
6
LEGEND
COMPLETED PROJECT
NOT COMPLETED
EASTERLY LIMIT OF PROJECT
0.0 Mi.
HENRY H. BUCKMAN LOCK
10 Mi.
20 Mi.
30 Mi.
RODMAN DAM
LAKE OCKLAWAHA
EUREKA LOCK & DAM
40 Mi.
50 Mi.
R.N. BERT DOSH LOCK
60 Mi.
70 Mi.
80 Mi.
90 Mi.
100 Mi.
107 Mi.
DUNNELLON LOCK
INGLIS LOCK BYPASS CHANNEL
INGLIS BYPASS SPILLWAY
INGLIS LOCK
INGLIS DAM & SPILLWAY
WESTERLY LIMIT OF PROJECT
ATLANTIC OCEAN
GULF OF MEXICO
INTRACOASTAL WATERWAY, JACKSONVILLE TO MIAMI, FLA.
NASSAU
DUVAL
CLAY
ST. JOHNS
PUTNAM
FLAGLER
ALACHUA
GILCHRIST
LEVY
MARION
VOLUSIA
CITRUS
SUMTER
HERNANDO
ORANGE
SEMINOLE
JACKSONVILLE
STARKE
GREEN COVE SPRINGS
ST. AUGUSTINE
PALATKA
BUNNELL
GAINESVILLE
TRENTON
BRONSON
OCALA
DUNNELLON
CRYSTAL RIVER
INVERNESS
LEESBURG
BUSHNELL
BROOKSVILLE
DADE CITY
DELAND
SANFORD
ST. JOHNS RIVER
OKLAWAHA RIVER
WITHLACOOCHEE RIVER
SUWANNEE RIVER
HOMOSASSA RIVER
LAKE ROUSSEAU
Lake George
Lake Harris
Lake Apopka
Lake Monroe
Lake Lochloosa
Orange Lake
CROSS-FLORIDA BARGE CANAL
SCALE IN MILES
10 0 10 20
DEPARTMENT OF THE ARMY
JACKSONVILLE DISTRICT, CORPS OF ENGINEERS
JACKSONVILLE, FLORIDA
7-25-97

GEOHYDROLOGY OF THE CROSS-FLORIDA BARGE CANAL AREA WITH SPECIAL REFERENCE TO THE OCALA VICINITY

By Glen L. Faulkner

U.S. GEOLOGICAL SURVEY

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P.20-Figure 6.-After U. S. Weather Bureau, 1962.

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P.30-Figure 11.-Partial sources: Applin, 1951; Applin and Applin. 1965; McNeil, 1949; Puri, 1957; Puri and Vernon,1964; Stringfield, 1966; and Vernon, 1951.

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PREFACE

This report is concerned in general with the possible effects of the Cross-Florida Barge Canal on the pre-canal hydrologic regime of the area, and in particular with the effect of the canal on the natural levels, movement, and quality of the ground-water. A discussion of the effects of water-table changes and surface-water impoundments on the ecology of the area is beyond the scope of the report, as is a comprehensive discussion of possible changes in surface-water quality resulting from the several impoundments along the canal route. However, the importance of maintenance of high surface-water quality standards is emphasized in order that the risks of movement of contaminated surface water from the canal into the aquifer may be minimized.

GEOHYDROLOGY OF THE CROSS-FLORIDA BARGE CANAL AREA WITH SPECIAL REFERENCE TO THE OCALA VICINITY

By Glen L. Faulkner

CONCLUSIONS

The geohydrologic investigation of the Cross-Florida Barge Canal area reveals that the design of the canal and the plan of operation are consonant with the hydrologic regime. Thus, canal operations likely will not seriously affect the regimen of the economically and ecologically important large springs--the water level, rate of flow, and the quality of water at Rainbow Springs, for example. Further, if Summit Pool lockage losses are essentially replaced and operating precautions are taken against pollution of Summit Pool waters, there should be no noticeable adverse effects on the water level, rate of flow, and quality of water of Silver Springs.

In the canal reaches below Eureka Lock and Dam, and below Dunnellon Lock, little change in water levels in the Floridan aquifer is expected, and pollution hazards other than possible lockage of water from the Gulf of Mexico are not considered significant. Also, little exchange of water between the canal and the aquifer is expected in these areas. However, some water likely will leak around Dunnellon Lock and Inglis Lock and Dam, some ground water will flow into the canal for a mile or two below Dunnellon Lock because of a head differential between the canal's operating stage and natural potentiometric surface of the Floridan Aquifer. Also, farther westward there is a possibility for some direct exchange of water between the Floridan Aquifer and Inglis Pool in a few places where the dredged canal channel through the pool is expected to penetrate rocks of the aquifer. The potentiometric surface in most of these places is probably close to the planned operating stage of the Pool, and it is likely any net exchange will be small.

Minimal locking up of Gulf water into Inglis Pool is indicated by past experience at the old Inglis Lock and Dam on the Withlacoochee River, and by the fact that considerable flushing action should result from Withlacoochee River flows augmented by possible increases in fresh water flow to the Gulf from some additional ground-water inflow to the Inglis Pool reach and from lockage releases at Dunnellon Lock. However, if significant lockage of Gulf water is found to occur, effecting special operational procedures will minimize such lockage into the canal pools.

The Eureka Pool level near the Eureka Lock and Dam site will be about 15 feet above the normal altitude of the potentiometric surface. The hydraulic head difference between the downstream part of Eureka Pool and the natural potentiometric surface is thus expected to raise the potentiometric surface of both the Floridan and shallow aquifers about 15 feet in the intermediate vicinity of the Eureka Lock and Dam site. An appreciable amount of water, therefore, is likely to seep from the northern or downstream part of the Eureka Pool into both the shallow aquifer and Floridan Aquifer. The steep gradient of the natural potentiometric surface to the river from west of the lock and dam area seems to preclude an appreciable change in the potentiometric surface very far to the west. However, it is possible that downward seepage from wet lands in the Eureka-Ft. McCoy area could be appreciably reduced by raised ground-water levels. Head loss through the sand and clayey sand east of the lock and dam area should be sufficient to result in only negligible change in the potentiometric surface east of Mount Dora Ridge. Although the seepage from downstream part of the Eureka Pool may be appreciable, the filtering action of the sand and clayey sand should be sufficient to at least restrict particulate contamination of water in the aquifer, although the natural filter will not preclude movement of possible dissolved contaminants from the pool to the aquifer. To the southwest, in the upstream part of Eureka Pool little change is expected as the natural potentiometric surface is higher than the planned stages for the pool.

A significant quantity of water is expected to flow from the canal to the Floridan Aquifer in a four-mile reach of the Summit Pool south of Silver Springs. The inflow-outflow relation between the Floridan Aquifer and the Summit Pool will result in some minor readjustment of the potentiometric surface near the canal. The change in the potentiometric surface is expected to be most significant in the immediate vicinity of the Summit Pool with the effects generally becoming insignificant within 2 or 3 miles of the canal. Ground-water levels should decline less than two feet from natural levels in those reaches of the Summit Pool where inflow to the canal will occur, except for that reach just east of Dunnellon Lock where the canal cuts through a local potentiometric high, and where the maximum decline in water level will be about 15 feet. However, here the decline of water levels should diminish with distance from the canal and a potentiometric high should remain on each side of the canal, although they will be lower than the natural high and some wet lands are expected to be drained as a result.

Levels should rise about 1 foot in the aquifer at zones of outflow from the canal. If the variation in the stage of the Summit Pool were to be independent of any control but the natural fluctuation of the potentiometric surface the Floridan aquifer, the predicted range in pool stage, according to the 36-year ground-water level record, would be about 12.3 feet. However, under controlled conditions the expected possible range in fluctuation of canal stage in the Summit Pool reach is only about 10.5 feet, with a maximum stage of 51.5 feet and a minimum stage of 41.0 feet with replacement of lockage releases back-pumpage from lower pools. The actual stage will depend on natural levels, the controlled stage of Eureka Pool, pumping schedules, and lockage schedules.

Some ground water that flows to Silver Springs from the south will be diverted through the Summit Pool from one zone of preferential flow in the aquifer to another. However, little water will be lost from Silver Springs drainage area nor will the flow of Silver Springs be reduced if lockage losses from the Summit Pool are replaced with pumpage from the lower pools.

The most likely place where the completed canal can affect the quality of ground-water is in the Summit Pool area, especially at one or two zones of outflow from the canal to the aquifer. Water in the Summit Pool will move directly from the principal zone of outflow into the Floridan Aquifer and in time will be discharged at Silver Springs. It is possible that some water will also move from the Summit Pool through a less important outflow zone toward the Withlacoochee River. Consequently, if the Summit Pool water becomes contaminated, water in certain parts of the aquifer could be contaminated and subsequently discharged at Silver Springs if measures are not taken to reduce Summit Pool levels at such times.

A workable method for direct measurement of time of travel of water and, therefore, of possible contaminants, especially dissolved substances, through the aquifer to Silver Springs is not available. However, from knowledge of the general hydraulic gradient and transmissivities, and from estimates of the porosity and thickness of the zone of flow, a approximate average velocity of about 200 feet per day and a time of travel of about 140 days for a distance of about 5 miles from the canal to Silver Springs has been estimated. Although these values are considered reasonable for the zone flow from the canal to Silver Springs, the values should be used with care, because actual porosity and thickness of the principal zone of flow may vary widely. If flow occurs preferentially through a cavernous zone or zones, then velocities in the aquifer would be considerably greater than given above.

Application of proper construction and operational procedures designed to avoid contamination of water in the Summit Pool will tend to minimize the possibility of polluting water in the aquifer and at Silver Springs. If Summit Pool water accidentally becomes contaminated, operational procedures can be instituted for a limited time to protect the aquifer and Silver Springs.

The studies documented in this report are considered of sufficient scope and intensity to adequately determine possible effects if the canal on adjacent ground water and, conversely, of the ground-water regimen of the canal. However, the additional studies discussed would further define certain geohydrologic relations, a knowledge of which might prove useful in refining operational procedures under various hydrologic conditions. As excavation of the canal progresses, additional geologic and hydrologic information revealed by the construction process would enhance the existing geohydrologic knowledge. Implementing certain special studies of the Summit Pool area likely will better define routes and rates of ground-water to and from the canal, and better predict the effects of certain operational procedures on the hydrologic system. Special studies could include methods for geohydrologic mapping, use of pumping tests and use of analog models. Testing the actual effects of canal operations on the hydrologic system before the canal is placed in full operation will provide information that will allow the planned operating schedules to be adjusted if necessary. This procedure will further assure compatibility of planned operating schedules with existing hydrologic conditions, and will more accurately delineate zones of outflow to the aquifer.

Such additional studies coupled with continuation of the hydrologic monitoring program begun in mid-1966 will help to further assure that the water resources of the Cross-Florida Barge Canal area are protected.

ABSTRACT

The Cross-Florida Barge Canal route commences at Palatka on the St. Johns River, about 75 miles upstream from the Atlantic Ocean, and extends 110 miles southwestward across Peninsular Florida into deep water in the Gulf of Mexico near Yankeetown. The canal will be equipped with five locks, each 600 feet long and 84 feet wide, and the channel will be a minimum of 12 feet deep and 150 feet wide. From near Ocala northeastward, the canal channel will replace much of the natural channel of the Oklawaha River, and will be excavated into beds of the so-called shallow sand aquifer of Miocene age and younger, which overlies limestone of the Floridan Aquifer. Westward from Ocala most of the canal will be excavated below the potentiometric surface into limestone and dolomite of the Floridan Aquifer. Water levels of Rodman, Eureka, and Inglis Pools will be controlled by dams and spillways with the limited exchange of water between the pools and the aquifers. The water levels in the Summit Pools will fluctuate with the natural changes in the ground-water level of the Floridan Aquifer, although the stage of the pool will be controlled partly by the stage held in the Eureka Pool. A dynamic inflow-outflow relationship will exist between the Summit Pool and the Floridan Aquifer.

