of chemical application.
2. It allows prescription application of chemicals.
3. It allows timeliness of chemical application.
4. It allows easy chemical incorporation and activation.
5. It reduces soil compaction.
6. It reduces mechanical damage to the crop.
7. It reduces operator hazards.
8. It may reduce chemical requirements.
9. It may reduce environmental contamination.
10. It is effective.
11. It is economical. (Threadgill, 1981).

The use of the center-pivot systems in Georgia has increased markedly over the past 11 years, as have self-propelled single-gun systems, while portable, hand-moved systems have declined (see Table II). However, because center-pivot systems cover a much greater area than cable-tow or hose-reel systems, they account for many more acres of farmland. For our purposes, it is also instructive to note where the water for this greatly expanded irrigation is coming from. Over the entire state of Georgia the number of systems supplied by surface water (ponds and streams) has increased from 6,000 in 1970 to 7,100 in 1981, while the number of systems supplied by wells has increased from 600 in 1970 to 4,200 in 1981 (Harrison, 1981).

Pollard and others (1978), studied irrigation practices in Southwest Georgia. For 1976 and 1977 they found that approximately 80 percent of the irrigation water, by volume, was supplied from groundwater. In 1977, 64 percent of all irrigation systems were of the center pivot type. In the Dougherty Plain area per se (excluding the Fall Line Hills and the Tifton Upland) they reported that 92 percent of all the water consumed for irrigation was from groundwater. For the five southwesternmost counties, which are the heaviest irrigators, the average daily consumption of irrigation water during 1977 was 116 million gallons per day. Industrial and municipal water use in the same five-county area probably did not exceed 10 mgd. Even in Albany, the industrial and population center of Southwest Georgia, the average industrial and municipal water use in 1977

TABLE II: Irrigation Statistics for Georgia, 1970-1981

	1970	1973	1975	1977	1978	1979	1980	1981
Total Acreage	144,629	193,857	307,416	592,088	722,075	842,235	988,356	1,094,778
Total # Hand-Move Systems	6,365	6,440	5,026	4,179	3,613	3,228	2,517	2,125
Total # Cable-Tow and Hose-Reel	69	694	1,090	2,585	3,157	3,624	4,254	4,800
Total # Center-Pivot, Side-Wheel, and Linear	90	246	489	1,019	1,675	2,327	2,892	3,481
Total # Other (Including Golf Courses and Athletic Fields)	32	338	533	560	700	794	876	1,153
Total # Using Well Water	582	972	1,118	1,771	2,342	2,939	3,387	4,175
Total # Using Water From Ponds and Streams	5,990	6,084	5,258	6,211	6,614	6,477	6,378	7,067

Compiled from U. S. Department of Agriculture (Harrison, 1982)

was only 30 mgd. Thus, we may say that irrigation is the major consumer of water in the Dougherty Plain, and that groundwater supplies the vast majority of that water.

Kwader and Wagner (1981) have examined the possible future repercussions of this increased water use in the southern Dougherty Plain. Several points bear mentioning, summarized from their report. The southern Dougherty Plain is ideally suited for center-pivot irrigation and groundwater supplies are greatest here. The dramatic increase in irrigation over the past 10 years (1970-1980) has not been accompanied by any long-term ground water level declines. In fact, 1978 levels were the highest for the 10 year period. The total thickness of the principle artesian aquifer in the southern Dougherty Plain is on the order of 1,400 feet. Considering an average porosity of 20 percent, the volume in storage alone is tremendous. The rate of expansion of irrigation must decrease in the future as all suitable land is utilized, and in fact it is already decreasing (see Figure 5). The total area of irrigated acreage will probably not exceed 3 times the area presently (1980) irrigated. According to Kwader and Wagner (1981) this may cause some decrease in the base flow of the Flint River, but "it is doubtful that any long-term water level declines or detrimental effects to the aquifer would take place" (ibid., p. v.). It must be noted that this conclusion applies to the <u>southern</u> Dougherty Plain area only. Other areas, such as the northern Dougherty Plain where the aquifer is thinner, or the Fall Line Hills area which taps a different aquifer, may well experience detrimental effects locally if irrigation is overdeveloped.

Geomorphology

The entire Dougherty Plain is a single unified, though certainly not homogeneous, hydrologic system. Surficial drainage is limited to the Flint River and several of its tributaries. Most precipitation on the Dougherty Plain quickly infiltrates through numerous sinkholes (dolines) and recharges the groundwater. All the groundwater in the Dougherty Plain flows to the Flint River or its tributaries except along the southwestern margin where a limited area drains to the Chattahoochee River, as can be seen from a recent map of the potentiometric surface (Figure 4).

The geomorphic elements of the Dougherty Plain may be examined in a sequence analagous to the hydrogeologic cycle as it pertains to this area. Following a general description of this sequence each of these elements will be examined in detail. The description may be clarified by Figure 2. The Pelham Escarpment marks the southeastern limit of the exposed (thinly mantled by residuum) limestones. The overlying insoluble, clastic Hawthorne Group sediments dip to the southeast, so drainage on the upland is away from the Dougherty Plain. Runoff along the scarp face flows down-slope along numerous gulleys. These short surface courses flow into deep, steep-walled sinks along the lower portions of the escarpment. Near Climax, Georgia, one of these sinks drains through a vertical shaft (genetically analagous to those described by Pohl, 1955) into Climax Cave (see Figure 12B). Several explored cave systems lie beneath the escarpment edge: Climax Cave, Glory Hole, Waterfall Cave, and Rockhouse Cave among others. All of these act as recharge points to the aquifer.

At greater distances to the northwest of the escarpment, drainage is entirely into sinkholes, although short surface streams may lead into them. Most of the sinkholes are plugged and hold small ponds after rainy periods. Some of these are reported by local residents to occasionally drain rapidly into the subsurface. None lead to any explored caves. The depth of sinkholes generally decreases as the distance from the escarpment increases. Between the escarpment and the Flint River, Big Slough is the only surface drainage feature. On small scale maps (such as Ga. Geological Survey, 1976) it is generally shown as a major continuous surface stream of at least third order, with tributaries flowing off the slopes of the Pelham Escarpment. However, examination of this feature on the appropriate 7½' topographic quadrangle maps shows that this concept is patently absurd; Big Slough is an aligned series of karst drainage features as was pointed out by Beck (1980a).

Between the Big Slough and the Flint River, surface drainage is virtually absent; there are generally not even short streams leading into the sinkholes. Numerous springs are present along the Flint River on both sides. Several of these, such as Radium Springs near Albany, are the outflow of water-filled cave conduits. Radium Springs was explored by divers for over 1,000 feet (Deloach, 1982); its flow varies from 3.8 to 0.1 cubic meters per second (87 to 2.6 mgd) (Wait, 1963). Others, such as Blue Spring on the east side of the Flint River south of Albany, and Berry Springs on the west side of the Flint River south of Albany, are simply karst fensters -- sinkholes which connect downward with water-filled conduits and may overflow when groundwater levels are high. These conduits most probably debauch directly into the bed of the river. At low water the presence of springs in the river bed is visible (Wait, 1963), and seismic surveys have detected deep sinkholes below water level (personal communication from Sam Pickering and Vernon J. Henry, 1976).

To the northwest of the Flint River, the hydrogeology is a bit different. Although here also most of the drainage is through sinkholes, several permanent tributaries to the Flint River have incised themselves into the limestone, notably Kinchafoonee Creek, Itchawaynochaway Creek, and Spring Creek. Between Albany and Leesburg, Kinchafoonee Creek flows between vertical limestone walls approximately thirty feet high for several miles. One cave, now above water level, and several small springs can be seen along this reach. These streams are largely effluent, as the name Spring Creek implies, and are the discharge points for local hydrogeologic circulation. The existence of surface drainage in this area of the Dougherty Plain may be due to its longer exposure to erosion, or the thinness of the Ocala Limestone in this updip area, thus perching the groundwater level closer to the land surface, or to other causes. This question has not been examined in detail.

The geomorphic development of the Dougherty Plain in the Baker and Mitchell County area was described by Herrick and LeGrand (1964). They interpreted geomorphic development as stages in a karst erosion cycle. Furthermore, they had identified the Pelham Escarpment as a retreating escarpment developed "chiefly by headward solution-sapping of the limestone composing the scarp by water from migratory springs situated at the head of pocket valleys" (ibid., p. 33-34). In the following detailed examination of the escarpment, it will be shown that this interpretation is partially incorrect.

Development of the Pelham Escarpment

Herrick and LeGrand (1964) concluded that the Pelham Escarpment was a retreating solution escarpment formerly dissected and undermined by karst springs. They described the process thusly (ibid., p. 33-34).

> "Reduction of the upland apparently was carried on chiefly by headward solution sapping of the limestone composing the scarp by water from migratory springs situated at the heads of pocket valleys. It is conceivable that the vast solution subsidence on the Dougherty Plain was accompanied by development of pocket valleys in a manner similar in principle to the tactics employed by military forces on the offensive. Such military tactics, in general, employ the principle of selective infiltration of enemy lines, followed by straightening of battle lines through subsequent operations. Selective headward solution sapping of the scarp by streams situated at the heads of pocket valleys produced deep reentrants into the face of the scarp. Reduction of the upland at such points on the scarp was then, for all practical purposes, brought virtually to a standstill by the resulting heavy accumulation of downslumped surficial material, the overlying clays of the Hawthorne Formation. Following this stage, those portions of the scarp heretofore relatively unreduced and relatively unprotected by surface residuum were then attacked in a similar manner and also forced to retreat. Removal of some of the clay by surface erosion then exposed parts of the limestone and facilitated the issuance of springs necessary for further attack. Such reduction was more or less piecemeal, but taken as a whole it resulted in an eastward, down-the-dip retreat of the entire solution scarp."

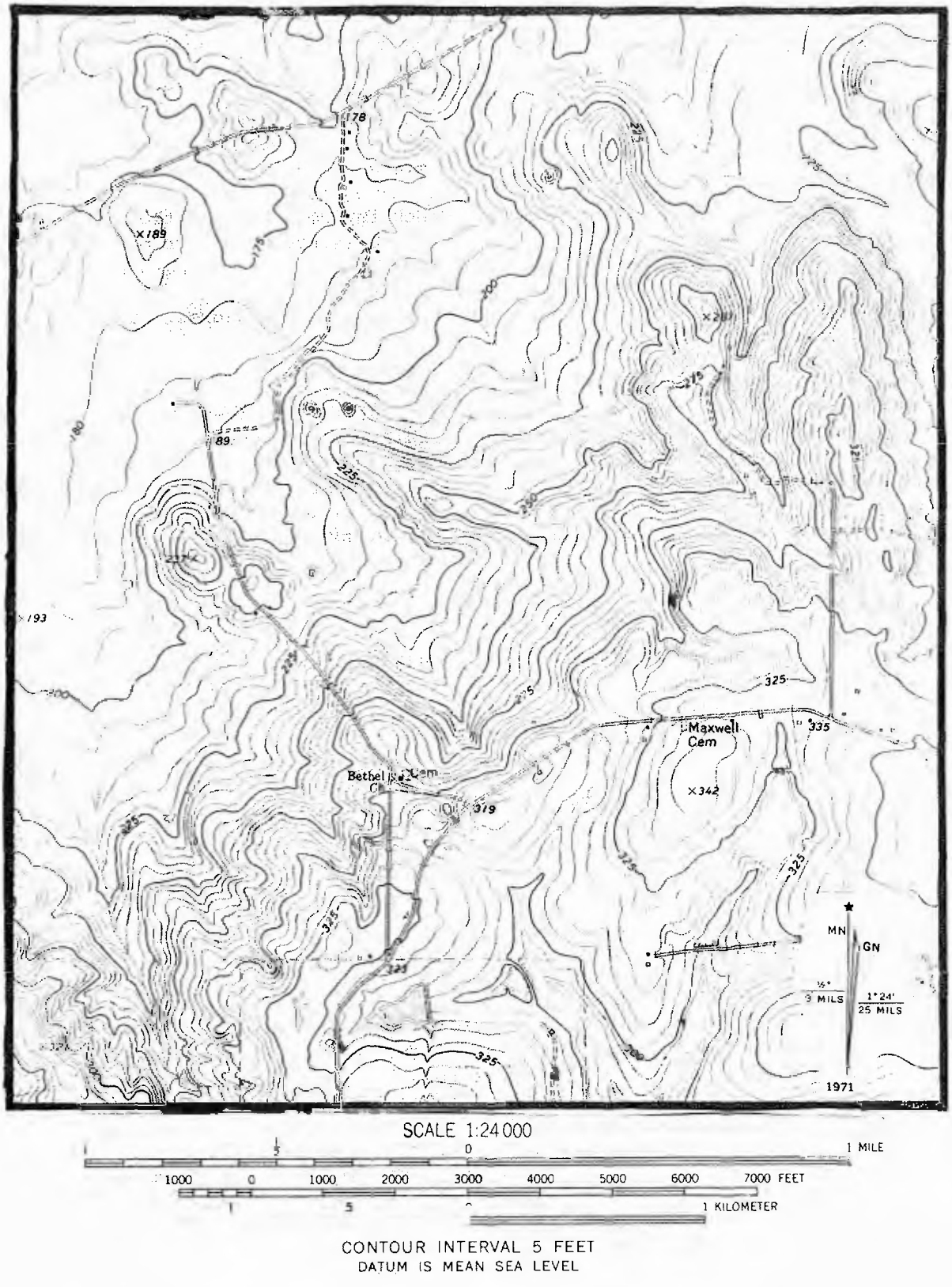


Figure 7: A representative section of the Pelham Escarpment. Tifton Upland to the southeast, Dougherty Plain to the northwest. (U.S.G.S. Gee Pond, Ga., 7½' topographic quadrangle map.)

Figure 8 (next page): Profiles across the Pelham Escarpment.