The Floridan Aquifer in the canal area is 1,000 to 1,200 feet thick and consists of limestone and dolomite of middle Eocene Miocene age, including from older to younger, the Lake City, Avon Park and Ocala limestones plus permeable sandy, dolomitic limestone in the lower part of the Hawthorn Formation. It is possible that most of the flow to the two major springs in the area occurs in the upper 100 feet or so of the aquifer in the Ocala Limestone. The aquifer is underlain by the Oldsmar limestone of early Eocene age and is overlain by sand, clayey sand, clay and shell beds of Miocene through Holocene age, in thickness from a few feet to 300 feet. The permeable beds overlying the Floridan Aquifer constitute the shallow aquifer, while the poorly permeable ones act as confining beds where the Floridan Aquifer is under artesian conditions.

A north-south line drawn separating the head of Silver Springs on the west from the Oklawaha River on the east marks the approximate western limit of a continuous blanket of materials of Miocene-Pliocene(?) age covering the rocks of the Floridan Aquifer. East of the line much of the aquifer is under artesian conditions, particularly in the Oklawaha River valley, although in some areas east of the valley direct recharge through thick permeable Miocene-Pliocene(?) sands occurs. West of the line only scattered remnants of a once continuous Miocene-Pliocene(?) cover remain. Lack of the cover is a result of erosion on the crest and flank of the Ocala Uplift, a broad northwest-southeast trending anticlinal upwarp, the axis of which is crossed by the canal route in the Dunnellon area. Over most of this area the Floridan Aquifer is unconfined, and receives direct recharge through a cover of a few tens of feet of sand and clayey sand of Quaternary age.

Tensional stresses during the structural evolution of the Ocala Uplift produced an intersecting system of fractures and normal faults in rocks of the Floridan Aquifer. The fractures and faults are important controls for orientation of solution channels, and, therefore, for development of ground-water circulation patterns.

When the system surface streams which once drained the Barge Canal area eroded the poorly permeable Miocene-Pliocene(?) cover from the flanks of the Ocala Uplift, surface runoff was reduced and precipitation began to directly infiltrate the underlying limestones. Now only principal streams, such as the Oklawaha and Withlacoochee Rivers and a few short tributaries, remain, while one of the most highly developed subsurface drainage systems in the world has evolved in cavernous limestone of the Floridan Aquifer. Two of the larger fresh-water springs in the world now discharge from the Floridan Aquifer in the canal area. Silver Springs near Ocala discharges an average 531 mgd (million gallons per day) down the 4-mi long Silver River which flows on poorly permeable beds to the Oklawaha River. Rainbow Springs near Dunnellon discharges on average 468 mgd from numerous orifices in the bed of the 5-mile-long Rainbow River, which flows into the Withlacoochee River. The heads of the springs have migrated to their present positions partly because of a tendency of ground-water levels to decline as permeability in the aquifer is increased due to removal of limestone by solution, and because of mechanical erosion of the limestone in the vicinity of the spring heads. Also, points of principal spring discharges have shifted in the past due to changes in ground-water levels in response to changes in sea level. The subsurface drainage system is continuing to evolve today, as evidenced in part by frequent occurrence of new sinkholes and by the presence of significant amounts of calcium bicarbonate in the spring waters.

Rodman Pool, at the east end of the canal, is separated from the Floridan Aquifer by poorly permeable materials.The pool's operating water level will be only a few feet above the potentiometric surface at the downstream end, and at or slightly below the potentiometric surface at the upstream end. Little exchange of water between the Rodman Pool and the Floridan Aquifer is expected. Eureka Pool, just upstream from Rodman Pool, will also be separated from the Floridan Aquifer by poorly permeable beds. However, the stage of the pool will be about 15 feet higher than the natural potentiometric surface at the pool's downstream end, and some seepage into the Floridan Aquifer is anticipated through faults and leaky parts of the poorly permeable beds, with a consequent rise in ground-water levels in areas adjacent to the lower end of the pool. Possibilities for particulate contamination of the aquifer will tend to be minimized because of the filtering capacity of the materials through which water must pass to reach the aquifer, although the natural filter will not preclude movement into the aquifer of contaminants which might become dissolved in the pool waters. No significant interchange of water between the pool and the aquifer is expected at the upstream end of the Eureka Pool. Present construction plans indicate an operating stage for Eureka Pool which will range between 38 and 40 feet above mean sea level, although it is possible to dredge the pool deep enough to permit a range in stage of 36 to 40 feet. A backwater effect extending up Silver River from the Eureka Pool is expected to regulate the stage at the head of Silver Springs between 39 and 44 feet above mean sea level if the pool ranges between 36 and 40 feet. If Eureka Pool ranges only between 38 and 40 feet, the range of stage at the head of the springs should be about 41 to 44 feet above msl. From Inglis Lock west, the canal will have direct connection with the Gulf, and canal stage will fluctuate with the Gulf tide. Since the canal stage will be slightly lower than the adjacent ground-water levels along much of the reach, there will be some ground-water inflow to the canal.

No significant changes in the existing ground-water regime are expected in the vicinity of Inglis Pool, the first step up in the canal east of the Gulf. Existing ground-water and surface-water levels in the area will not change appreciably, and the natural stage and flow of Rainbow Springs, which will flow by way of Rainbow River into Inglis Pool, should not be affected by canal operations. A possible adverse effect of the canal on the Inglis Pool area could result if sea water is locked-up from the Gulf through Inglis Lock. However, the high step of 25 feet at the lock, flushing action of continuous flow from Inglis Pool to the lower reaches of the Withlacoochee River, and use of possible preventive locking procedures should minimize the problem.

The potential for adverse effects on the ground-water regime is greatest in the area of the Summit Pool. Through most of the length of the pool, the canal channel will be excavated into limestone of the Floridan Aquifer to depths of 12 to 27 feet below the potentiometric surface. Changes that will take place in the ground-water flow system in the Silver Springs drainage area, once the canal is completed, were estimated by flow-net analysis. Variation in aquifer transmissivity was determined by calculating transmissivity in 25 different flow cells surrounding Silver Springs. Transmissivity in the 25 cells averages about 15,600,000 gpd/ft [2,090,000 ft^2/day (feet squared)] but transmissivity in the six cells through which the Summit Pool passes ranges from 9,000,000 to 44,000,000 gpd/ft (1,210,000 to 5,900,000 ft^2/day). Transmissivity was used to compute static stage of the Summit Pool under given ground-water level conditions. Had the canal existed in May 1968 and had the stage of Eureka Pool been held at 36 feet at the time, the static stage in Summit Pool would have been about 42.1 feet above mean sea level. Thus a conceptual model of the changes in the potentiometric surface wrought by the finished canal was drawn, and zones of ground-water inflow and outflow were delineated. Most outflow from the Summit Pool to the aquifer should be limited to one 4-mile-long zone along the north side of the pool, about 5 miles south of Silver Springs. It is estimated that a water volume equivalent to about 8 percent of the daily flow of Silver Springs will enter the Summit Pool each day from the southern one-third of the Silver Springs drainage area. A like amount will reenter the aquifer at the main zone of outflow and move toward Silver Springs at an estimated average velocity of about 200 feet per day, if something close to the natural static stage of the pool is maintained by return pumpage of the lockage losses. At a velocity of 200 feet per day, water from the Summit Pool would discharge at Silver Springs about 140 days later. However, any estimate of velocity in the highly cavernous limestone aquifer in the area should be used with caution, because difficult-to-measure changes in porosity and thickness of major zones of flow may cause large variations in velocity.

If all lockage losses are returned the Summit Pool by pumping from Eureka Pool, no net loss from the Silver Springs drainage area, except for some evaporation from the water surface in the canal and possible leakage around locks, will result from canal operations. The zone of outflow from the Summit Pool to the aquifer will be in a natural potentiometric trough, and the zone of inflow will be in a potentiometric ridge area. The equilibrium water level in the pool will tend to be about 1 foot higher than the altitude of the lowest level in the pre-canal potentiometric through, and about 2 feet lower than the highest level on the pre-canal potentiometric ridge. West of the Silver Springs drainage area just east of Dunnellon Lock, in the area of a local potentiometric high, the water level in the Summit Pool is expected to be about 15 feet below the natural potentiometric surface. In most areas 2 to 3 miles away from the Summit Pool, effects of the canal on the natural potentiometric surface should be slight. The stage of the Summit Pool, judging from the 36-year record for ground-water level changes and the anticipated indirect effect of the controlled stage in Eureka Pool, should have a maximum range of about 10.5 feet with a maximum water level of about 51.5 feet above msl and a minimum of about 41.0 feet above msl. Of particular importance in the Summit Pool is a implementation of well planned construction and operational procedures designed to minimize risks of ground-water contamination.

INTRODUCTION

In 1964 the U. S. Army Corps of Engineers started construction of the Cross-Florida Barge Canal, with completion now (1970) scheduled for 1974. The canal is the ultimate result of plans considered for over a hundred years for a short-cut waterway connection across North Florida or South Georgia, between the Atlantic and Gulf Coasts. Numerous routes were considered before one in north-central Florida was finally selected (fig.1).

The chosen route, officially designated as Route 13B, extends southwestward from Jacksonville a distance of 185 miles to deep water in the Gulf of Mexico near Yankeetown. One reason this route was considered most practical was that it could utilize parts of the St. Johns and Oklawaha Rivers on the east and the Withlacoochee River on the west.

Construction of a sea-level ship canal was actually begun along the present canal route in the mid-1930's. However, largely because of great concern over the probable ill effects of the sea-level canal on Florida's ground-water resources, work was discontinued after only 2 years.

Early in World War II, interest in a canal was again aroused partly because of the protection such an inland waterway would offer to wartime shipping, which otherwise would be exposed to attack by enemy submarines on the much longer trip in open seas around the southern tip of Florida. This time, however, it was decided to overcome the threat to the State's ground-water resources by constructing a comparatively shallow barge canal in which water levels would be regulated by a series of five locks to approximate natural ground-water levels. Figure 2 is a profile along the canal alignment from the St. Johns River to the Gulf. It shows the locations of the five locks, the altitudes of the bottoms of the several pool reaches, and the ranges of water levels which the locks will maintain.