A. From Sale City along a straight line to the northwest.

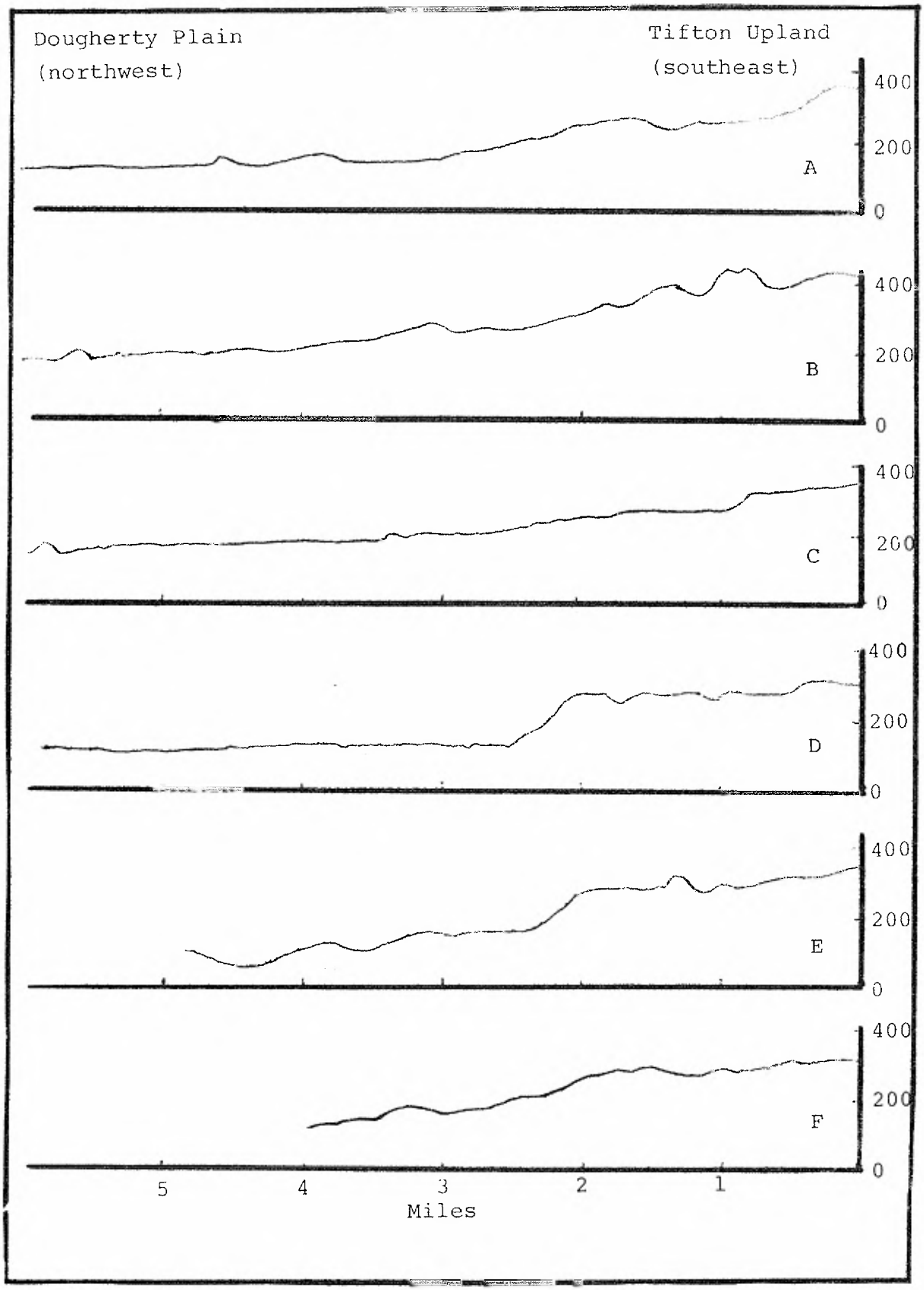
B. From Bridgeboro along a straight line to the northwest.

C. From Pelham along a straight line to the northwest.

D. From the peak of the ridge near Fowlstown, to the north along GA 309 and following the interfluve to Fowlstown Swamp.

E. From Climax to the northwest along an interfluve.

F. From a hill southeast of the intersection of GA 262 and GA 312, to the north along GA 312 and then following the interfluve to Open Pond.


Dougherty Plain
(northwest)
Tifton Upland
(southeast)
400
200
0
A
B
C
D
E
F
5
4
3
2
1
Miles

The more extensive body of hydrogeologic and geomorphic data available today contradicts this interpretation somewhat. Certainly the escarpment is being eroded and retreating down-the-dip. However, there is no evidence that karst springs exist, or did formerly exist along the Pelham Escarpment. In fact, Herrick and Legrand (1964) present no physical evidence of the former existence of those springs. They do explain (p. 33) that these hypothetical springs would only flow when the potentiometric surface beneath the Tifton Upland was higher than the water table below the Dougherty Plain, thus causing groundwater to flow northwestward beneath the upland toward outlets along the Pelham Escarpment where the limestone is unconfined. However, it would appear that any increased precipitation during Pleistocene pluvial periods would cause the water level in the Dougherty Plain to become relatively higher with respect to the groundwater underlying the Tifton Upland, because infiltration is virtually 100 percent over the Dougherty Plain but must be substantially less beneath the Tifton Upland where surface run-off and drainage are significant.

Further, the caves developed along the Pelham Escarpment (see below) are influent, not effluent; that is, they pirate surface drainage underground. None of them show any evidence of former surface outlets, such as scalloped passage walls or upward trending passages plugged by residuum. Rockhouse Cave, near Cordele, is the swallow hole of a semi-blind valley -- obviously influent. Waterfall Cave is the partially plugged drain of a swallow hole receiving the drainage from a small but significant stream -- again influent. Climax Cave is considerably more extensive and complex than the others and is discussed in more detail below; however, the present entrance to Climax Cave is a sinkhole, the bottom of which is a vertical shaft, analagous genetically to the vertical shafts described by Pohl (1955) in Kentucky. Vertical shafts are sharply grooved by the dissolving action of agressive water descending abruptly into the limestone.

In the Mammoth Cave area of Kentucky, water running off the sandstones which cap the Dripping Springs Escarpment infiltrates along the rim of the escarpment into the underlying limestone and has created numerous vertical shafts. These vertical shafts contribute to the dissection and retreat of the escarpment edge.

Pocket valleys are described by Sweeting (1973, p. 113) as "valleys with steep walls, and a steep abrupt cliff at their head ... a stream normally issues from its foot." It is the stream, in fact, which is responsible for undercutting the slopes and creating such a steep valley head. A representative portion of the Pelham Escarpment is shown in map view in Figure 7; several profiles across the Escarpment are shown in Figure 8. Although present day modification of Pleistocene topography is to be expected, there is no significant evidence that any of the short valleys here should be described as "pocket valleys." The majority of the profiles shown are very gradual and do not show any evidence of an oversteepened back-slope such as might indicate former undercutting. While two profiles do show an oversteepened back-slope, and this might be taken as some evidence in favor of undercutting by springs, this is the normal profile of a cuesta escarpment.

It is the present authors' conclusion that the Pelham Escarpment is indeed a retreating escarpment, the erosion of which is being aided by karst processes. However, it is dissection by vertical shafts and enlarging swallow holes which is aiding the normal surficial erosion, not undercutting by karst springs.

The Caves

Four significant caves with lengths in excess of 500' exist along the Pelham Escarpment: Climax Cave, Glory Hole, Waterfall Cave, and Rockhouse Cave (Figure 1). Boyer (1975) studied the geology of Climax Cave. Huddlestun (personal communication, 1976) measured the stratigraphic section in the entrance sink and both authors concur on the stratigraphic sequence (Figure 9). The uppermost, sloping portion of the sink is developed in the poorly consolidated sandy clay of the Hawthorne Group. The upper 25' (approximately) of limestone which forms vertical walls is dolomitic and impure and is designated as Chattahoochee Formation. The Suwannee Limestone begins approximately 5' above the first obvious "bench" in the entrance pit. Within the cave Boyer (1975) reports fossils of Clypeaster sp., Pecten sp., a tooth from Carcharadon sp., and Lepidocyclina sp.: an assemblage which he designates as Oligocene. This data, and a cross-section prepared by Sever (1965), indicate that the major horizontal development of Climax Cave is in the Suwannee Limestone, approximately 45' below the top. The primary level of horizontal development in Waterfall Cave is below 30-40' of limestone composed dominantly of algal balls and green clay. This is obviously the same unusual facies of the Suwannee Limestone as is seen in the quarry near Bridgeboro (see Stratigraphy of the Dougherty Plain). The geomorphic position of the other caves along the slopes of the Escarpment, along with their limited depth, indicates that they may be developed at a similar level.

Rockhouse Cave near Cordele is geomorphically simple. It marks the swallow-hole (ponor) in a semi-blind valley. Because of its limited size, this feature does not appear on the 7½' topographic map of the area (Penia Quadrangle). Rockhouse Cave has two entrances, apparently marking the upstream reatreat of the sinking point of a small stream. That is, the more westerly downstream entrance is higher and probably marks the original sinking point of a stream. Later, a second subterranian piracy developed further upstream as erosion removed the insoluble cover and entrenchment proceeded (see Figure 10). Solutional fluting of the present entrance demonstrates that aggressive water flowing off the clastic Hawthorne Formation is still actively dissolving the limestone.

Inside Rockhouse Cave few features of note exist. Passages are generally small, but higher than wide. Recent leaf detritus on the ceiling is evidence of total flooding. Frequent flooding is the apparent cause of the absence of any speleothem growth. One section of passage is perched on a large, irregular nodular chert layer that supports a waterfall. The flow at the western and lowest end of the cave descends deeper into the aquifer via impenetrable vertical cracks. Ponding due to the limited capacity of these cracks has caused the terminal area of the cave to act as a sediment trap, where large banks of coarse sediment have been deposited.

Climax Cave, Glory Hole, and Waterfall Cave show many similarities (Figures 11, 12, and 13). All three caves are entered via steep sinkholes which receive local drainage. The sinks at Climax Cave and Waterfall Cave show vertical solutional grooving indicative of enlargement by agressive descending water.

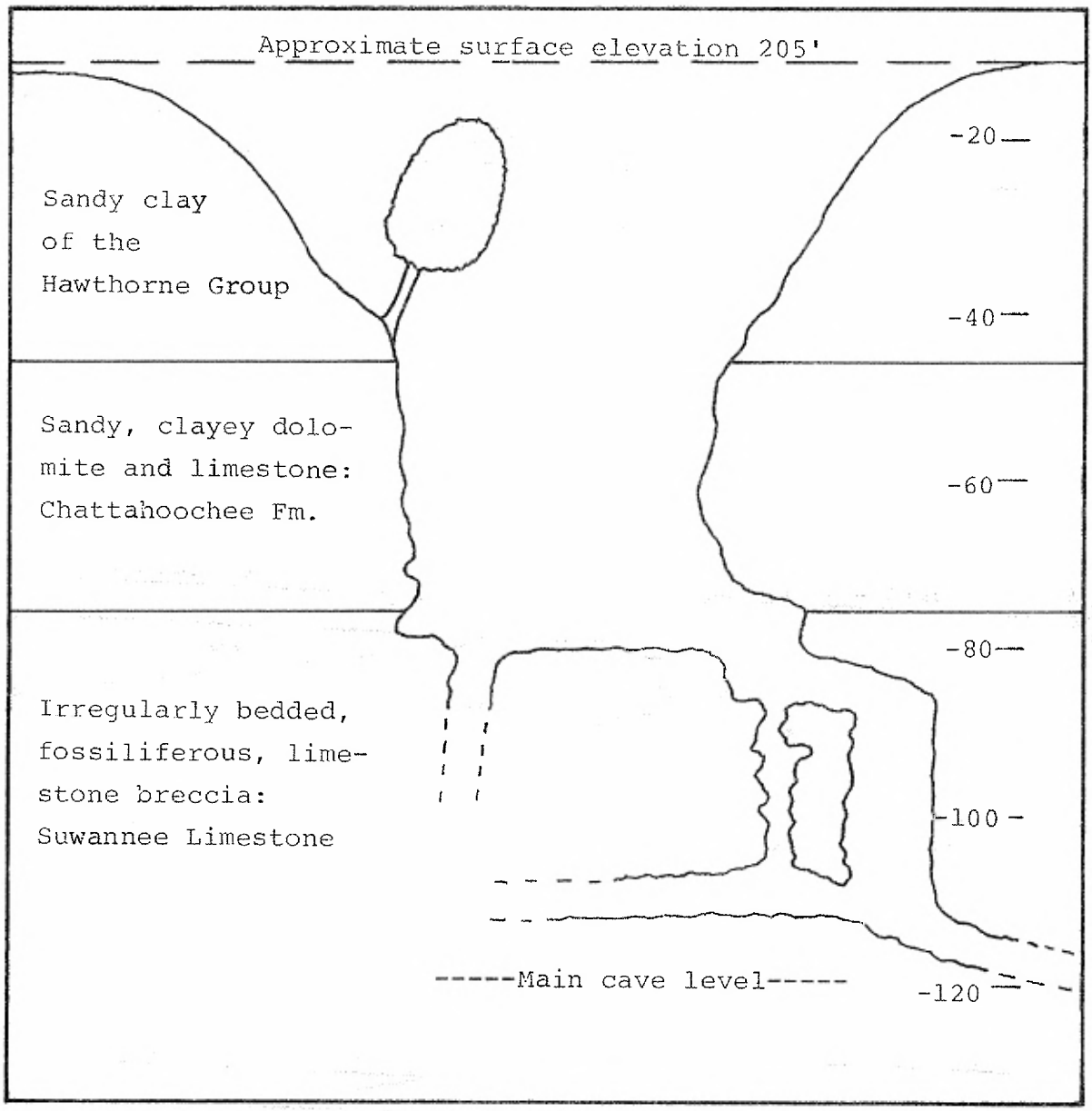


Figure 9: Geologic section at the entrance to Climax Cave. Compiled from Boyer (1975) and Huddlestun (pers. com., 1976).

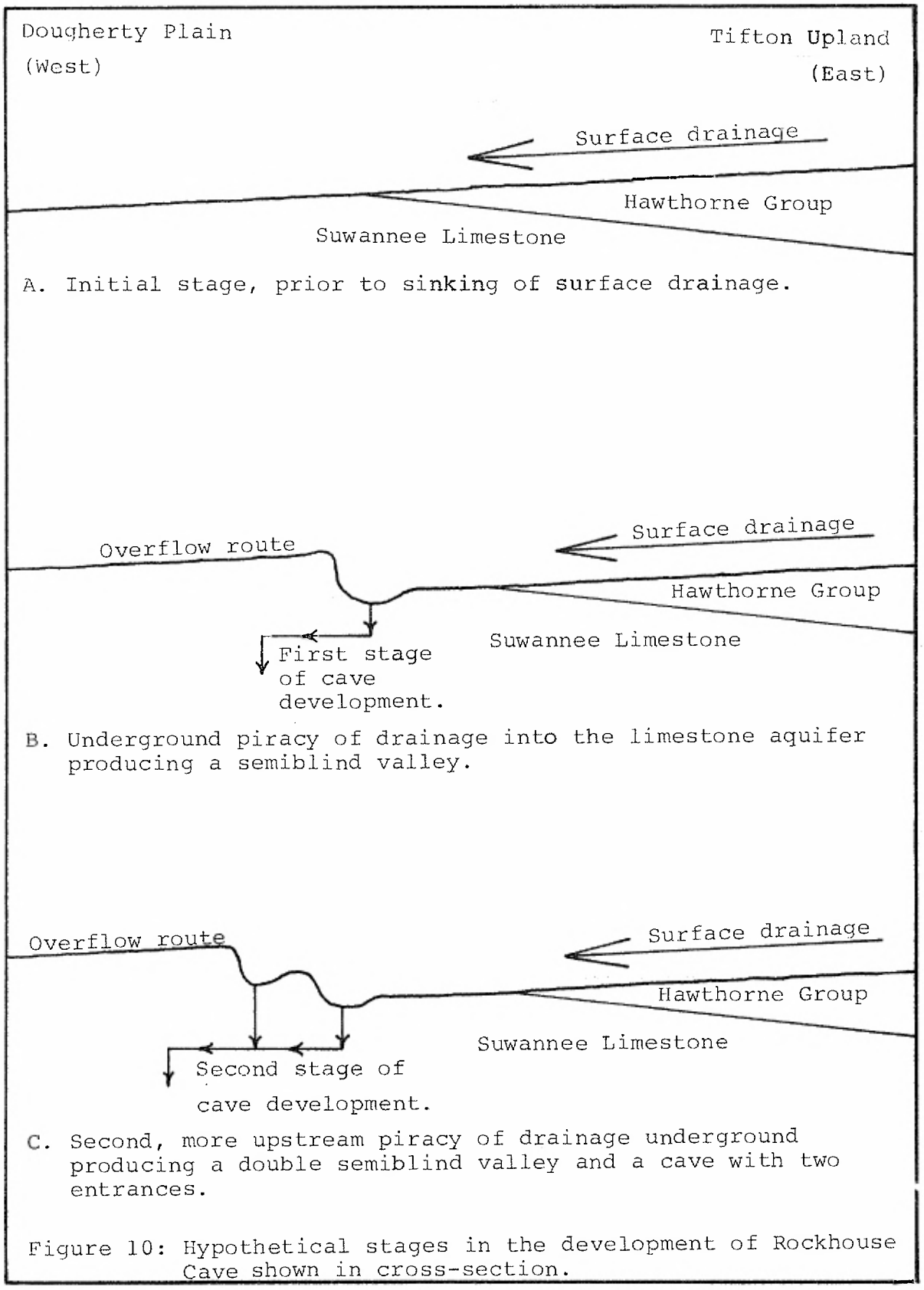


Figure 10: Hypothetical stages in the development of Rockhouse Cave shown in cross-section.

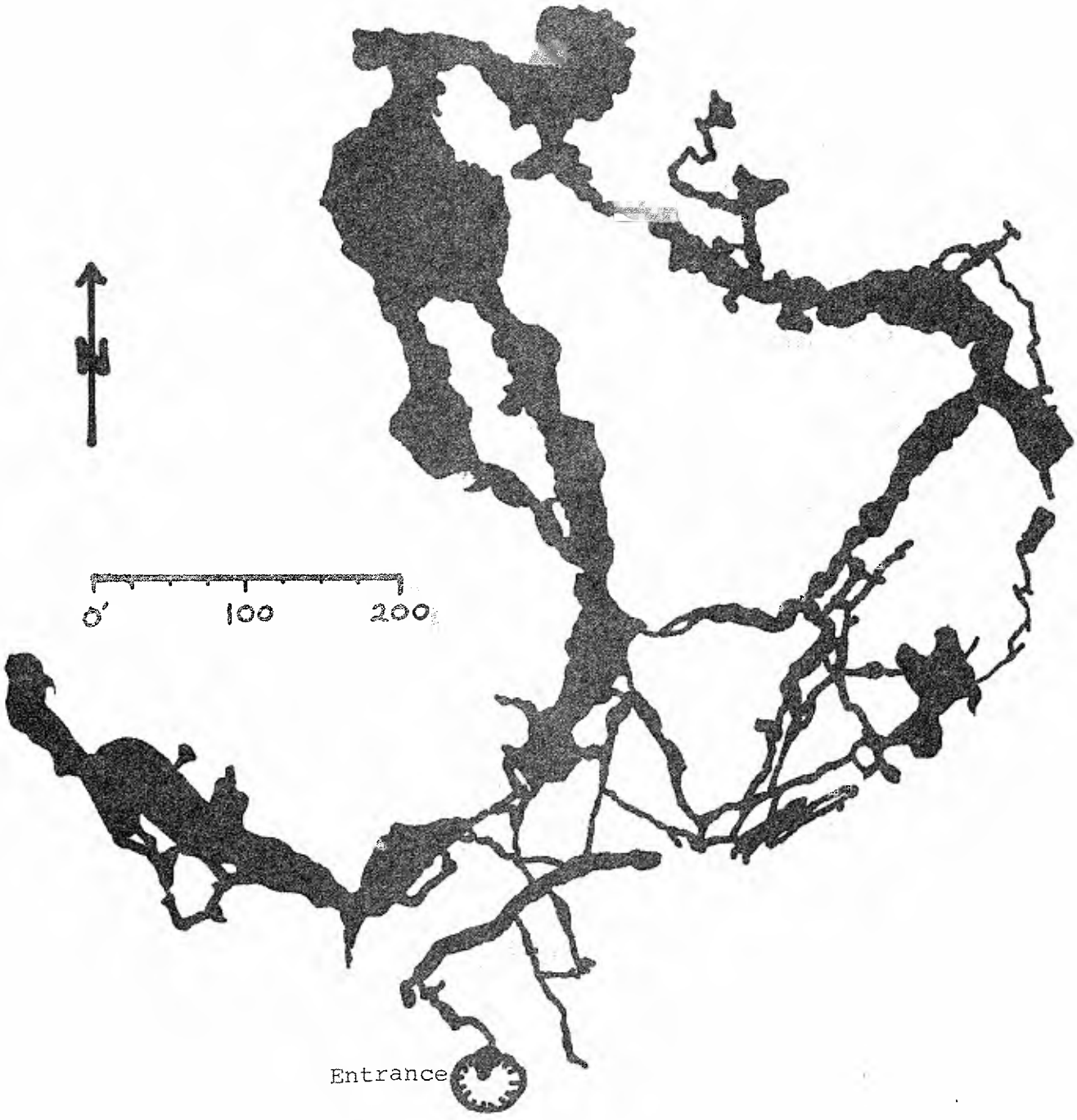


Figure 11: Passage outline map of Glory Hole (Glory Hole Cave), Grady County, Georgia. This cave is gated and access is available only through the Florida State University Caving Club. Map prepared by the Florida State Caving Club and the Fort Rucker-Ozark Grotto of the National Speleological Society, 1976.

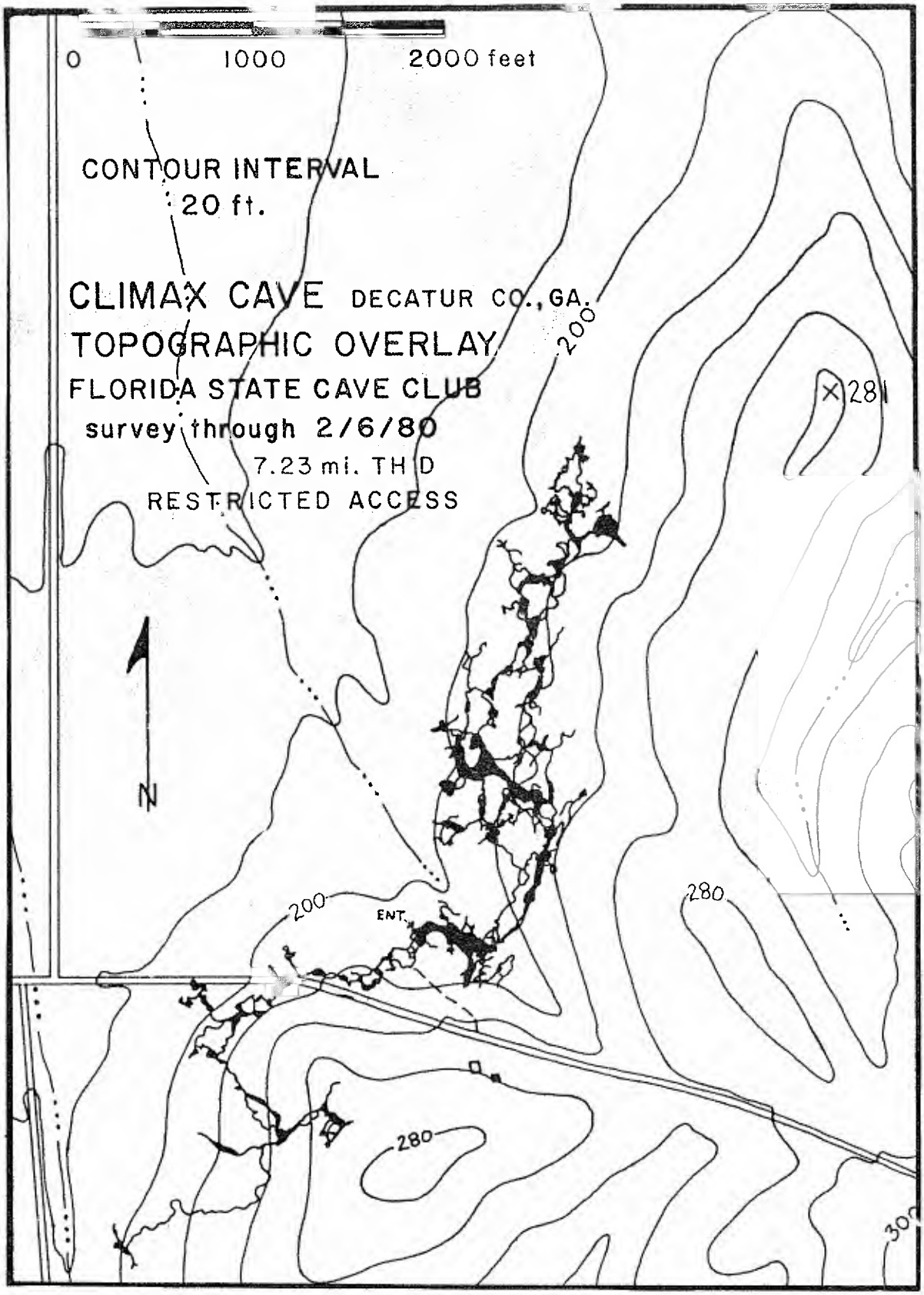


Figure 12: Climax Cave topographic overlay.