Purpose and Scope

The design for the canal calls for an accommodation to the local ground-water regime to the extent that the canal will alter natural conditions as

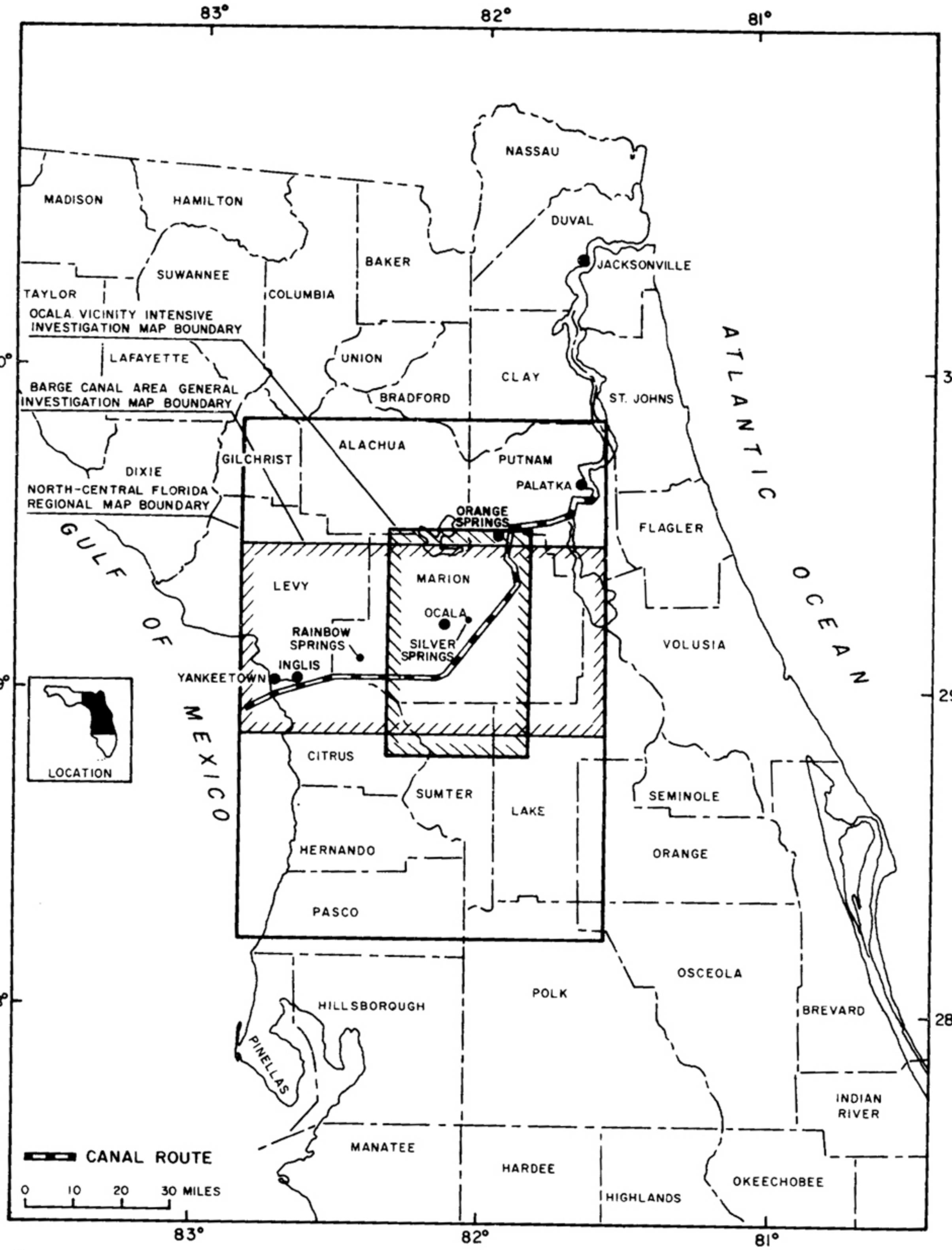


Figure 1. Index map of North Florida showing route of Cross-Florida Barge Canal, outlines of geohydrologic investigation areas, and monitoring network area.

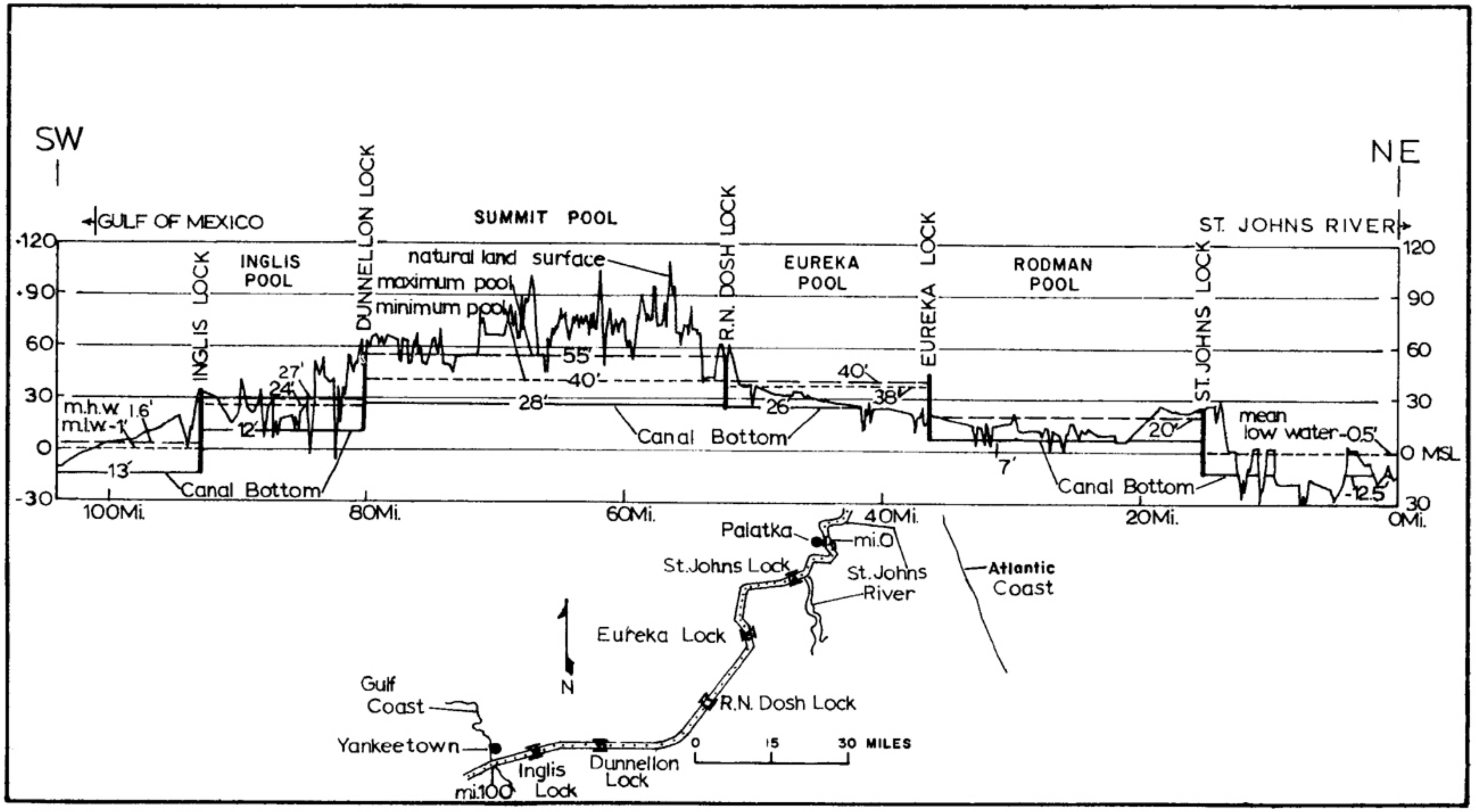


Figure 2. Profile along Cross-Florida Barge Canal from St. Johns River at Palatka to the Gulf of Mexico.

little as possible and that it will have no significant adverse effect on the ground-water system. In a further effort to see that this aim be accomplished, plans were made in 1965 for the U. S. Geological Survey to monitor the ground-water in the area of the Barge Canal before, during, and after canal construction (fig. 1). The monitoring program was to serve two basic purposes. First, it would help to point out or signal undesirable effects, if any, that the canal might have on the ground-water regime, and thereby permit corrective measures before any lasting harm might result. Second, data collected during the program could be essential to proper litigation of possible future damage claims against the builder and operator of the canal.

In addition to the general hydrologic investigation incidental to the collection and interpretation of water-level and water-quality data in that part of the monitoring network area concentrated near the canal alinement, it was decided to conduct an intensive geohydrologic investigation in the Ocala vicinity of the canal area (fig. 1). Although the basic geologic and hydrologic facts of the locality were generally understood, several factors indicated the desirability of such a special investigation before canal excavation and lock construction were started in this critical reach of the canal.

Much of that part of the Barge Canal that will traverse the Ocala area will be excavated into the limestone of the Floridan Aquifer to a depth of 12 to 27 feet below the water table. People in the Ocala area depend on water from the Floridan Aquifer for practically all of their water supplies. Also, within the special study area is Silver Springs, one of the largest freshwater springs in the world, and a multi-million-dollar tourist and recreational facility. As some water is expected to seep from the Summit Pool of the canal toward Ocala and Silver Springs, concern has been expressed over the possibility of adverse effects of canal construction and operation on Ocala's ground-water supplies, and on the flow and quality of water issuing from Silver Springs.

Scope of the intensive investigation includes determination of: principal recharge and discharge areas; direction, rates, and principal routes of ground-water flow; amounts of recharge and discharge; quality of water; and seasonal and long-term ranges in water-level and water-quality fluctuations. The basic purpose is to predict the effect of the canal on the ground-water regime, and thereby to enable recognition of any special hydrologic problem in time to permit necessary changes in design or planned operational procedures.

Acknowledgments

The author expresses his appreciation to the many individuals who aided him in various ways to complete the investigation. Residents and city and county officials were very cooperative in permitting use of privately and publicly owned wells for observation of water-level changes and the collection of water samples. Several water-well drillers in the area were particularly helpful in providing driller's logs and other information on wells essential to the preparation of the various sections contained in the report. The Bureau of Geology of the Florida

Department of Natural Resources, and the Corps of Engineers, Jacksonville, Florida District, generously supplied basic geological and engineering data needed for the study. Many U. S. Geological Survey personnel throughout the Florida District and elsewhere, but especially in the Ocala Subdistrict, were of immeasurable assistance, not only in the collection and reviewing of basic data, but also by providing helpful technical advice and comments pertaining to the many hydrologic and geologic problems examined in the process of the investigation.

The writer acknowledges the several very rewarding discussions with Robert 0. Vernon and Harbans S. Puri of the Florida Bureau of Geology on the geology of central Florida. Special thanks is due Lewis 0. Stafford of the Corps of Engineers for his aid in the exchange of various types of data needed for the investigation.

The help provided the author by several members of the Survey was particularly outstanding. Walter S. Wetterhall first introduced the author to the general hydrology and geology of central Florida. Both Mr. Wetterhall and Frank N. Visher generously discussed results of work they had done earlier on the investigation of the hydrology of Marion County, Florida. Charles A. Appel was most helpful in planning a method of flow-net analysis for determination of quantitative aquifer characteristics. Discussions with Gordon L. Stewart about natural concentrations of radioactive isotopes in water provided valuable information pertaining to recharge characteristics in the Barge Canal area. Technical comments and advice from Hilton H. Cooper, Jr., Robert R. Bennett, and John D. Bredehoeft were especially helpful to the writer. Many opportunities were taken during the study for fruitful discussion of geologic and hydrologic problems with Darwin D. Knochenmus. Warren Anderson was of invaluable assistance as an advisor on questions pertaining to surface-water problems. The author sought and received much valuable information and advice on matters of water quality from Boyd F. Joyner, Donald L. Goolsby, Michael E. Beard and Robert T. Kirkland, Jr.

Grateful acknowledgment is extended to Charles P. Laughlin, who had a major responsibility for collecting and processing basic data for the study, and to Vance W. Fabella who assisted Mr. Laughlin in the data collection work. Their aid was an indispensable contribution to the investigation.