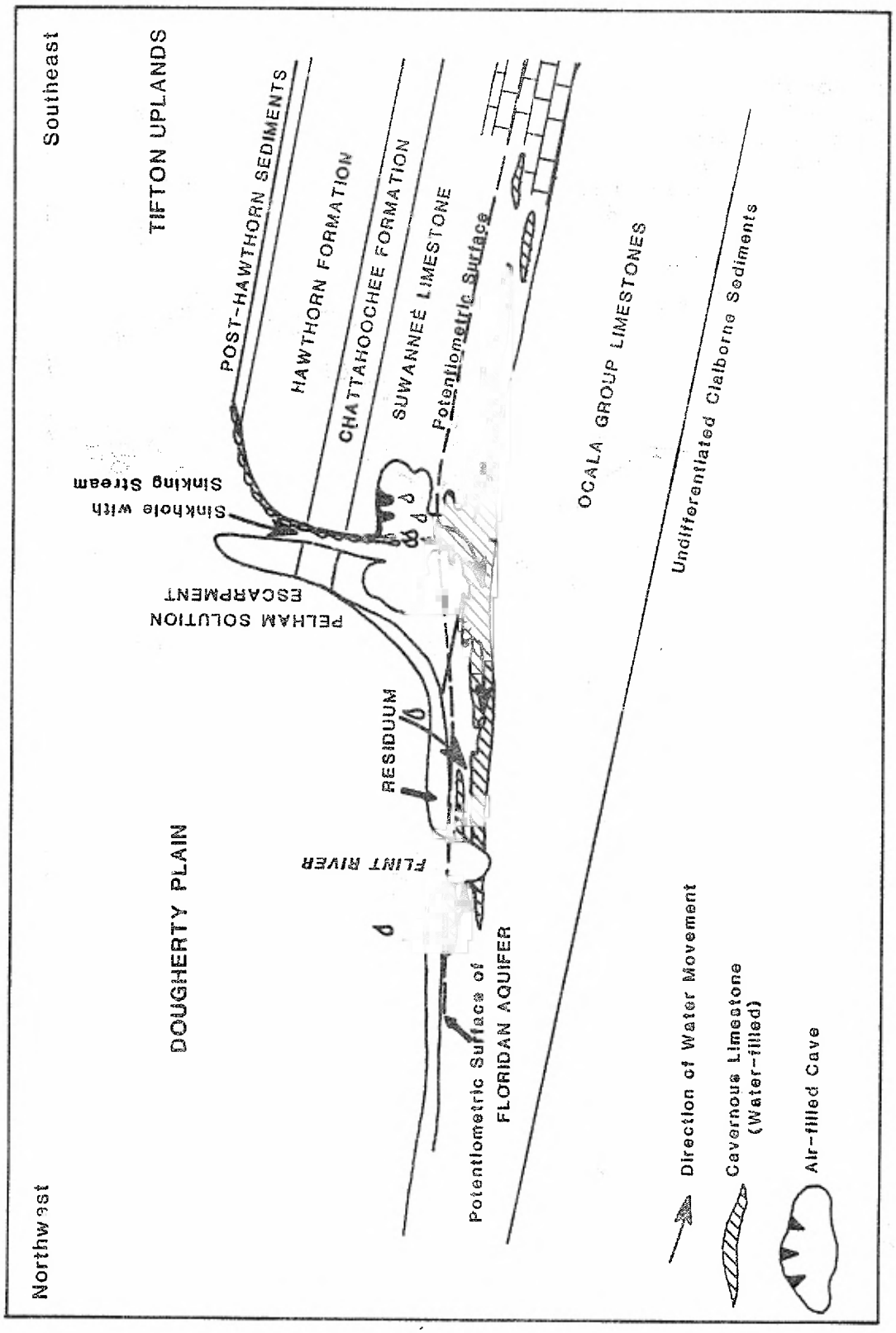


Figure 12B: Cross-section of the Pelham Escarpment and the Dougherty Plain showing the development of Climax Cave and the associated hydrogeologic relations (Korosy, 1984).

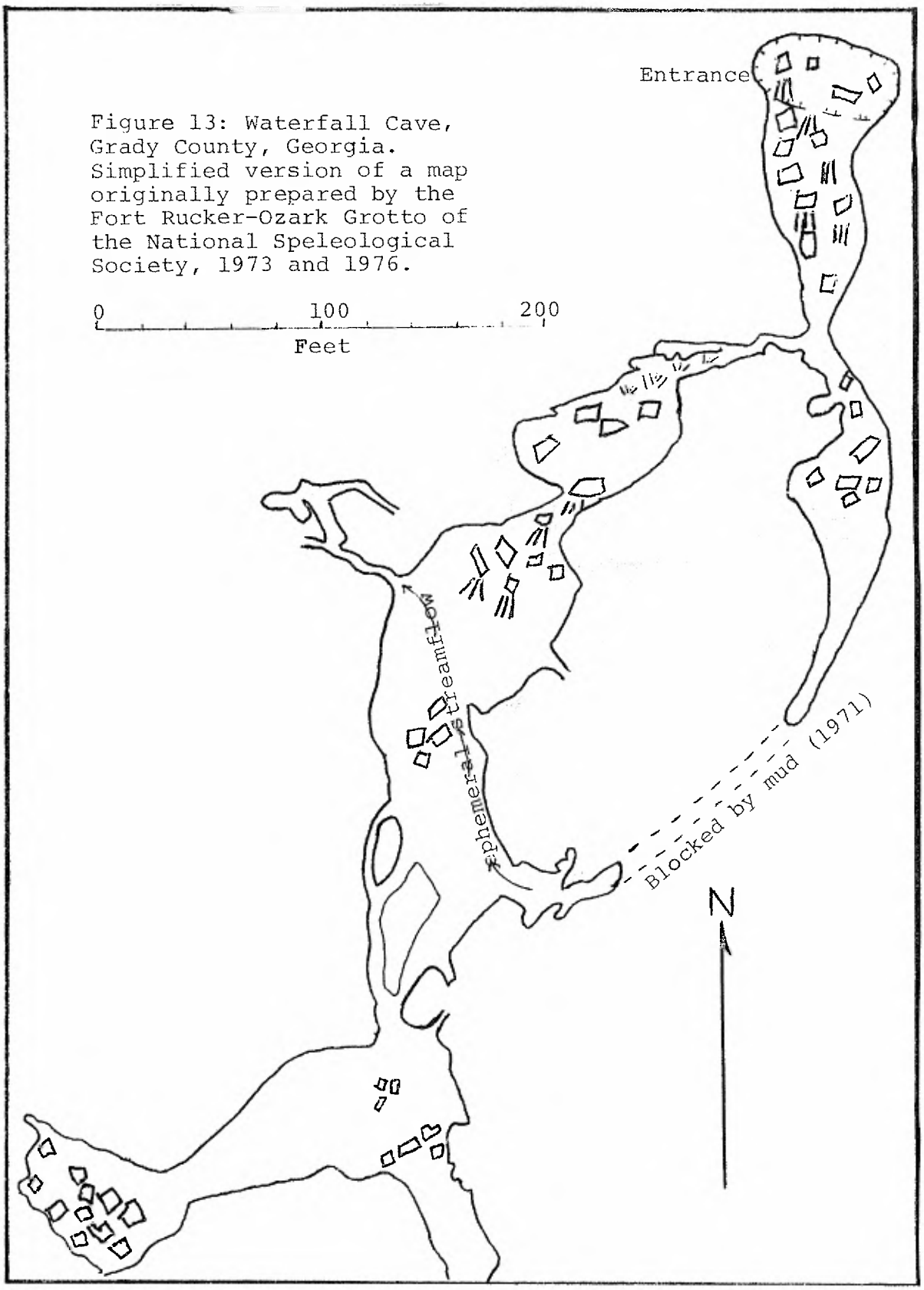


Figure 13: Waterfall Cave, Grady County, Georgia. Simplified version of a map originally prepared by the Fort Rucker-Ozark Grotto of the National Speleological Society, 1973 and 1976.

The sink at Glory Hole is filled with rubble and the limestone walls are not visible. In all three caves the entrances conduct drainage underground, but the caves do not appear to have formed solely as a consequence of the water draining into the present entrances. In Climax Cave there are at least twenty additional vertical shafts (Korosy, 1984).

In many areas the walls and smaller passages of these caves are a spongework, although the overall geometry of the larger passages appears to have an anastamotic maze pattern (terminology after Palmer, 1975, p. 58). Nevertheless, a computer analysis of passage trends has shown that all three of these caves are joint controlled by joints trending approximately N41°E and N42°W (Elwood, 1973). None of these caves show any evidence of directed waterflow, except for late stage vadose infiltration. In Climax Cave numerous clear pools are encountered, all at approximately the same elevation. Additional passages have recently been mapped below the water level. None of these pools show any detectable current. The water level in all the pools appears to fluctuate uniformly and does not respond rapidly to rainfall. No increase in turbidity is noted after heavy rainfall. The cave is simply coincident with a normal, diffuse water table developed in a zone of macroporosity (see Figure 12B).

Climax Cave and Glory Hole both contain extensive layers of fine-grained sediment. Coarser clastic sediments are only present near the entrances. Currents were apparently negligible during the deposition of most of the clay and silt since Elwood (1971) was able to obtain an interpretable paleomagnetic curve from the sediments. An average sedimentation rate of 6.9 centimeters per 100 years was calculated for a sequence of laminae deposited over a 600 year period (ibid., p. 308). The deposition of such fine-grained sediment is further evidence of a normal diffuse water table with no directed currents.

Climax Cave and Glory Hole both also contain extensive areas of finely textured speleothems, notably helictites, anthodites, and spar crystals. The density, delicacy, and variety of speleothems in Glory Hole is especially unusual. To preserve these from vandalism and destruction, the cave is gated and locked. Access may be obtained only through the Florida State Univeristy Grotto of the National Speleological Society. Future visitors to any of these caves are reminded that any collection or destruction of speleothems is not only contrary to conservation principles, it is also against the law in Georgia.

Climax Cave is considerably more extensive (more than 7 miles of passage) than Glory Hole or Waterfall Cave. A topographic overlay (Figure 12) shows that it is developed parallel to the edge of the escarpment and conforming to the local topography. Note that the major elongation of Waterfall Cave (Figure 13) is also along the escarpment. Palmer's (1975) model of maze cave formation by diffuse infiltration beneath a permeable caprock would not apply to these caves because they are developed far (circa 70 feet) below the clastic/limestone contact.

Is the position of the groundwater divide virtually under the escarpment, pertinent to the cave development? Probably not. It would be reasonable to assume that there would always be a groundwater divide under the escarpment edge. In the Dougherty Plain groundwater levels should rise with increasing distance from the Flint River which is the local base level. However, beneath the Tifton Upland, where infiltration must be less because surface run-off is greater, the groundwater level should be lower if all other factors are similar. Thus, the position of this high near the escarpment edge has nothing to do with cavern development.

These three caves (Climax Cave, Glory Hole, and Waterfall Cave) appear to be formed by the concentrated infiltration of aggressive water through vertical shafts along the escarpment edge. During periods of base flow for surface runoff the influent water is probably near saturation by the time it reaches the water table. However, during peak flows, such as are caused by the locally ubiquitous summer thunder storms, large volumes of highly unsaturated water must recharge the water table creating a localized zone of aggressive groundwater (Thrailkill, 1968). Within this zone the pre-existing macroporosity is further dissolved into a spongework of enterable proportions. The spongework character itself is evidence that no substantial current flow was involved in the speleogenesis.

Big Slough

Big Slough, as mentioned previously, is shown on most small scale maps as a permanent third order stream, tributary to the Flint River. It is so shown on Figure 1; it is the stream running along the base of the escarpment. However, Big Slough as shown does not exist. The topographic feature called Big Slough consists of a linear series of karst features, in some areas linked by short surface streams.

The course of Big Slough runs southwest from Camilla as a broad, generally bankless, low swampy channel which may carry water in rainy periods. However, the entire flow disappears into a large pond known as Medicine Sink, approximately five miles southwest of Camilla (follow the description on Figures 14 and 15). The linear, "valley-like" low trend continues to the southwest, but ephemeral drainage in this area is actually flowing to the northeast ("upstream)" into Bay Pole Sink (Figure 14). In fact, a large area of ephemeral surface drainage which is supposedly tributary to Big Slough actually drains into this sink. Immediately southwest of this area there is a broad flat-bottomed sink known as Gee Pond. Gee Pond is not actually in the course of Big Slough, but breaches the "valley wall" on the northwest side (Figure 14). The intermittant drainage into Gee Pond flows perpendicularly across the channel of Big Slough. To the southwest of this area, for several miles, Big Slough is shown as a dry valley periodically punctuated by sinkholes. This is the area where Big Slough crosses the E-W portion of Georgia Highway 262.

For approximately three to four miles south of this road crossing, Big Slough flows in a well-defined, v-shaped valley with short, equally well defined tributary valleys. (See the eastern part of Figure 15.) However, this normal fluvial surface drainage sinks underground at Big Rock Sink. If Big Rock Sink overflows, the drainage continues a half mile further and sinks into Little Rock Sink. Beyond this point, the low area called Big Slough widens markedly into a series of very broad (circa one mile), shallow, flat-bottomed sinkholes which have coalesced into a linear array. Classically, this is known as a uvala. An obvious example of the individual members of this array is the nearby sinkhole known as The Level (see Figure 15).

Why do all these karst features exist in a NE-SW linear array along the trend known as Big Slough? Note that the course of Big Slough is approximately parallel to the Pelham Escarpment and to the Flint River. However, the topography to the east of the slough is distinctly different from that to the west. From Big Slough to the Flint River, surface drainage is virtually absent, while between Big Slough and the Pelham Escarpment numerous small streams, ponds, and swamps are present. This is quantitatively demonstrated by calculating the drainage density, herein calculated as the total length of all streams

TABLE III: Karst Indices for the Southeastern Dougherty Plain

	NW			SE
Drainage density, km/km^2	0.035	0.028	0.74	0.97
Sinkhole density, % area	18%	24%	1.2%	1.8%
Total area measured	30 km^2	22 km^2	15 km^2	23 km^2
	Flint R.		Big Slough	Pelham Escarpment