Previous Investigations

Several geologic and hydrologic studies have been made that include parts or all of the area of the Cross-Florida Barge Canal. Although no reports have been published that specifically pertain to the geology and hydrology of the Barge Canal, the Corps of Engineers has made studies and prepared reports unpublished on many aspects of the Barge Canal project, including hydrology and geology.

Cooke and Mossom (1929) and Cooke (1945) included in their reports on the geology of Florida, discussions of the stratigraphy in much of the area through which the canal route passes. Vernon (1951) described the geologic structure and stratigraphy of a large part of central Florida, including much of the Barge Canal area. Later work by Espenshade and Spencer (1963), Puri (1964), Vernon and Puri (1964 and 1965), and Chen (1965) provided further details on the geology, stratigraphy in particular, of the general area of investigation. Cooke (1939), MacNeil (1950), and White (1958) described the physiography and geomorphology of central Florida.

Stringfield (1936) reported on artesian water in Peninsula Florida and prepared a map of the potentiometric surface of the Floridan Aquifer, which included the area of the Barge Canal. Unpublished reports by the U.S. Army Engineers (1938) and the U. S. Engineers Office (1943) also contained potentiometric maps of the Florida aquifer in the canal area. In addition, these reports discussed geology and other hydrologic aspects of the canal route. The area of the canal was again included in potentiometric surface map of Florida prepared by Healy (1962). County-wide water resources investigations have been made in several of the counties bounding the canal area, including studies by: Wyrick (1960), Volusia County, Bermes, Leve, and Tarver (1963), Flagler, Putnam and St. Johns Counties; and Clark, Musgrove, Menke, and Cagley (1964), Alachua, Bradford, Clay and Union Counties. In 1966, Stringfield comprehensively re-examined the geohydrology knowledge acquired through the years concerning artesian water in Tertiary limestone in the southeastern United States. Much of the work has application to the Geohydrology of the canal area.

The above listings are not exhaustive, as various other reports and articles, both early and recent, touch on the geology and (or) hydrology of that part of Florida through which the canal route passes. Many such reports are noted in the list of references at the end of this report.

The Hydrologic Monitoring Network and Data Collection Program

The Cross-Florida Barge Canal hydrologic monitoring program will be described and documented in detail in a companion report. However, because of the interdependence of the monitoring program and the geohydrologic investigation the monitoring program is described briefly for purposes of convenience and clarity. Figure 3 shows the locations of observation stations, and Table 1 lists the number and types of the several categories of stations contained in the monitoring network.

In 1965 a network of 73 ground-water and 10 surface-water observation stations was planned to monitor variations in water levels and water quality in that part of the Barge Canal area extending from the vicinity of Orange Springs on the northeast to the vicinity of town of Inglis on the south-west. As there is little likelihood of undesirable effects from the canal on the Floridan aquifer east of Orange Springs, the network does not extend into that part of the canal area. A few widely dispersed observation stations on the fringes of the canal area provided limited regional control. Such stations are located as far north as northern Putnam County and western Alachua County and as far south as central Pasco County.

Water-level and water-quality data collected from a basic network now 1970 consisting of 90 ground-water stations and 13 surface-water stations

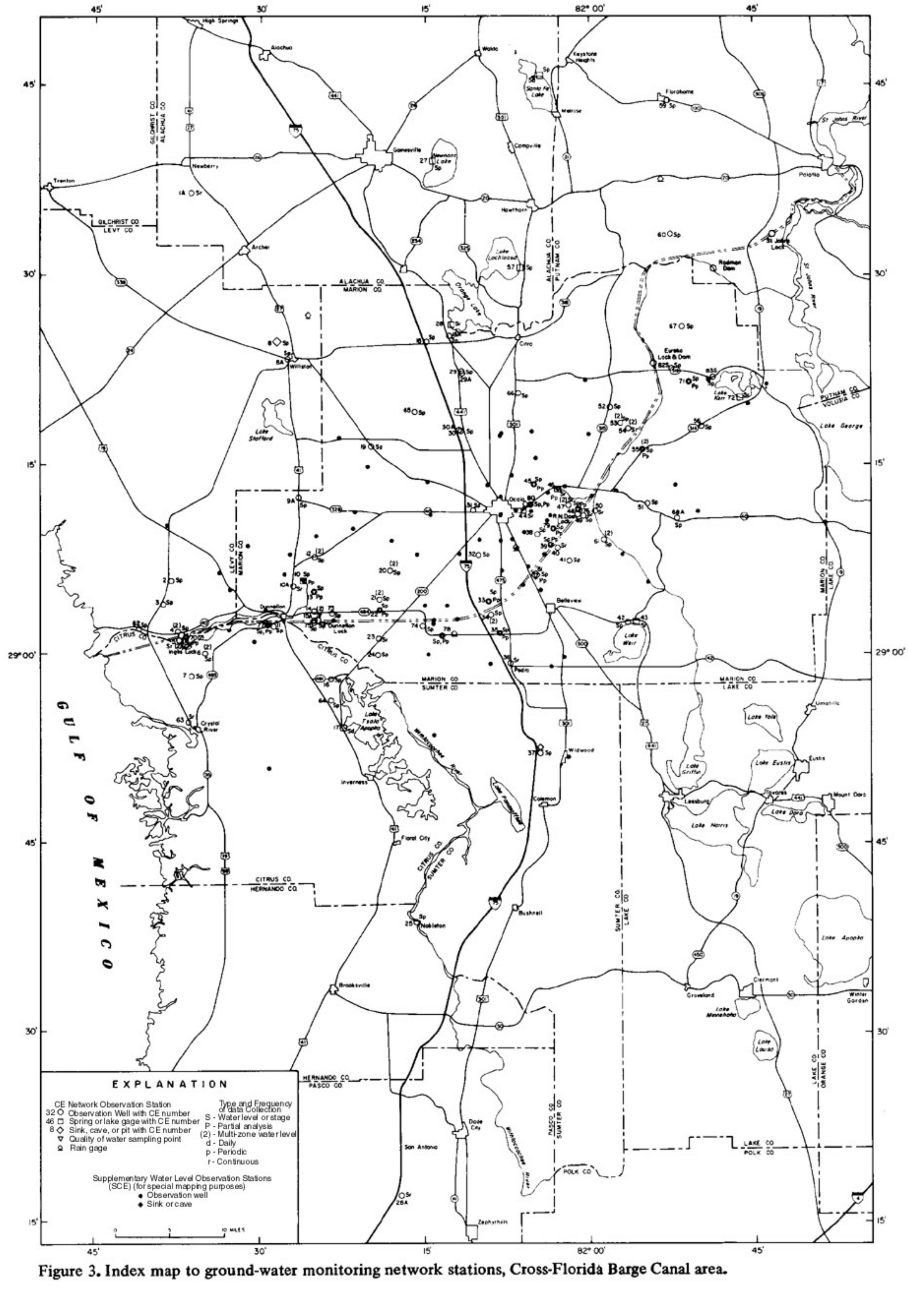


Figure 3. Index map to ground-water monitoring network stations, Cross-Florida Barge Canal area.

Table 1.--Types of observation stations and frequency of data collection in hydrologic monitoring network.

Observation Station Types	Total Number	Frequency of Data Collection		
		Continuous	Bimonthly	Semi-annually
Ground Water				
Floridan Aquifer	143			
Water Level	143	16	57	70
Quality of Water	21	-	15	6
Shallow Aquifer	17			
Water Level	17	-	17	-
Quality of Water	0	-	-	-
Surface Water				
Springs	2			
Stage	2	1	1	-
Quality of Water	2	-	2	
Lakes	7			
Stage	7	3	*	-
Quality of Water	-	-	-	-
Streams	12			
Stage	10	9	-	1
Quality of Water	11	-	1	10
Barge Canal	4			
Stage	4	-	2	2
Quality of Water	4	1	1	2

* Three weekly and one daily lake stage station.

have been reported on a bimonthly basis to the Corps of Engineers since July 1966. Data from a supplementary network of about 70 ground-water and 12 surface-water stations are collected on a semi-annual basis, once in May (low-water period) and once in September (high-water period).

DESCRIPTION OF AREA

Location and Extent

The Cross-Florida Barge Canal route stretches across Florida a distance of 185 miles from the Atlantic Ocean near Jacksonville to the Gulf of Mexico near Yankeetown (fig. 1). From Jacksonville to Palatka, a distance of about 78 miles, the waterway utilizes the St. Johns River. The excavated part of the canal extends from Palatka southwestward about 107 miles to deep water in the Gulf.

The ground-water monitoring program and general geohydrologic investigation is most concerned with that 70-mile reach of the canal extending from the vicinity of Orange Springs in the northeast to the vicinity of the town of Inglis in the southwest. The area of interest measures about 2,500 square miles, and includes most of Marion County and parts of Putnam, Alachua, Lake, Sumter, Citrus, and Levy Counties.

The Ocala vicinity, as delineated for the intensive geohydrologic investigation (fig. 1), comprises about 1,000 square miles mostly in Marion County and within the bounds of the eastern part of the ground-water monitoring network. The city of Ocala and Silver Springs are situated near the center of the area, and the canal route traverses the area in a northeast-southwest direction.

Topography

As is typical of Florida, the Cross-Florida Barge Canal area is one of low relief. Although much of the area is fairly flat, the canal route does pass through some low rolling hill country. Over 60 percent of the excavated part of the canal is located within Cooke’s (1939, p. 14) Central Highlands physiographic division (fig. 4). Northeastward from the vicinity of St. Johns Lock to Palatka, and from just west of the town of Dunnellon to the Gulf Coast, the route traverses the Coastal Lowlands physiographic division.

Land-surface altitudes along the canal route range from near sea level at each end to about 110 feet above sea level near Ocala. In the area of the general geohydrologic investigation, altitudes range from only a few feet to approximately 215 feet above sea level. The lowest altitudes are at the northeast and southwest extremities of the area, and the highest altitude is at a point about 20 miles northwest of Ocala in the Fairfield Hills physiographic subdivision of Puri and Vernon (1964, p. 15).

A more detailed physiographic subdivision of the Barge Canal area appears in Figure 5. Shown are land forms recognized by Puri and Vernon (1964, pp. 13-15, fig. 6) as subdivisions of Cooke’s major physiographic divisions.

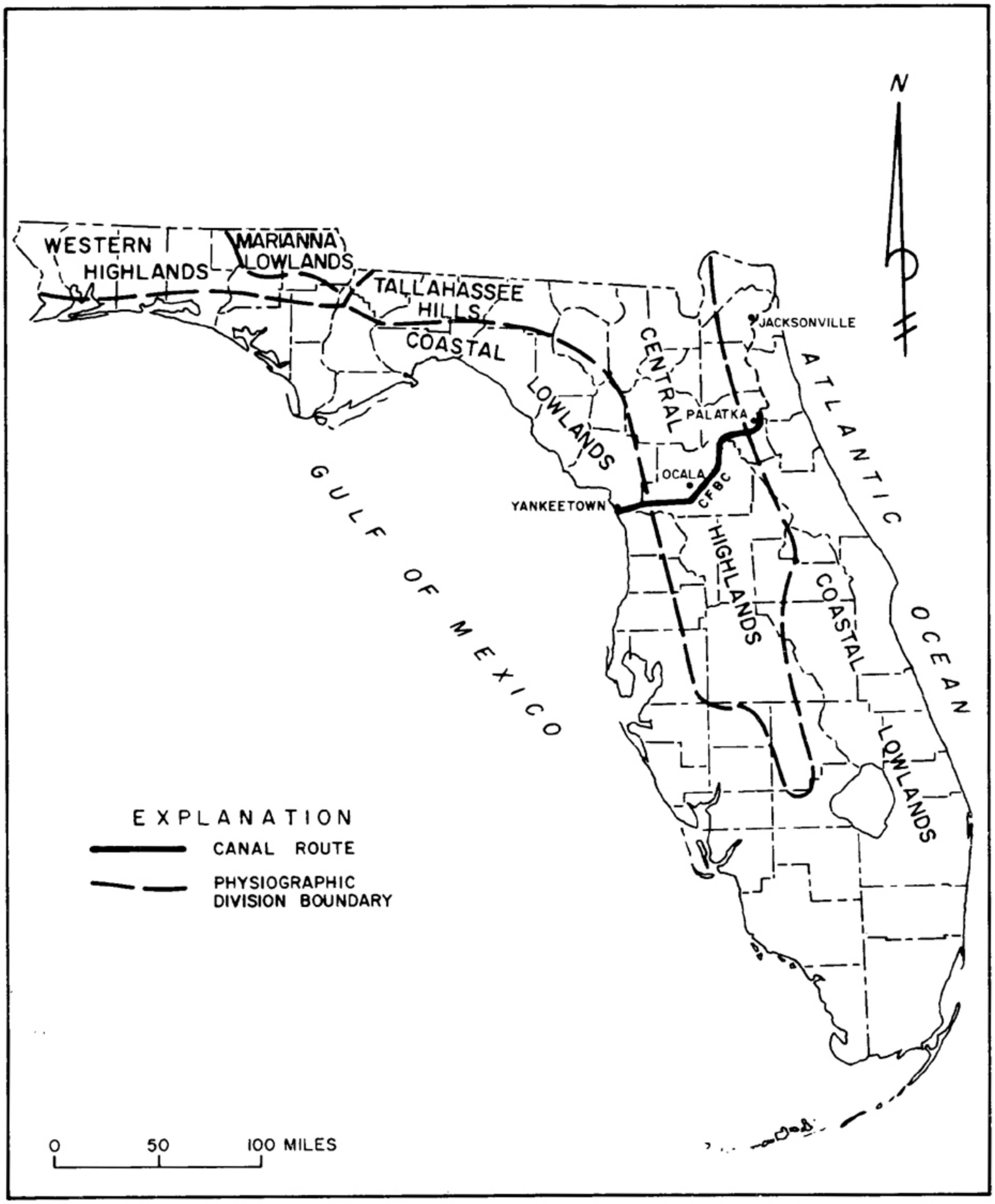


Figure 4. Physiographic divisions of Florida.

From Figure 5 it will be noted that the canal route at Palatka is situated in the Eastern Valley, a major subdivision of Cooke's Coastal Lowlands, where land surface altitudes are only a few feet above sea level. Southwestward along the canal route between the St. Johns and Oklawaha Rivers, surface altitudes increase to an average of about 25 feet above sea level and then decrease to less than 10 feet as the canal converges with the Oklawaha River near the Marion-Putnam County line. At this point the canal enters the Central Valley of the Central Highlands through the Kenwood Gap, and near Orange Springs it bears southwestward toward Ocala alone the flood plain the Oklawaha River.

From Orange Springs to the mouth of Silver River on the west side of the Central Valley, the flood plain of the Oklawaha gradually rises from an altitude of less than 15 feet to nearly 40 feet above mean sea level. The terrain of the Central Valley, where it is traversed by the Oklawaha River, tends to be nearly flat to gently rolling, and lakes, ponds, and swampy areas are common, especially on the southeast side of the river.

A few miles east of Ocala, the canal route enters the Sumter upland in southwest direction from the Central Valley. The Sumter Upland, which mostly ranges in altitude between 65 and 100 feet, gently rolling terrain characterized by numerous shallow sinkhole depressions that produce subdued karst topography. Often the depressions have permeable bottoms and generally do not contain perennial ponds or lakes.

Midway across the Sumter Upland, 8 to 10 miles south of Ocala, the canal turns to a due west course and passes through the summit part of its route across the Florida Peninsula. The highest natural land surface altitude along the centerline of the canal, about 110 feet above sea level, is in this area. As the route curves into its westward course, it passes between two comparatively high topographic features. The one to the north is identified as Ocala Hill by Puri and Vernon (1964, p. 15) and the one to the south is called Belleview Hill by this writer. Both features are comparatively hilly with numerous large sinkhole depressions, many of which contain ponds perched above the ground-water table. Except for some sink areas, land surface altitudes in both of these hill areas range from 100 to 150 feet above sea level.

Westward, just before the canal route enters the Western Valley of the Central Highlands, it crosses the southern tip of the Cotton Plant Ridge, a narrow, northwest trending feature ranging in altitude from 100 to 150 feet above sea level. From Cotton Plant Ridge the canal continues due west across the Western Valley to its confluence with the Withlacoochee River just south of Dunnellon. Where the canal route crosses the Western Valley, the terrain is flat to gently rolling, with an average altitude ranging from about 60 feet on the east to less than 50 feet near Dunnellon.

From Dunnellon, the canal alinement follows a modified course of the Withlacoochee River through the Dunnellon Gap, which allows passage of the Withlacoochee River through the Brooksville Ridge, a prominent, relatively high, physiographic feature, which forms the western boundary of the Central Highlands physiographic division. Altitudes in the gap range between 30 and 50 feet, while 1 mile to the south and 3 or 4 miles to the north, altitudes of about 150 feet above sea level are not uncommon in the Ridge area.

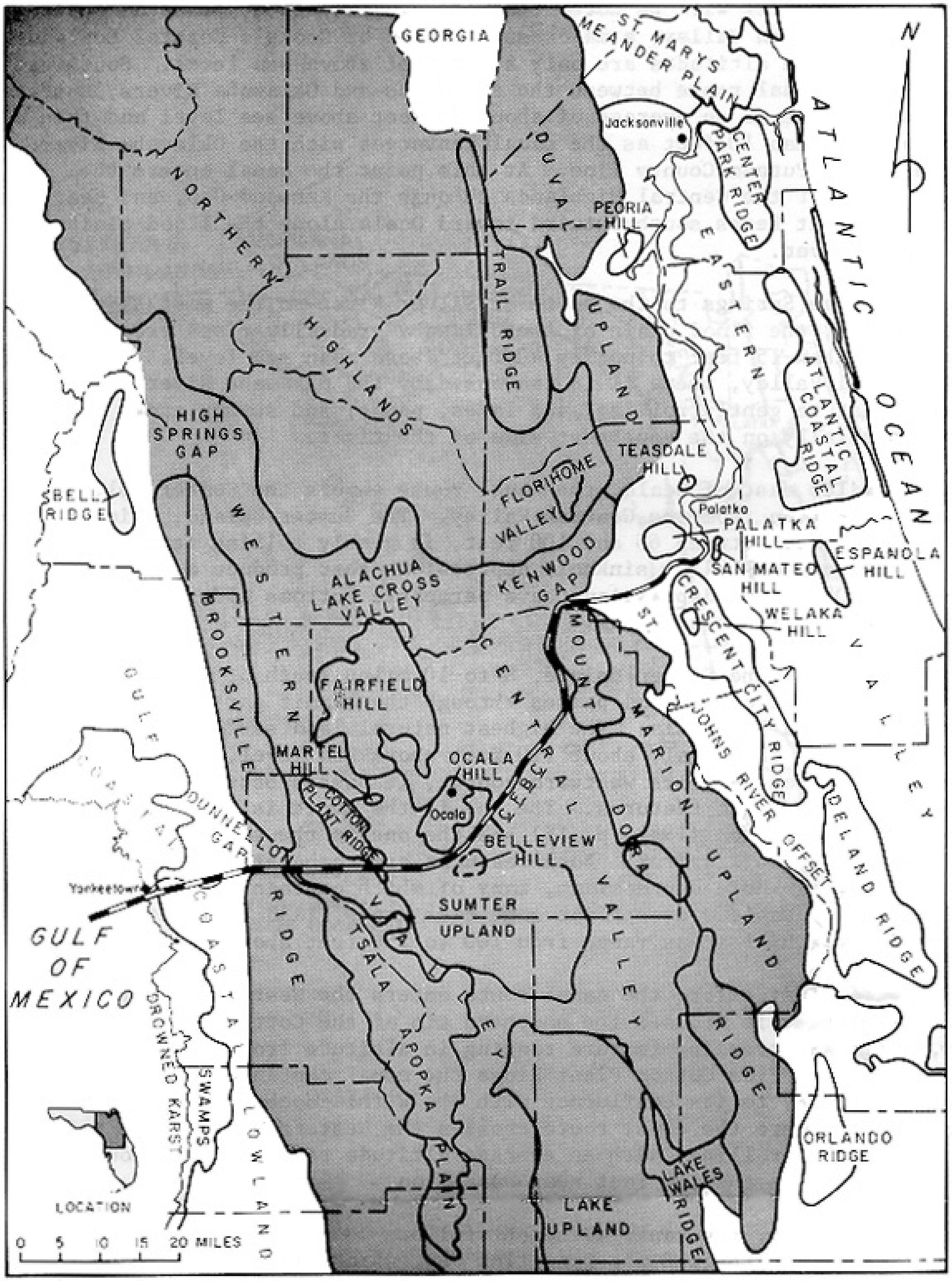


Figure 5. Physiographic subdivisions of north Peninsular Florida.

From the Dunnellon Gap, the canal bears to the southwest across the Gulf Coastal Lowlands, and thence through a narrow span of the Coastal Swamps before entering the Gulf, with the canal excavation terminating in deep water about 10 miles seaward from the coast.

Climate

The general climatic conditions of the Barge Canal area are subtropical. As shown in Figure 6, the route of the Cross-Florida Barge Canal traverses parts of the north and north-central climatic divisions of Florida (U. S. Weather Bureau, 1962, p. 24). Most of the area covered by the investigation is contained within the northern part of the north central division, although that part of the investigation area which extends into Levy County is considered to be in the north division.

The canal area has two distinctive seasons-summer, the rainy season and winter, the dry season. In a normal year over half the annual rainfall occurs during the 4-month period, June-September, although the rainy season in some years starts in early May and in some as late as the end of June.

Low latitude, nearness to the Ocean and the Gulf, and to some extent the presence of inland lakes, are the chief factors that cause warm humid summers and mild winters. However, cold air masses which occasionally move into Florida from the north generally preclude a winter free of frost and below freezing temperatures. Cold spells usually last only 2 or 3 days, however.

Rainfall

Rainfall, on the average, is abundant in the area, but it can be very unevenly distributed geographically, seasonally and from year to year. Excessively wet periods occur, as do extend droughts. According to Weather Bureau records for the 30-year period 1931-60, average annual rainfall for both the north-central and north climatic divisions (fig 6) is about 53 inches. Annual rainfall extremes for the 1931-60 period in the north central division range from 41.44 inches in 1956 to 73.21 inches in 1953.

Most summer rainstorms are local and of the convectional type (thunder-showers) that usually occur in the afternoon. The storms are shortlived, generally not more than 1 to 2 hours long, but they often yield large amounts of rain. The canal area is in a region that has more thunderstorms each year than any other part of the United States--80 to 90 such storms are expected annually, on the average. Winter rainstorms are commonly associated with cold fronts moving down from the north and such rainstorms are, therefore, rather widely distributed and of comparatively long duration.

Long lasting rainstorms during the summer months are uncommon, but, when they do occur they are usually associated with tropical storms or hurricanes. The greatest 24-hour rainfall recorded in Florida occurred at Yankeetown at the west end of the Barge Canal route, when 38.7 inches of rain fell on September 5-6, 1950, as the result of a hurricane. Two-hour rainfalls in excess of 3 inches and 24-hour amounts of around 10 inches occasionally occur in the study area.