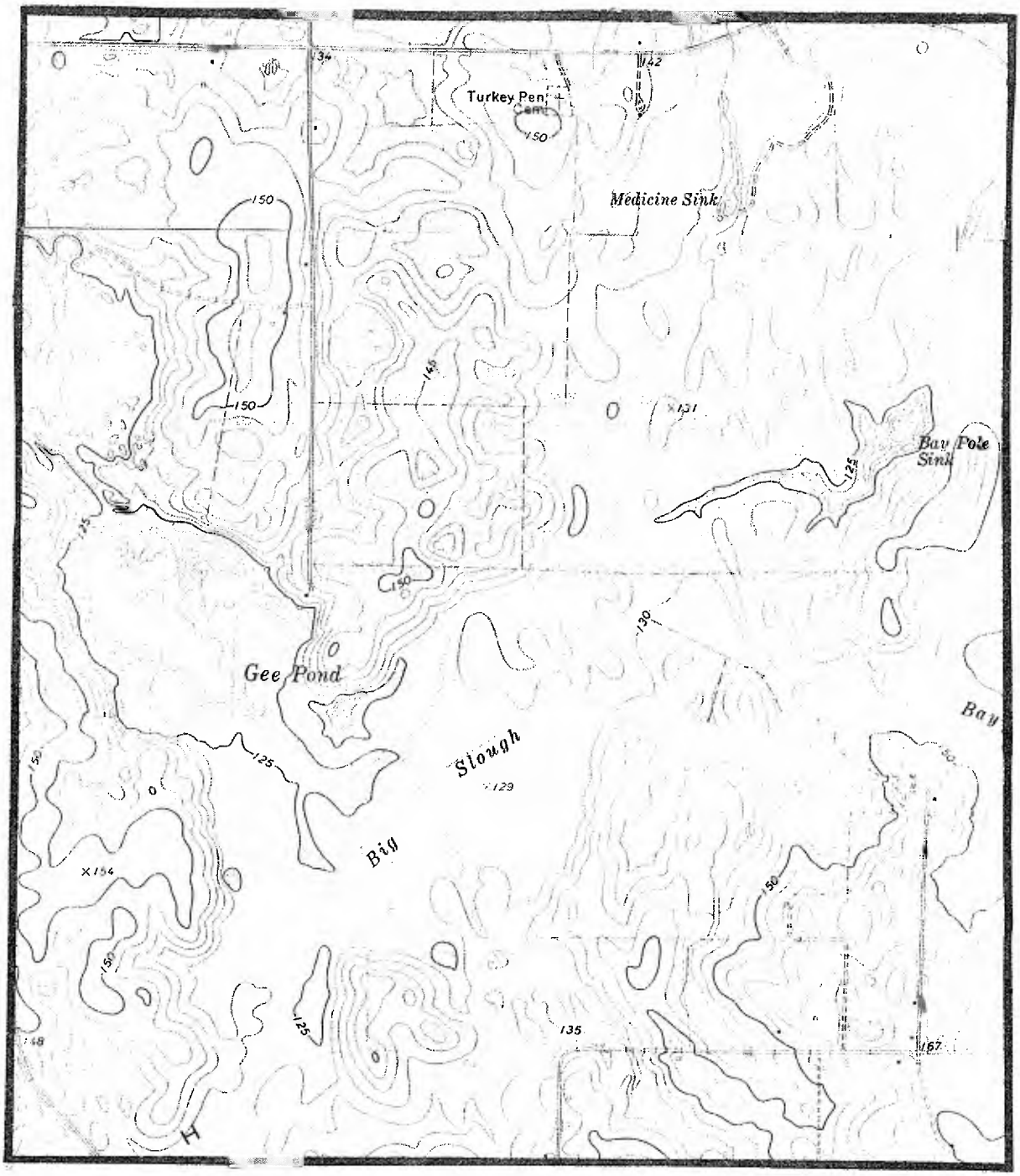
(including intermittent reaches) and ponds and swamps (the latter two being measured along their long dimensions) per area of land surface. Measured values are shown in Table III. Four separate areas were measured in a line from the Flint River to the Escarpment, two on each side of Big Slough. The sinkhole density was also calculated for the same areas: the percent of the land surface within hachured contours. While there is no trend apparent, there is an obvious major demarcation between the two sides of Big Slough. The Big Slough apparently marks some type of a boundary.

An earlier study of this feature (Beck, 1980a) noted that the course of Big Slough is very close to the Ocala-Suwannee contact, beneath the residuum (see Figure 1). This contact is unconformable and a zone of macroporosity which may control the geomorphology (see Hydrology). Thus, it was postulated that these karst features were localized along this contact. To the northwest the land surface is underlain by Ocala Limestone, to the southeast by Suwannee Limestone, thus accounting for the distinctly different topography and drainage.

However, there is also an alternate hypothesis which is applicable. The course of Big Slough is approximately equidistant from the Escarpment. It could also be localized along a former position of the Escarpment where a momentary pause in the down-dip retreat caused more extensive development of influent karst features along its base. The land northwest of that position would have been exposed to the surface elements for a much longer period, thus more thoroughly developing subsurface drainage. The land surface more recently exposed as the escarpment retreated downdip would have been subjected to solutional attack more briefly, thus leaving more vestiges of surface drainage.

Figure 14 (facing page): A portion of the Big Slough. From the northeast the Slough flows into Medicine Sink. Note that the drainage in the dry valley of Big Slough flows both northeast into Bay Pole Sink, and northwest, across the dry Slough, into Gee Pond.

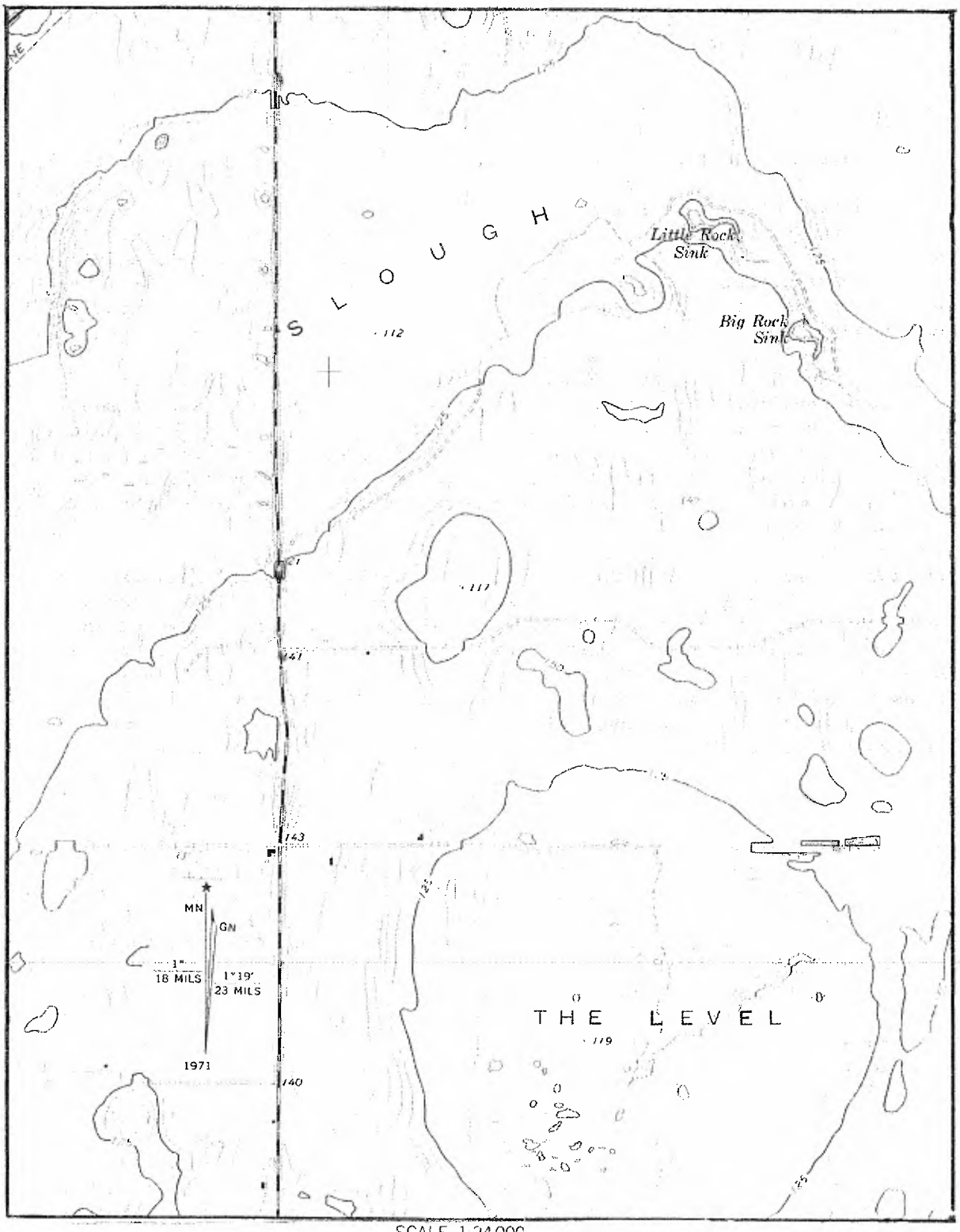
(U.S.G.S. 7½' topographic quadrangle maps for Gee Pond, Ga., and Branchville, Ga.)



True North is vertical. For a distance scale see Fig. 15, p. 41.

Figure 15 (facing page): A portion of the course of the Big Slough. Flow enters from the east in a "normal" fluvial channel and sinks at Big Rock Sink. In high flow Big Rock Sink may overflow into Little Rock Sink. Note that the course of the Slough beyond these two sinking points is a broad uvala composed of several coalesced sinkholes which individually were probably like The Level, in the southeast corner of the map.

(U.S.G.S. 7½' topographic quadrangle map for Vada, Ga.)


SLOUGH
Little Rock Sink
Big Rock Sink
THE LEVEL
MN
GN
1°
18 MILS
1°19'
23 MILS
1971
SCALE 1:24000
1 0 1 MILE
1000 0 1000 2000 3000 4000 5000 6000 7000 FEET
1 5 0 1 KILOMETER
CONTOUR INTERVAL 5 FEET
DATUM IS MEAN SEA LEVEL

At present, both hypotheses (stratigraphic control versus control by an erosional stillstand) appear tenable. Additional quantitative landform data may help to resolve this question.

Sinkholes

The surface landforms which are most characteristic of the Dougherty Plain are its ubiquitous sinkholes, or dolines. To be more specific, most of these should be called subsidence dolines. The mode of development of these features is not unique to the Dougherty Plain. Many sinkholes in Florida develop similarly, and the shakeholes in England are subsidence dolines developed in glacial debris overlying limestone. A generalized description given by Jennings (1971, p. 126) aptly describes the development of subsidence dolines (sinkholes, limesinks) in the Dougherty Plain.

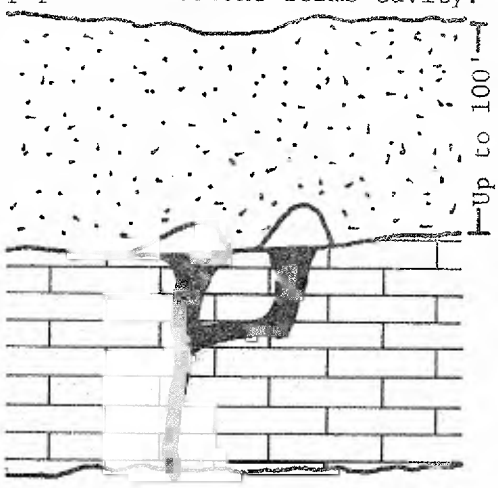
> "Subsidence doline. Where superficial deposits or thick residual soils overlie karst rocks, dolines can develop through spasmodic subsidence and more continuous piping of these materials into widened joints and solution pipes in the bedrock beneath. They vary very much in size and shape. A quick movement of subsidence may temporarily produce a cylindrical hole which rapidly weathers into a gentler conical or bowl-shaped depression."

Sweeting (1973, p. 59) describes similar features as "alluvial dolines", but implies that the English shakeholes, which are her primary illustration of these features, are the result of slow, continuous subsidence rather than collapse. Dolines in the Dougherty Plain, and to the south in the Floridan karst, are very frequently the result of a sudden collapse of the overlying, unconsolidated residuum. For this reason, and because the adjective "alluvial" seems misplaced and misleading, the term subsidence dolines is preferred for these features. Engineers refer to them as ravelling sinks (Jammal and Assoc., 1982).

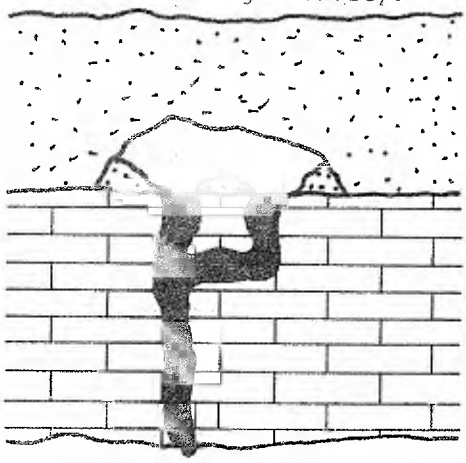
The catastrophic surface collapse of these sinkholes is only the final stage in a gradual, continuous process which has been occurring out of sight, underground prior to its sudden surface manifestation (see Figure 16). Initially, water infiltrating into the residuum percolates downward to the underlying limestone surface. The water moves laterally, perched on the solid limestone until it encounters an opening: usually an enlarged joint or the intersection of two joints. The water then descends deeper into the limestone through a network of joints, bedding planes, and porous zones. Over a period of time this downward path becomes enlarged by the solution action of the descending water acidified by carbon dioxide from the soil zone. The solution process is slow and enlargement to a significant size probably requires tens of thousands of years, or longer. A vertical conduit so formed along intersecting joints is referred to as a solution pipe (Jennings, 1971, p. 50).

When these solution pipes enlarge to a sufficient size, the turbulent flow of water draining into them will also wash sediment down the pipe, a process sometimes termed suffosion. Thus, a cavity will develop in the residuum directly over the active pipe. Over a period of time the roof of this cavity may crumble and this new debris will also be washed down the pipe, thus enlarging the cavity upward. This upward stoping collapse and suffosion continues until some time when the weight of the remaining residuum composing the cavity roof exceeds its cohesive strength and the roof suddenly collapses. This upward stoping process may be rapid or slow depending on the cohesion of the unconsolidated sediments.

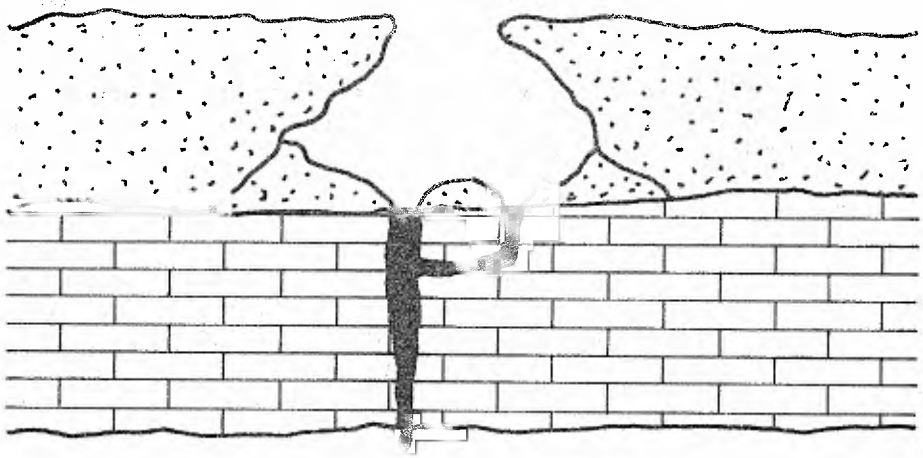
A. Suffosion through solution pipe in limestone forms cavity.



B. Upward stoping and continued suffosion enlarges cavity.



C. Continued upward stoping eventually breaches the land surface forming a young, steep-sided subsidence doline.



D. Continued surface erosion of the unconsolidated residuum produces a wider sinkhole with gently sloping sides.

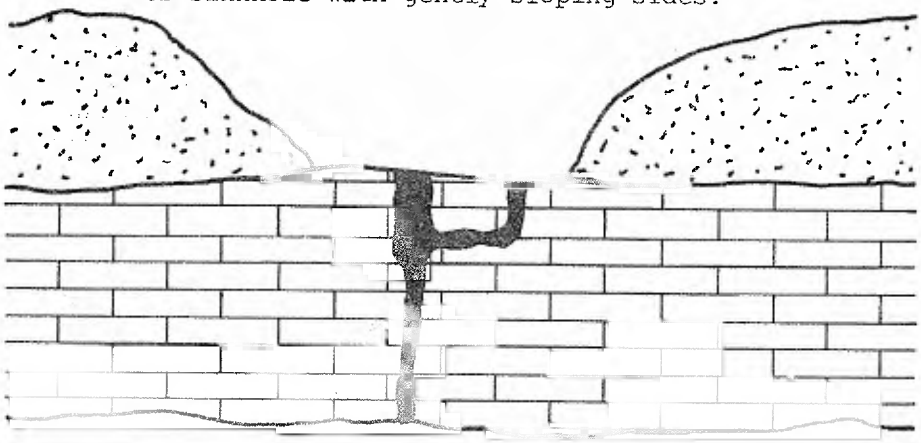


Figure 16: Development of subsidence doline.

While this upward stoping will eventually reach the surface, certain events or conditions, either man-induced or natural, may activate the final surface collapse. Three basic types of triggering mechanisms exist.

A period of unusually low groundwater levels, which allows the upper residuum to dry out, followed by heavy rains which saturate the surface sediments, may cause increased weight and lower cohesion in the cavity roof, thus activating collapse. Lowered groundwater levels may be caused by drought, heavy well pumpage, or mine dewatering; all have the same affect.

Any activity which concentrates an abnormally large volume of infiltration in one area will accelerate the suffosion process and may cause collapse to occur much sooner than it would naturally. The construction of lakes, ponds, or water treatment lagoons can concentrate infiltration as can modifications of surface drainage, or rupturing of water lines. When such a concentration of drainage results in sustained, high volume infiltration, the residuum may be literally flushed down the solution pipe as a slurry, and surface collapse may occur quite rapidly.

Unusual stresses which overload, or weaken, the sediment bridging an upward stoping cavity may activate immediate collapse. The construction of buildings or other large structures, or the impoundment of surface water in lakes or lagoons, may overload potential collapse sites. New lakes or ponds are particularly potent triggering mechanisms because they increase both infiltration and weight. Excavations, which reduce the roof thickness, or blasting which reduces roof cohesion, may also activate collapse.

Areas which are particularly prone to sinkhole formation are characterized by certain natural conditions. Zones of high porosity in the limestone are necessary for suffosion. A linear array of elongate sinkholes may indicate the presence of a controlling subsurface fracture. Obviously, areas along this fracture are prone to future collapse. Brook and Allison (1981) identify three factors which correlate with collapse prone areas. Troughs in the potentiometric surface may indicate zones of high permeability and increased likelihood of sinkhole formation. Areas where the groundwater level fluctuations are greatest will alternately wet and dry the residuum and accelerate collapse. Areas with thin residuum are more likely to develop sinkholes.

Just prior to the actual collapse, various associated surface features may develop. Cracking and settling of structures and the gradual development of a low area outlined by small circular fissures in the ground surface may preceed more serious events. Water in nearby wells may become turbid and muddy, particularly if the cone-of-depression caused by the well is accelerating the suffosion process. Finally, numerous small holes, on the order of 10 cm in diameter, may develop over the site of future collapse.

The actual collapse event initially forms a steep-sided or overhanging, sinkhole (see Figure 16). However, since the sides are made up of unconsolidated residuum, this will quickly evolve through normal erosion into a broader sink with gentler side slopes. Eventually, the continued erosion of the sides will produce a broad, saucer-shaped depression. Inflowing sediment may plug the solution pipe draining the sink and a lake may form. These sinkhole lakes frequently drain rapidly into the subsurface when this plug collapses or is washed away.

A few deep, steep-sided sinkholes with limestone walls are located along the base of the Pelham Escarpment, usually associated with caves. In the sinkhole at Climax Cave the removal of the limestone appears to be primarily the result of erosion and solution by inflowing surface water. However, at Glory Hole and Waterfall Cave, large amounts of collapsed limestone debris floor the sink and form breakdown mounds in the caves below. Along the escarpment, then, collapse of the limestone per se into underlying voids (caves) is part of the sinkhole forming process. However, it is not known whether this collapse occurred into a pre-existing cave system or whether the development and enlargement of the sink was an integral part of the cave's origin. The latter seems probable.

Springs and the Flint River

While numerous publications have documented the springs in Florida (for instance, Rosenau and others, 1977), springs in Georgia are virtually unstudied. All the groundwater flow in the Dougherty Plain eventually drains into the Flint River, either from springs along the Flint River itself or via springflow into the major tributaries: Spring Creek, Ichawaynochaway Creek, and Kinchafoonee Creek and their drainage systems. Big Slough, while a tributary, is influent, as discussed previously. Only in the lower reaches of the tributaries, near the Flint River itself, does spring flow occur. The streams in the lower order portions of the drainage network have wide, flat flood plains with low gradient, braided channels through wooded, swampy terrain. They are generally perched well above the limestone and do not directly interchange water with it.

The Flint River, and the higher order reaches of its major tributaries, are incised into the limestone. Wilson and Pickering (1973, p. 6-10) describe the Flint River thusly.

> "(The Flint River) flows from the Piedmont Province, near Atlanta, to join the Chattahoochee River near Bainbridge and then, as the Appalachicola River, to the Gulf of Mexico. (The Flint River is) characterized by year-round flow, deeply incised beds, strong natural levees, wide flood plains, and heavy flooding in the winter from a large Piedmont drainage basin. Floodwater is highly turbid, colored reddish tan to reddish brown. The river may recharge the upper, highly cavernous portions of the aquifer during flood, as evidenced by raising of water levels in sinks near the river after floods. At times of very low flow river water is clear, green, and alkaline. Limestone outcrops are exposed almost continuously along the banks at low water, and shoal areas are caused by resistant chert layers in the limestone prominent in the river bed. During low flow, numerous circular springs line the banks, and large boils in midstream indicate the presence of powerful underwater springs. (The Flint River) as seen on visible red band earth resources satellite photos, is little wooded, in contrast to other nearby large streams. In summary, (the Flint River) interchanges water with the aquifer, depending on river level and ground-water level."

The higher order reaches of the tributary streams also interchange water with the aquifer.

All of the springs along the Flint River (known to the author) flow from conduits in the limestone. Radium Springs and Berry Springs, just south of Albany, are good examples. Radium Springs does not flow directly into the river but forms a small rise pool approximately 0.1 miles east of the river (Fig. 17). From here a short run flows to the Flint. The location of Radium Springs appears to be on the bank of an abandoned channel of the river. It is fed through a large cavern (Fig. 18) which connects to a plugged sinkhole across the road from the spring. Flow from this spring was measured as high as 87 mgd on 4-19-1904 (Wait, 1963). During the early 1950's flow generally ranged between 15 and 50 mgd (ibid.), but during the summer of 1981 the spring ceased flowing for the first time on record.

Berry Spring, along the west side of the Flint River south of Albany (Baconton North Quadrangle), is in reality a karst fenster which functions as an estavelle. It is a funnel-shaped sink in residuum leading down to a vertical channel in limestone. During periods of moderate or low water it contains clear standing water and does not flow. It apparently is simply an opening into a subterranean conduit conducting water toward the river. A surface profile of electrical resistivity measured inland from the spring detected a strong anomaly near the spring, probably indicative of the path of this conduit. Between the spring and the river, another small vertical solution pipe can be found in the woods, apparently also marking the path of this conduit. During periods of high water this sinkhole overflows with muddy brown water, probably indicative of backflooding from the river. Because Berry Spring may be either influent or effluent, it is termed an estavelle.

Blue Hole Spring, on the east side of the Flint River near the power plant south of Albany, is also a fenster and an estavelle. Discharge may be through a spongework zone of megaporosity because divers have not located a conduit beneath this feature (Deloach, 1984, pers. com.). This "spring" is a swallow hole for the semi-blind valley of Dry Creek. It is a large bowl-shaped basin adjacent to the channel of Dry Creek and connected to it by a short diversion channel. The bottom of this basin contains several cavities penetrating limestone spongework and normally partially filled with clear water. Immediately adjacent to the Blue Hole, there is an upward step of several feet in the channel of Dry Creek. That is, in moderate flows Dry Creek runs toward the Flint River to this point and then flows into Blue Hole Spring to finish its path to the Flint via a zone of megaporosity in the limestone. In the event of heavy local precipitation, Dry Creek overflows this low bench and flows to the Flint via both surface and subsurface paths. During periods of high river stage, after local flow has dissipated, the Blue Hole Spring overflows brown muddy water from the river and backfloods the channel of Dry Creek.

Some of these subsurface conduits probably debauch directly into the bed of the Flint River. Wilson and Pickering (1973) show a photo of a large boil in the river surface at low flow marking strong spring flow. A seismic survey by Henry and Pickering (pers. com., 1976) located a large, deep sink in the bottom of the river, probably marking a point where the river and aquifer interchange water (Fig. 19).

However, the persistent and relatively uniform discharge of Radium Springs indicates that the spring has a large diffuse flow contribution. Divers have mapped the underlying conduit which appears to be a dendritic system growing from the spring outflow headward (Fig. 18). Because the river backfloods most of the other springs, the character of the flow is difficult to ascertain. Based on the nature of Radium Springs, and by analogy with similar terrain in Florida, it is hypothesized that these springs are predominantly fed by diffuse flow through the highly permeable limestone and that conduit development occurs at the outflow point due to the convergence of groundwater flow (Beck and others, 1984; Rhoades and Sinacori, 1941).

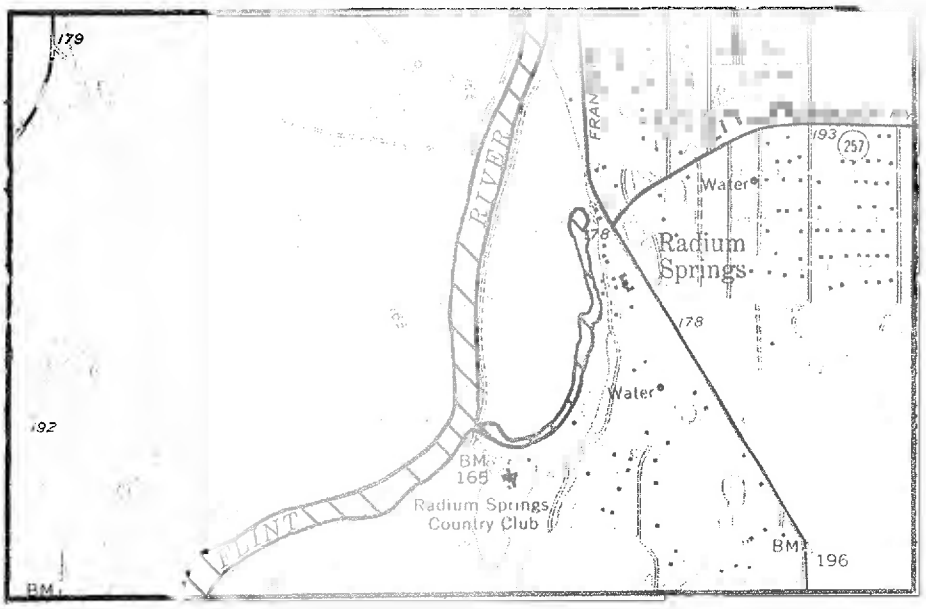


Figure 17: Map of Radium Springs area showing the spring, the Flint River, and the associated sinkholes. North is vertical; scale is 1:24,000.