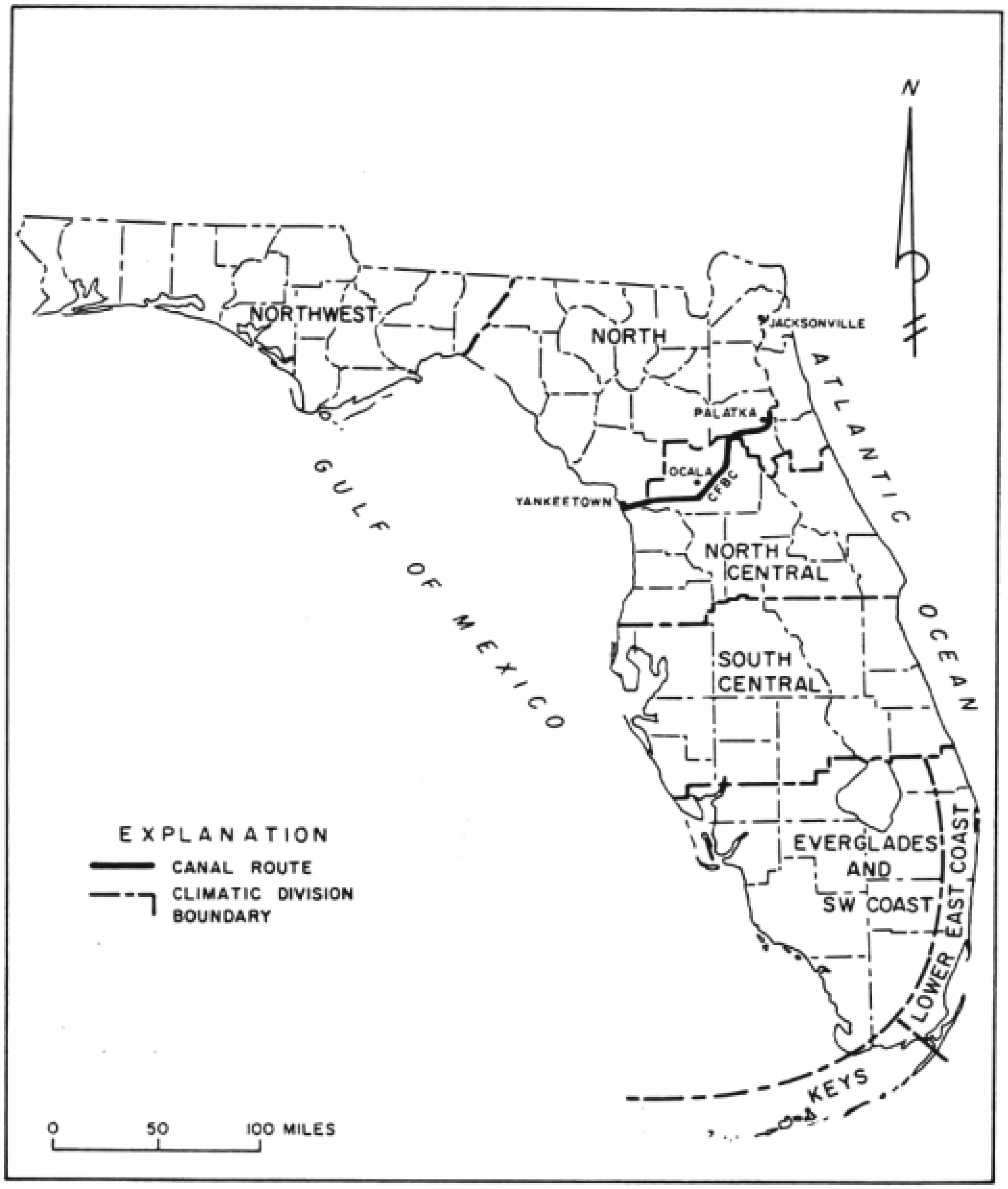


Figure 6. Climatic divisions of Florida.

Temperature

The annual average temperature for the north central division, as determined by the U. S. Weather Bureau for 1931-60, 30-year period, is 71.7°F (Fahrenheit). The coolest month is January, with an average temperature of 60.5°, and the warmest is August, with an average of 81.7°. Temperatures in the 90's during the summer are common, but temperatures above 100° are rare. Although average minimum temperatures during the coolest months range in the upper 40's, temperatures may be a few to several degrees below freezing several times during the winter. An occasional severe cold wave may drop temperatures to as low as 15° to 20°.

Winds

Wind directions in the area are often affected by local convectional forces, particularly in the summer when strong, but short-lived, local winds may accompany thunderstorms. Also, the "land and sea breeze" effect has some influence near the coast. Although prevailing directions are somewhat erratic, they follow a pattern: northerly in the winter and southerly in the summer. March and April tend to be the windiest months. Tornadoes may occur in any season, but they are not particularly prevalent in the canal area.

Tropical storms and hurricanes produce the principal high winds, and the winds of hurricanes can be quite destructive. Tropical storms are most likely to occur in the area in September and October, but they may be expected any time from June through November.

Humidity

The climate of the canal area is Florida. The inland areas have slightly coastal areas. In the summer relative during the night and from 50 to 65 percent during the afternoon. Except during the cooler months, heavy fog is generally limited to night and early morning hours.

Sunshine

The sun shines in the area about two-thirds of the time sunlight is possible. The frequency of sunshine ranges from 70 percent of possible in April and May, to 60 percent in December and January.

Drainage

The eastern part of the Cross-Florida Barge Canal area is drained to the Ocean by the Oklawaha and St. Johns Rivers, and the western part to the Gulf by the Withlacoochee River. The gradients of these streams are quite low, and poorly drained swampy areas are common in the flood plains of the streams. Most of the area between the Oklawaha and Withlacoochee River flood plains is drained internally, and has no interconnecting surface drainage system. A few swampy prairies are located in the area of internal drainage where the land surface is low and the water table is near the land surface.

Figure 7 shows that the canal route transverses, from northeast to southwest, the following three surface drainage basins as outlined by Kenner, Pride, and Conover (1967): (1) St. Johns River basin below Oklawaha River, (2) Oklawaha River basin, and (3) Withlacoochee River basin. In addition to the above listed drainage basins, the west end of the barge Canal area is no exception. Includes parts of the coastal drainage area between Withlacoochee and Suwannee River basins and between the Hillsborough Withlacoochee River basins. Important tributaries within the report area include Silver River and Orange Creek in the Oklawaha River basin, and Rainbow River in the Withlacoochee River basin.

As Kenner and others (1967) point out, most of Florida has little topographic relief; therefore, drainage divides are often difficult to delineate and frequently are not meaningful. The Barge Canal area is no exception. In addition to low relief in much of the area, geologic conditions have allowed development of a subdued karst or sinkhole topography, generally indicating good subsurface drainage which further complicates definition of surface drainage basins. In much of the Barge Canal area subsurface drainage takes the place of surface drainage.

Much of the area of investigation drains internally, either directly into the numerous local closed depressions of the subdued karst terrain, or by direct seepage into the limestone or shallow sand aquifer through permeable surficial materials on the topographic highs separating the depressions. Many of the sinks have permeable bottoms which permit rapid seepage into the highly porous cavernous limestone which underlies them. Some are open directly to the limestone, and therefore allow direct flow into the aquifer. Others have clay bottoms which retard or practically prevent the flow of water into the aquifer.

The subsurface drainage system is well developed, and the boundaries of the ground-water drainage areas do not necessarily correspond to the surface drainage basin divides (fig. 7). For example, a considerable section of the northwest part of the topographically defined Oklawaha River basin drains underground by way of the Floridan Aquifer to Rainbow Springs in the Withlacoochee River basin, and thence to the Gulf instead of to the Ocean by way of the Oklawaha and St. Johns Rivers. Neither does the Oklawaha River receive surface runoff from that part of its topographically defined basin. Other similar situations exist in the study area.

Silver and Rainbow Springs are very important features of the drainage system of the Barge Canal area. Silver Springs, a few miles east of Ocala, discharges an average of 531 mgd (million gallons per day) from the Floridan Aquifer. Silver River, which heads at the spring, carries the discharge a few miles eastward to the Oklawaha River. The flow rainfall which has drained or seeped into the subsurface mostly in northeastern and central parts of the Oklawaha River basin. Rainbow Springs discharges, on the average, 468 mgd to the Withlacoochee River. This is the accumulated subsurface flow from recharge areas not only in the northwestern Oklawaha River basin, but in part of the eastern coastal area between the Withlacoochee and Suwannee Rivers, and in the extreme north end of the Withlacoochee River basin.

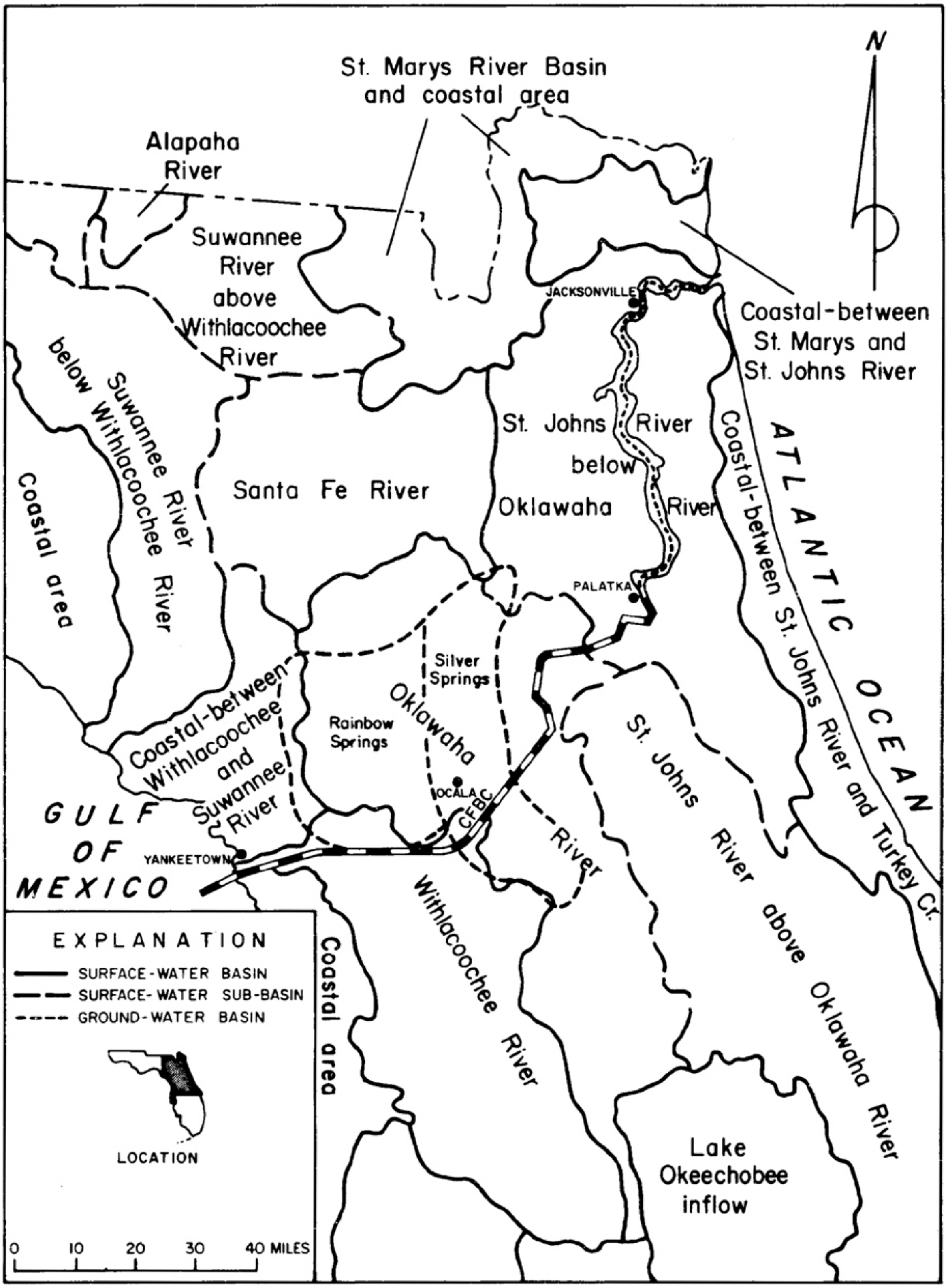


Figure 7. Drainage basins in north Peninsular Florida.