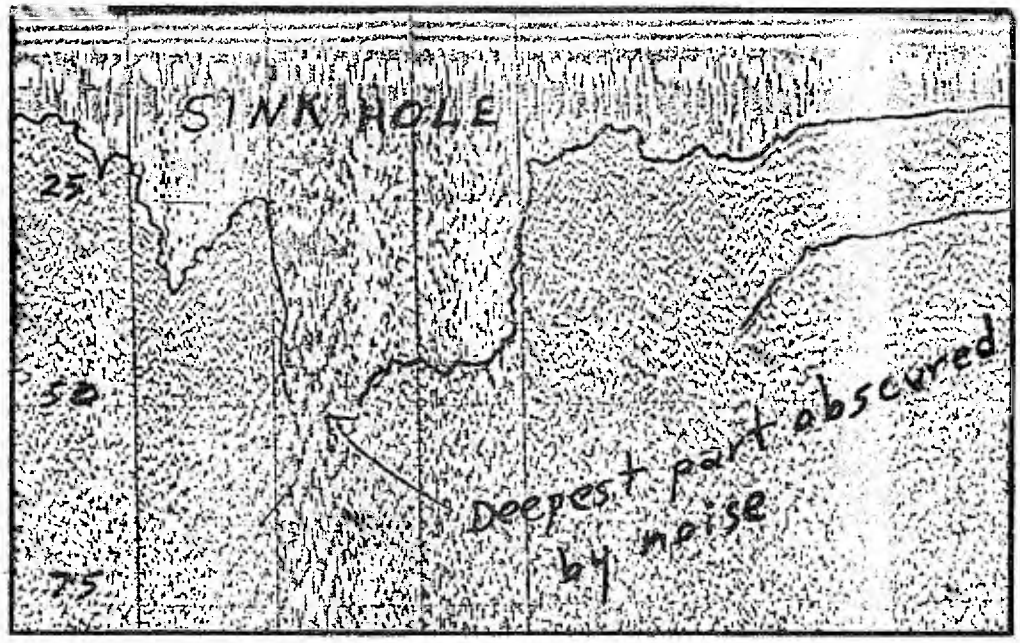


Figure 19: Field record of a seismic traverse on the Flint River showing a large sinkhole in the bed of the river. Original record by Vernon J. Henry and Sam Pickering.

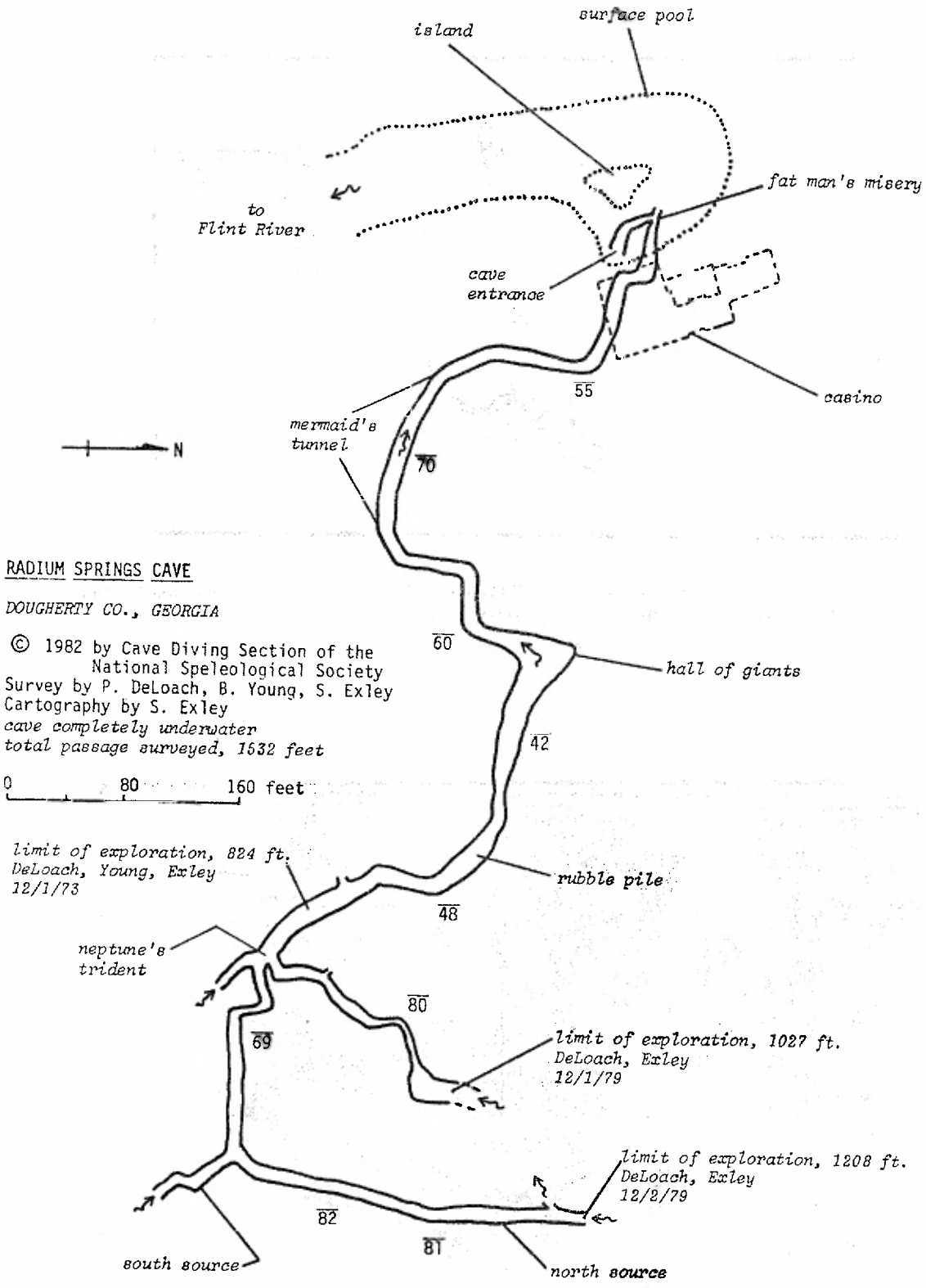


Figure 18: Map of Radium Springs Cave as surveyed by divers. The entire cave is water-filled. The location of the "rubble pile" coincides with a surface sinkhole across the road from the spring.

REFERENCES

Applin, E.E.R., and Applin, P.L., 1964, Logs of selected wells in the coastal plain of Georgia: Ga. Geo. Surv. Bull. 74, 229 p.

Beck, B.F., 1980, Karst indices, subsurface lithology, and the drainage of Big Slough, Southwest Georgia (abs.): G.S.A. Abstracts with Programs, V. 12, no. 4, p. 170.

-----, 1980a, An introduction to caves and cave exploring in Georgia: Ga. Geol. Surv., Geologic Guide 5, 43 p.

Beck, B.F., Ceryak, Ron, Jenkins, D.T., Scott, T.M., and Spangler, D.P., 1984, Field guide to some illustrative examples of karst hydrogeology in central and northern Florida: University of Central Florida, Florida Sinkhole Research Institute, Report 84-85-2, 43 p.

Bogli, A., 1964, Mischungskorrosion: ein beitrag zum verkarstungsproblem: Erd Kunde, V. 18, no. Z, p. 83-92.

Bowman, J.K., and Manker, J.P., 1982, Smectite-rich clays associated with Oligocene algal reefs (abs.): Ga. J. of Sci., V. 40, no. 1-2, p. 20.

Boyer, Paul, 1975, Unpublished manuscript on Climax Cave. (Dr. Paul Boyer, now deceased, was a geology professor at Okaloosa-Walton Jr. College, Ft. Walton Beach, Florida.)

Brook, G.A., and Allison, T.L., 1981, Subsidence susceptability models for Dougherty County, Georgia, from sinkhole and fracture distribution data: in Beck, B.F. (ed.), 1981, Proceedings of the Eighth Int. Cong. of Speleology, July 18-24, 1981, Bowling Green, Kentucky, p. 50-52; published by the Nat. Spel. Soc., Cave Ave., Huntsville, Al.

Burnson, Terry, 1981, Hydrogeology of the Suwannee River Water Management District: in Fisher, George (ed.), Groundwater in Florida -- Proceedings of the First Annual Symposium on Florida Hydrogeology, NW. Fla. Water Mgt. Dist. Pub. Info. Bull. 82-2, p. 51-64.

Clark, W.Z., and Zisa, A.C., 1976, Physiographic map of Georgia: Georgia Geol. Survey GM-8, one sheet.

Cooke, C.W., 1943, Geology of the coastal plain of Georgia: U.S.G.S. Bull. 941, 121 p.

Cooke, C.W., and Mansfield, W.C., 1936, Suwannee Limestone of Florida (abs.): G.S.A. Proc. 1935, p. 71-72.

Dall, W.H., 1892, Correlation papers - Neogene: U.S.G.S. Bull. 84, 349 p.

DeLoach, P., 1982, Map of Radium Springs Cave: prepared by the Cave Diving Sec. of the Nat. Spel. Soc., unpublished.

Ellwood, B.B., 1971, An archeomagnetic measurement of the age and sedimentation rate of Climax Cave sediments, Southwest Georgia: A.J.S., V. 271, p.304-310.

-----, 1973, Unpublished computer analyses of cave passage trends in Southwest Georgia. (Dr. Brooks Ellwood is a geology professor at the U. of Ga.)

Fisk, D.W., and Exley, I.S., 1977, Exploration and environmental investigation of the Peacock Springs Cave System: in Dillamarter, R.R., and Csallany, S.C. (eds.), Hydrologic Problems in Karst Regions, Bowling Green, Ky., West. Ky. U, p. 297-302.

Georgia Geologic Survey, 1976, Geologic map of Georgia: Georgia Dept. of Nat. Resources, Ga. Geol. Surv. Div., Atlanta, Ga.

Glawe, L.N., 1974, Upper Eocene and Oligocene Pectinidae of Georgia and their stratigraphic significance: Ga. Geol. Surv. Inf. Circ. 46, 27 p.

Hanshaw, B.B., and Back, W., 1979, Major geochemical processes in the evolution of carbonate aquifer systems: J. of Hydro., V. 43, p. 287-312.

Harrison, Kerry, 1981, 1981 irrigation survey: Coop. Ext. Serv. of U. Ga. Dept. of Ag. and the U.S.D.A., 4 p.

Herrick, S.M., and LeGrand, H.E., 1964, Solution subsidence of a limestone terrain in Southwest Georgia: Int. Assoc. of Sci. Hydro., V. 9, no. 2, p. 25-36.

Hickey, J.J., 1984, Field testing the hypothesis of Darcian flow through a carbonate aquifer: Groundwater, V. 22, no. 5, p. 544-547.

Hicks, D.W., Krause, R.E., and Clarke, J.S., 1981, Geohydrology of the Albany area, Georgia: Georgia Geol. Surv. Inf. Circ. 57, 31 p.

Jammal & Assoc., 1982, The Winter Park sinkhole: An investigative report for the city of Winter Park, Florida; privately published.

Jennings, J.N., 1971, Karst: Cambridge, Mass., the M.I.T. Press, 252 p.

Korosy, M.G., 1984, Ground water flow patterns as delineated by uranium isotope distributions in the Ochlockonee River area, Southwest Georgia and Northwest Florida: unpublished M.S. thesis, Fl. St. U., Dept. of Geo., 179 p.

Kwader, Thomas, and Wagner, J.R., 1980, Ground water resources of the Dougherty Plain, Southwest Georgia: private report prepared for Decatur County, Ga.; available from the authors, Box 10452, Tallahassee, Fl., 32302.

LaForge, Laurence; Cooke, Wyeth; Keith, Arthur; and Campbell, M.R., 1925, Physical geography of Georgia: Geol. Sur. of Ga. Bull. 42, 189 p.

LaMoreaux, P.E., 1948, Fluoride in the ground water of the Tertiary area of Alabama: Al. Geol. Surv. Bull. 59, 77 p.

-----, 1982, Groundwater from the coastal plain for the future: in Arden, D.D., Beck, B.F., and Morror, Eleanore (eds.), Proceedings of the 2nd Symp. on the Geol. of the S.E. Coastal Plain, Ga. Geol. Surv. Inf. Circ. 53, p. 132-135.

Lichter, W.F., Warren, A., and Joyner, B.F., 1968, Water resources of Orange County, Florida: Fl. Geol. Surv., Report of Inv. No. 50.

Mansfield, C.W., 1937, Mollusks of the Tampa and Suwannee Limestone of Florida: Fl. Geol. Surv. Bull. 15, 334 p.

Mitchell, G.D., 1981, Hydrogeologic data of the Dougherty Plain and adjacent areas, Southwest Georgia: Ga. Geol. Surv. Inf. Circ. 58, 124 p.

Palmer, A.N., 1975, The origin of maze caves: Nat. Spel. Soc. Bull., V.37, no. 3, p. 57-76.

Pirkel, E.C., 1956, The Hawthorne and Alachua Formations of Alachua County, Florida: Fl. Acad. of Sci. Quart. J., V. 9, no. 4, p. 197-240.

Pohl, E.R., 1955, Vertical shafts in limestone caves: Nat Spel. Soc. Occ. Paper 2, 24 p.

Pollard, L.D., Grantham, R.G., and Blanchard, H.E., 1978, A preliminary appraisal of the impact of agriculture on ground-water availability in Southwest Georgia: U.S.G.S., Water-Resources Inv. 79-7, 22 p.

Popenoe, P., Kohout, F.A., and Manheim, F.T., 1984, Seismic-reflection studies of sinkholes and limestone dissolution features on the Northeastern Florida shelf: in Sinkholes: Their Geology, Engineering, and Environmental Impact, Beck, B.F. (ed.): Rotterdam, NL, A.A. Balkema Publishers, p. 43-57.

Puri, H.S., 1957, Stratigraphy and zonation of the Ocala Group: Fl. Geol. Surv. Bull. 38, 248 p.

Puri, H.S., and Vernon, R.O., 1964, Summary of the geology of Florida and a guidebook to the classic exposures: Fl. Geol. Surv. Spec. Pub. No. 5, 312 p.

Randazzo, A.F., 1976, Petrographic and geohydrologic model of aquifer limestones in Florida: Fl. Water Resources Research Center (U. of Fl.), Pub. No. 5, 51 p.

-----, 1982, Comments on the geology of the fieldtrip area: in Cenozoic vertebrate and invertebrate paleontology of North Florida: S.E. Geol. Soc. Guidebook No. 24, p. 5-13.

Rhoades, R., and Sinacori, M.N., 1941, Pattern of groundwater flow and solution: J. of Geol., V. 49, p. 785-794.

Roseneau, J.C., Faulkner, G.L., Hendry, C.W., Jr., and Hull, R.W., 1977, Springs of Florida: Fl. Bureau of Geol. Bull. 31, 461 p.