Orange Creek, which drains into the Oklawaha River near Orange Springs at the northeast end of the study area, is an important part of the surface drainage system. It conveys overflow from Orange Lake and, during periods of high runoff, receives significant amounts of surface water inflow along its course from Orange Lake to the Oklawaha River. A few other small tributaries drain sinkhole lakes or swampy areas in or adjacent to the flood plains of the Oklawaha and Withlacoochee Rivers. Probably the most important of these is Eaton Creek, which drains the Lake Eaton-Mud Lake area and empties into the Oklawaha about 3 miles south of Eureka.

A system of drainage wells drilled into the limestone of the Floridan Aquifer has been developed in the city of Ocala in the bottoms of partly plugged sinkholes and excavated retention ponds. The drainage wells were drilled to augment the natural internal drainage as storm runoff increased with urban expansion. However, present State regulations against the drilling of additional drainage wells, because of the of pollution of ground-water supplies, precludes further expansion of the drainage well system. A few other drainage wells are no doubt located within the study area, but Ocala is the only place where there are many them.

GEOLOGY

General

The Florida Peninsula, a prominent southeastward protrusion of the North American Continent, forms the eastern boundary of the Gulf, and separates the Gulf from the Ocean. The peninsula is about 400 miles long and 150 miles wide at its widest point. The peninsula is the emergent part of a much larger feature called the Floridan Plateau (fig. 8), the submerged parts of which form the continental shelves surrounding Florida. The plateau extends southward only a few miles beyond the Florida Keys, but its overall width averages nearly twice that of the emergent peninsula. The shelf area off the east coast of Florida is comparatively narrow, and at its narrowest point, off southeast Florida, is only about 10 miles wide and ends abruptly in the Straits of Florida. On the other hand, the continental shelf, as delineated on Figure 8 by the 300-foot water-depth contour, is extensive off the west coast of Florida. Its widest part is located off the southwest tip of the peninsula, where it extends about 185 miles westward into the Gulf. Off Yankeetown, at the west end of the Cross-Florida Barge Canal, the shelf is nearly 150 miles wide.

Geologically, the Floridan Plateau is a large, more or less tectonically stable, carbonate platform on which have accumulated thick deposits of Cretaceous and Tertiary limestones and dolomites, some evaporites, and comparatively small amounts of clastic sediments. These rocks were deposited in shallow transgressive and regressive seas over the southeastward plunging Peninsular Arch (fig. 8), the principal subsurface structural axis of the peninsula, which has been referred to by some as the “backbone” of the Florida Peninsula. The Arch is generally believed to have originated during late Paleozoic or early Mesozoic time from crustal stresses, which caused very gentle warping or arching of the Coastal Plain Floor, a nearly flat, possibly peneplaned surface underlain by early Paleozoic and Precambrian(?)

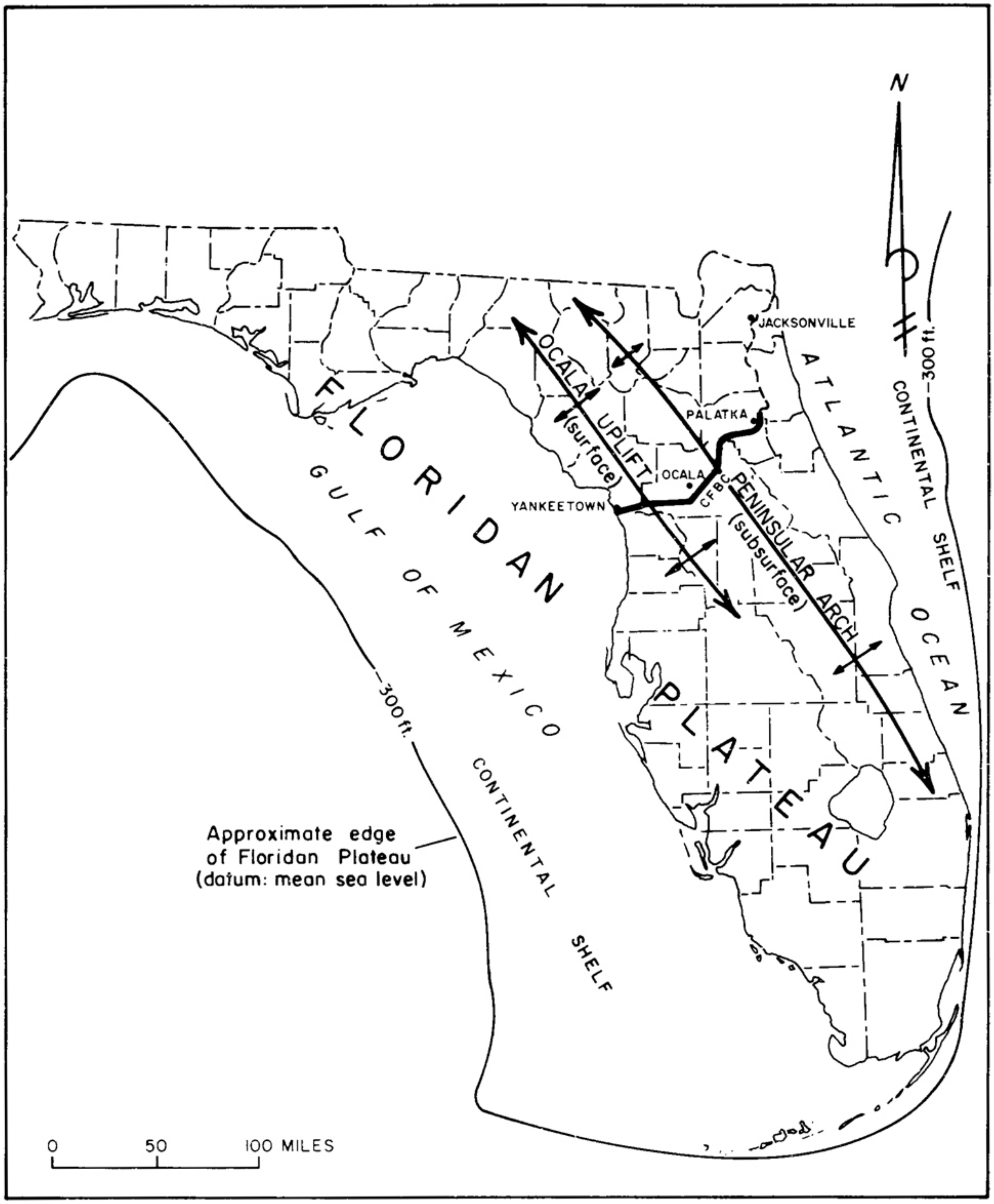


Figure 8. Index map to major structural features of Florida which affect the area of the Cross-Florida Barge Canal.

sedimentary and igneous rocks (Applin, 1951, p. 3). It is possible, too, that the Arch could have developed during Cretaceous and early Tertiary time due to differential subsidence of the Coastal Plain Floor under the accumulating load of shallow sea sediments which now comprise the Floridan Plateau. The route of the Barge Canal probably passes over the axis of the Arch somewhere between Orange Springs and Ocala.

Possibly as early as late-middle Eocene time, the axis of warping shifted some 30 to 35 miles westward, to a line approximately parallel to the trend of the Peninsular Arch. This younger axis of gentle folding is expressed at the surface in the form of the Ocala Uplift (fig. 8), a broad, rather ill defined, tension faulted anticlinal structure mappable in exposures of middle and upper Eocene limestone and dolomite (fig. 9). Again, the origin of this new structural axis is not clear, but it could have resulted from gentle upwarping of the early Tertiary rocks due to shallow lateral compressional stresses, or possibly it was simply a shift in the pattern of differential subsidence of the Coastal Plain Floor. A discussion of the possible causes of this subsidence is beyond the scope of this report. The canal route crosses the axis of the Ocala Uplift in the vicinity of the Dunnellon Lock site.

Beds of Oligocene age are not present in the canal area, although the Suwannee Limestone of Oligocene age is present in Peninsular Florida some distance from the canal area down both the north and south plunges of the Ocala Uplift (fig. 8).

During Miocene and possibly the early part of Pliocene time, shallow marine clastic and deltaic sediments were deposited all along the northeast flank of the Ocala Uplift, and probably to some extent over the crest of the structure although the writer knows of no Miocene deposits mapped between the anticlinal axis and the Gulf Coast in the canal area. They may have been removed by later erosion, of course, but it is thought by some (Vernon, 1951) that the crest area of the structure was emergent during much of Miocene and probably all of Pliocene time, and, therefore, no marine sediments were deposited on the apex of the structure at that time. As may be seen on Figure 9, Miocene-Pliocene(?) rocks are present far down both plunges of the Ocala Uplift, as well as down the northeast flank of the structure.