Schmidt, W., and Scott, ToM., 1984, Florida karst - its relationship to geologic structure and stratigraphy: in Sinkholes: Their Geology, Engineering and Environmental Impact, Beck, B.F. (ed.): Rotterdam, NL, A.A. Balkema Pub., p. 11-16.

Sever, C.W., 1965, Groundwater resources and geology of Seminole, Decatur, and Grady Counties, Georgia: U.S.G.S. Water Supply Paper 1809-Q, 30 p.

Stringfield, V.T., LeGrand, H.E., and LaMoreaux, P.E., 1977, Development of karst and its effects on the permeability and circulation of water in carbonate rocks, with special reference to the Southeastern States: Geol. Surv. of Al. Bull. 94, pt. G, 68 p.

Sweeting, M.M., 1973, Karst Landforms: N.Y., Columbia U. Press, 362 p.

Thrailkill, J., 1968, Chemical and hydrologic factors in the excavation of limestone caves: Geol. Soc. Am. Bull. V. 79, p. 19-46.

Threadgill, E.D., 1981, Why chemigate: in Young, J.R. (ed.), Proc. of the Nat. Symp. on Chemigation, Rural Development Center, Tifton, Ga., p. 1-6.

Wagner, J.R., 1982, Hydrogeology of the Northwest Florida Water Management District: in Fisher, George (ed.), Proc. of the 1st Annual Symp. on Fl. Hydrogeology, Northwest Fl. Water Mgt. Dist. Bull. 82-2, p. 37-50.

Wait, R.L., 1963, Geology and ground-water resources of Dougherty County, Georgia: U.S.G.S. Water Supply Paper 1539-P, 102 p.

Watson, T.W., 1981, Geohydrology of the Dougherty Plain and adjacent area, Southwest Georgia: Ga. Geol. Surv. Hydro. Atlas 5, 4 sheets.

Wilson, J.D., and Pickering, Sam, 1973, Geologic analysis of Dougherty County area, Georgia: Ga. Geol. Surv. contract report to the Southwest Ga. Planning and Development Comm., 40 p.

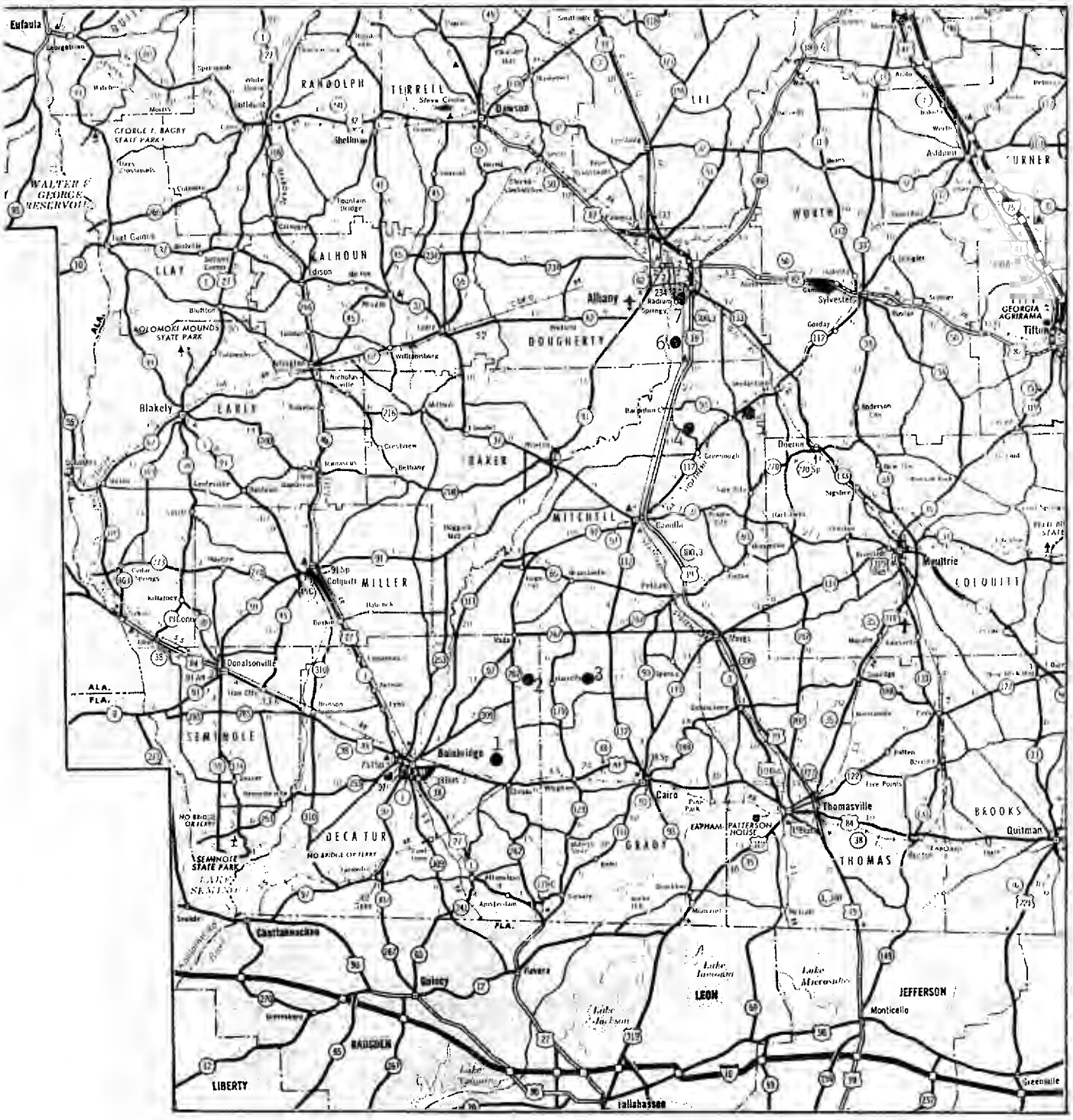


Figure 20: Approximate location of field trip stops (1-7) as listed in the following Road Log. The base map is the state highway map and many of the roads to be taken are not shown. One inch = app. 13 miles and true north is vertical.

ROAD LOG

Mileage Between Points	Total Mileage	
0.0		Begin on U.S. 27 at Georgia-Florida line.
1.1	1.1	Go north to intersection of U.S. 27 and Ga. 262.
14.5	15.6	Bear right and drive north on 262 to intersection with U.S. 84 in Climax.
0.3	15.9	Turn right (east) on Ga. 262 and U.S. 84 and go through Climax to intersection where 262 bears left (north).
0.7	16.6	Go left (north) on Ga. 262 to dirt road bearing left at Mt. Horum Church.
1.2	17.8	Take dirt road north. You are driving along the drainage divide between the Dougherty Plain to the left (west) and the Tifton Upland to the right (east). Go north to intersection with Ga. 312.
1.1	18.9	Turn left (west) on Ga. 312 and drive to crude dirt "road" on right.
		STOP 1
		Walk in dirt road to sinkhole entrance to Climax Cave. See Figure 9 for a stratigraphic section of the entrance, Figure 12 for a map of the cave and the topography, and pages 28-36 for a discussion of Climax Cave. The Pelham Escarpment is just to the northwest of the entrance.
0.5	19.4	Board vehicles and continue straight, descending Pelham Escarpment, to intersection at base of hill.
2.9	22.3	Turn right (north) and drive to the first intersection.
1.5	23.8	Turn right and drive to Ga. 262.
4.9	28.7	Turn left (north) on Ga. 262 and drive across intersection with Ga. 309, continuing straight to an unpaved road marked New Home Church.

Mileage Between Points	Total Mileage	
0.5	29.2	Turn right and drive to noticeable downhill slope. You are now crossing the bottom of a very wide (approximately one mile), shallow, flat-bottomed sinkhole called The Level. It is shown on the map in Figure 15. This must represent an old age stage of development. Obviously, backwearing of the sides has been dominant over downward erosion. Possibly a more cohesive (clayey) layer of the residuum has impeeded deepening. Note on the map that there is a deep ponor, or swallow hole, where the drainage sinks in the middle of The Level. Note also the similarity between this feature and the course of Big Slough to the northwest.
0.9	30.1	Continue driving straight to noticeable uphill slope. This is the far edge of The Level.
0.4	30.5	Continue driving to intersection.
1.0	31.5	Turn left and drive to bridge over small creek. This is one of the reaches of the Big Slough where it appears to be a normal surface drainage feature.
0.3	31.8	Continue straight to gate on left near top of hill.
0.6	32.4	Turn left into gate and follow the crude dirt road across the field. This is private property and should not be entered without the owner's permission.* Park by the side of Big Rock Sink.
		STOP 2
		Big Rock Sink and Little Rock Sink are sinking points for Big Slough (see Figure 15). Big Rock Sink swallows the normal flow of Big Slough but in flood may overflow into Little Rock Sink. After periods of heavy precipitation, both may fill up completely and overflow into the broad uvala downstream. Return to vehicles and retrace route to Ga. 262.
0.6	33.0	Drive back to gate.
1.3	34.3	Turn right and drive to intersection.

*The owner's name is James Boutwell. He lives in the first house on the left if you continue straight on the main dirt road.

Mileage Between Points	Total Mileage	
2.0	36.3	Turn right and drive to Ga. 262.
0.8	37.1	Turn right and drive north on 262 -- to south "bank" of Big Slough.
		Continue straight crossing Big Slough. Note the lack of any defined channel and how this resembles The Level. This reach of the Slough appears to be a uvala composed of several sinks which have coalesced as they widened. See Figure 15.
1.3	38.4	North "bank".
0.9	39.3	Continue straight to intersection with Ga. 97.
0.9	40.3	Bear right and follow Ga. 97 and 262 until Ga. 262 turns right.
3.3	43.6	Turn right on 262 and drive to west bank of Big Slough. This reach of the Big Slough is a broad, bankless, dry valley punctuated by periodic sinkholes. It is southwest of Gee Pond (Figure 14).
0.7	44.3	Continue to east "bank" of Slough.
4.9	49.2	Continue straight to intersection with Ga. 112.
4.6	53.8	Turn right (south) on Ga. 112 and drive up and down the dissected escarpment, past the Fire Lookout Tower to the dirt road on the right.
1.7	55.5	Turn right on dirt road and drive to fork.
0.3	55.8	Bear right and drive to fork at Church.
1.4	57.1	Bear right and continue, irregularly descending the Escarpment, to a crude dirt road on the right.
0.2	57.4	Turn right and follow road to back of field where a very rough road leads right into the woods.
		STOP 3
		Park and walk along the road through the woods to Waterfall Cave. The two sinks you pass on the left lead to Polley's Lost Cave and Shingle Cave, which are both small. For a discussion of Waterfall Cave, see page 28 through 36 and Figure 13.

Mileage Between Points	Total Mileage	
		LUNCH STOP
		Return to vehicles after lunch. Please bring all trash with you.
0.2	57.6	Drive back around field to main dirt road.
1.4	59.0	Turn left and drive to Church.
0.3	59.2	Bear left and drive to T intersection.
1.7	60.9	Turn left and drive back to Ga. 112.
3.1	64.0	Turn left on Ga. 112. Wooded area in field to the right is the sinkhole entrance to Glory Hole.
12.8	76.9	Continue straight into Camilla where Ga. 112 turns right (east).
0.6	77.5	Continue on 112 to U.S. 19.
4.1	81.6	Cross U.S. 19 and continue on Ga. 112 to dirt left into pecan grove.
1.9	83.5	Turn left (west) and drive to Hays Chemigation Research Farm.
		STOP 4
		Numerous small center-pivot systems are used to experimentally apply various chemicals to various crops. See pages 15 through 21 for a discussion of irrigation and chemigation.
1.9	85.4	Return to Ga. 112.
7.6	93.0	Turn left (north) and drive to unused limestone quarry. Park.
		STOP 5
		This is the quarry described under "Stratigraphy", pages 7 through 10. A measured section in the quarry is shown on pages 9 and 10. This same section of algal Suwannee Limestone can be seen inside Waterfall Cave.
		The clays and algae in this section were studied by Bowman and Manker (1982) who tentatively

Mileage Between Points	Total Mileage	
		identified the algae as Archaeolithothamnium, but were unable to definitely establish whether the smectite-rich clay which occurs both as individual layers and as amorphous fillings in the algal limestone was bentonite (volcanic) or an alteration product of detrital clays (kaolinite and gibbsite) in a hypersaline tropical marine environment. Mollusc, coral, and foraminifer fossils are also well preserved here. Return to vehicles.
0.5	93.5	Turn right and go back south on Ga. 112 to paved county road.
5.4	98.9	Turn right (west) on county road and drive to U.S. 19.
0.75	99.6	Cross U.S. 19 and continue straight to Ga. 3.
5.9	105.5	Turn right (north) on Ga. 3. Drive to old scales on left, just past power plant. Park. STOP 6 Walk to Blue Hole Spring near Dry Creek. As you walk downstream along the bed of Dry Creek, notice that the creek bed actually goes uphill next to the Blue Hole Spring. In moderate flows, this diverts Dry Creek into the Blue Hole where it sinks underground to flow through a spongework to the Flint River (approximately 0.5 miles west). In high flow, the Flint backfloods via these same conduits and the Blue Hole Spring is filled with highly turbid Flint River water. Since this feature may be either influent or effluent, it should be called an estavelle. Return to cars.
5.4	110.9	Continue north on Ga. 3 to the Radium Springs Inn, now destroyed by fire. Park. STOP 7 Walk to spring rise pool. Note the spongework megaporosity in the limestone around the pool and the linear openings due to additional solution along joints. For a discussion of Radium Springs, see page 46 and Figures 17 and 18.

Mileage Between Points	Total Mileage	
		Return to cars. Return to Tallahassee. Go east from the Radium Springs parking lot on Ga. 257 (a right turn from the way you arrived) to U.S. 19. Turn right on 19 and drive south to Thomasville. Turn right on U.S. 319 and drive through Thomasville to Tallahassee.