Much of the rock shown on the map as being either at or near land surface is actually covered by a veneer of Pleistocene sand and clayey sand distributed as marine terrace deposits at high stands of the sea during interglacial periods. However, some of the sandy cover is material from the leaching of Miocene and younger deposits

Figure 10 is a geologic section (X-X') along a northeast to southwest line drawn across north Peninsular Florida through the area of the Cross-Florida Barge Canal. The section illustrates the general stratigraphic and structural relationships that exist from the land surface down into the early Paleozoic or Precambrian rocks that underlie the Coastal Plain Floor. Although the oldest stratigraphic unit to be penetrated by the Barge Canal excavation is the Avon Park Limestone of middle Eocene age, a knowledge of the geology of the deeper strata in the area is essential to a proper understanding of the geohydrologic relationships. As is discussed in detail later in

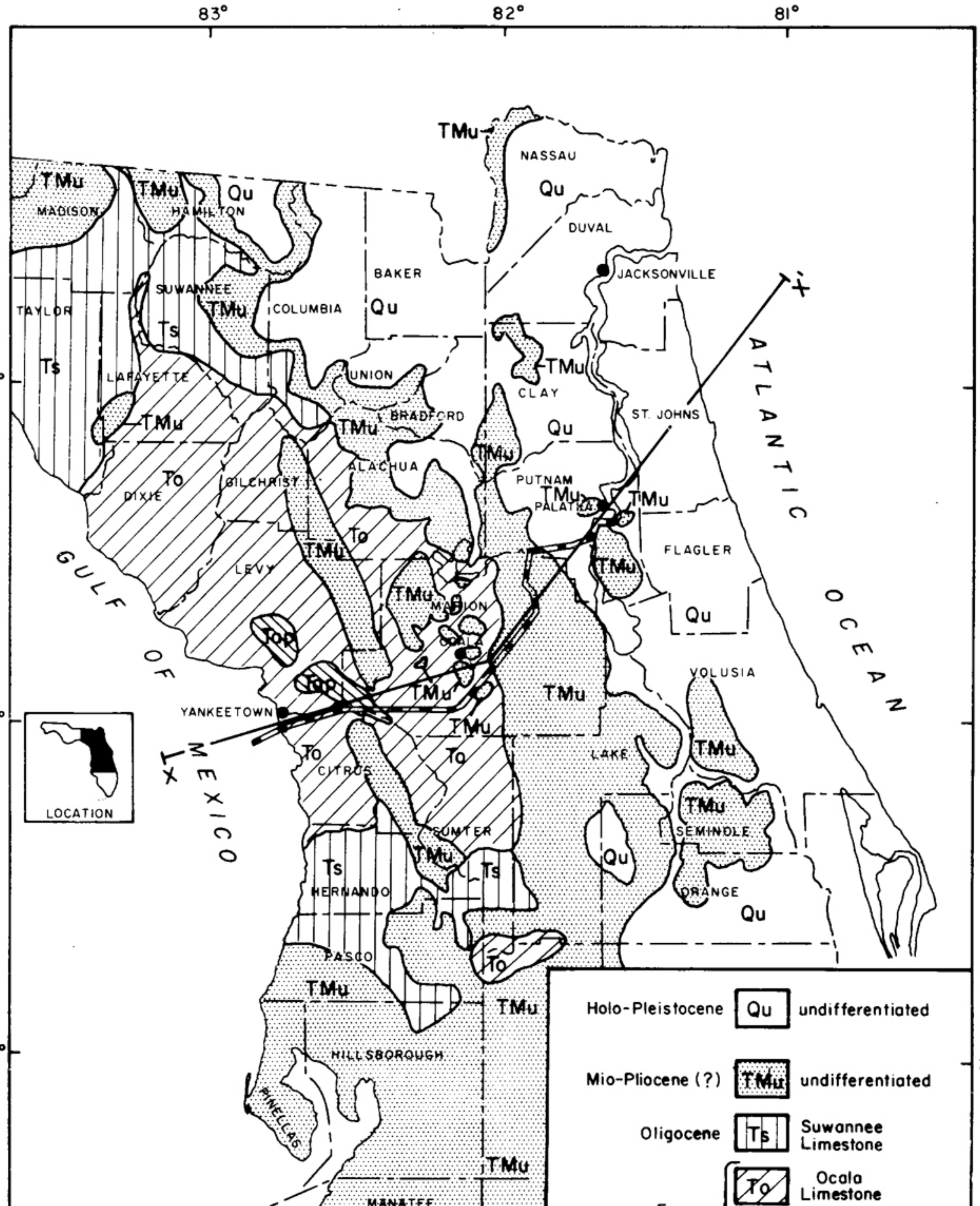


Figure 9. Geologic map of north Peninsular Florida showing rocks present at or near the surface of the ground.

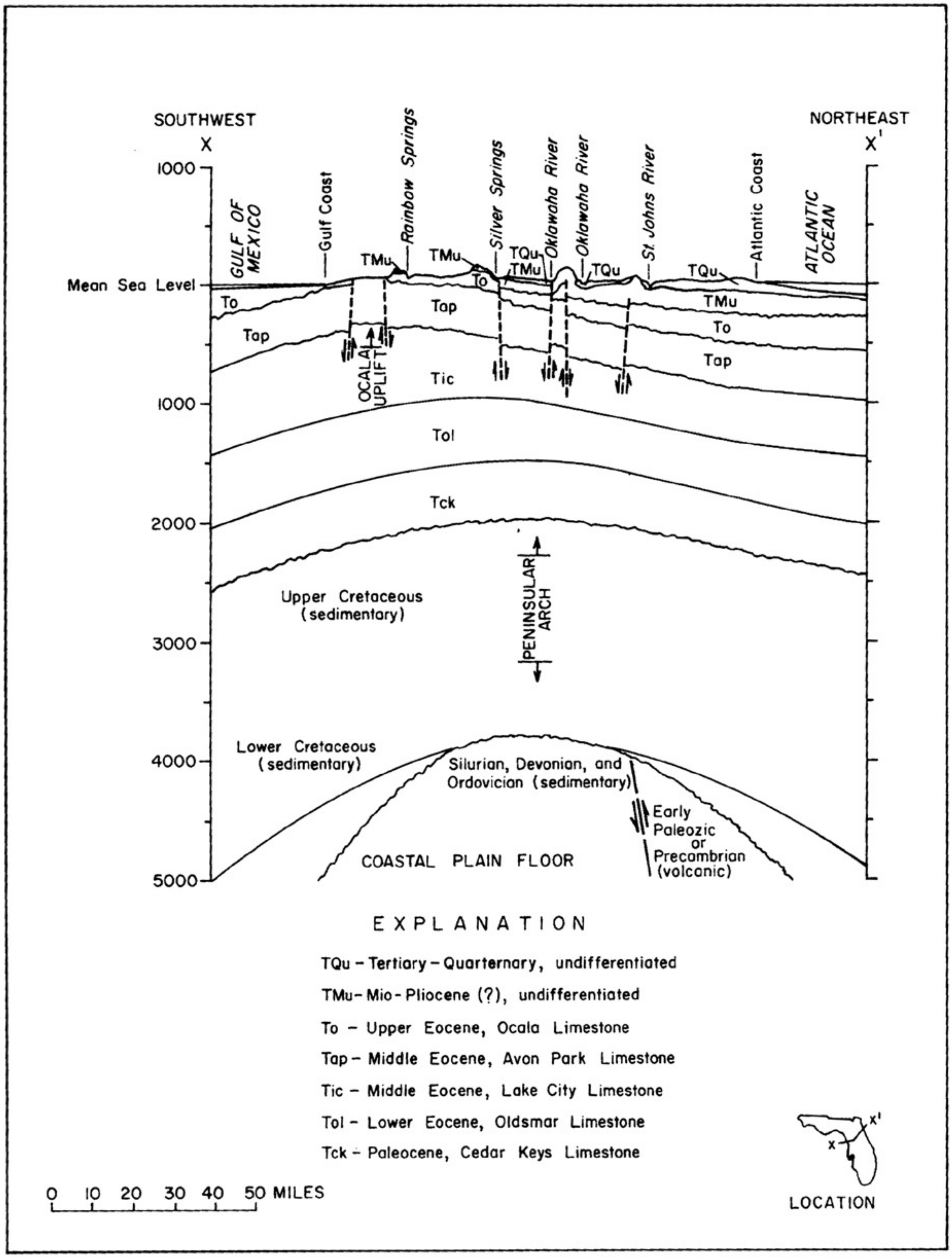


Figure 10. Geologic section (X-X') of north Peninsular through area of Cross-Florida Barge Canal.

this report, the relationships of the structure and Cenozoic stratigraphy of the Ocala Uplift have had a pronounced effect upon the hydrology of the Cross-Florida Barge Canal area.

When examining section X-X' (fig. 10), it should be kept in mind a the vertical scale is exaggerated about 160 times. Actually the flanks of both the Peninsular Arch and the Ocala Uplift have very low dips. Maximum true dip along the line of section at the Coastal Plain Floor is about one-half degree. Regional dips of the Mesozoic and younger beds are, therefore, considerably less than one-half degree.

Stratigraphy

In the Barge Canal area, about 4,000 feet of sedimentary rocks (Applin and Applin, 1965, p. 17, fig. 3), mostly carbonates, overlie the post-Paleozoic Coastal Plain Floor. Figure 11 describes the stratigraphic section in the area of investigation, from the rocks of early Paleozoic or Precambrian age, which underlie the Coastal Plain Floor, to the Quaternary deposits at land surface.

Pre-Cenozoic

Since the Barge Canal is directly concerned only with rocks of early Tertiary age and younger, if suffices to say here, for the pre-Cenozoic rocks, that some 1,500 to 2,500 feet of Lower and Upper, mostly Upper, Cretaceous carbonates and subordinate amounts of clastic sediments uncomfortably overlie the Coastal Plain Floor. The Cretaceous is in turn uncomfortably overlain by a thick sequence of early Tertiary marine carbonate rocks topped by a much thinner section of marine, transitional, and terrestrial clastic deposits of late Tertiary and Quaternary age.

The cumulative thickness of Cenozoic deposits overlying the Cretaceous beds in the canal area ranges between 2,000 and 2,500 feet (figs. 10 and 12).

Early Tertiary

The basal Tertiary stratigraphic unit in the area is the Cedar Keys limestone of Paleocene age. It is 400 to 700 feet thick and consists mostly of light gray dolomite and considerable amounts of anhydrite and gypsum.

Conformably overlying the Cedar Keys Limestone are 600 to 700 feet of the lower Eocene Oldsmar Limestone, a unit composed mostly of light brown to chalky white limestone with some interbedded dolomite and minor amounts of anhydrite and gypsum.

Conformably overlying the Oldsmar Limestone is the basal unit of the principal artesian aquifer of the southeastern United States (Stringfield, 1966, p. 30-31, table 3), that thick secession of permeable often cavernous limestones and dolomites referred to in Florida as the Floridan Aquifer. The Floridan Aquifer underlies all of the state and is the principal source of ground water in Florida, except in the southernmost and western most parts, and in some east coastal areas (Hyde, 1965). The sequence of limestones comprising the Floridan Aquifer ranges from about 1,000 feet to 1,200 feet in

Era	System	Series	Stratigraphic Unit	Thickness (feet)	Lithology	Aquifer
CENOZOIC	QUATERNARY	Holocene	Unnamed alluvial, lake, and windblown deposits	0-10 or 20	Alluvium, fresh water marl, peats and muds in stream and lake bottoms. Also, some dunes and other windblown sand.	
					unconformity	
		Pleistocene	Pamlico and other marine and estuarine terraces	0-100±	Mostly marine quartz sand, unconsolidated, and generally well graded. Also, some fluviatile and lacustrine sand, clay, marl and peat deposits.	Shallow aquifer
					unconformity	
	TERTIARY	Miocene-Pliocene(?)	"Jackson Bluff formation"	0-75±	Marine sands, argillaceous, carbonaceous; and sandy shell marl. Some phosphatic limestone.	
		?Miocene-Pliocene(?) / Middle Miocene-Pliocene(?)	Alachua Formation	0-100±	Nonmarine interbedded deposits of clay, sand, and sandy clay; much of unit is phosphatic, base characterized by rubble of phosphate rock and silicified limestone residuum in a gray and green phosphatic clay matrix.	
			"Fort Preston formation"	0-100±	Nonmarine fluviatile sand, white to gray, variegated orange, purple and red in upper part, fine to course grained to pebbly, clayey, cross-bedded.	
		Miocene, lower & ?middle	Hawthorn Formation	0-140	Marine interbedded sand, cream, white and gray, phosphatic, often clayey; clay, green to gray and white, phosphatic, often sandy; dolomite, cream to white and gray, phosphatic, sandy, clayey; and some limestone, hard, dense, in part sandy and phosphatic. Tends to be sandy in upper part and dolomitic and limy in lower part.	
					unconformity	
		Eocene, upper	Ocala Limestone 1/: Upper member 2/	0-100±	Marine limestone, cream to white, soft, granular, highly porous; coquinal, often consists almost entirely of tests of formanifers; cherty in places.	Floridan aquifer
			Ocala Limestone 1/: Lower member 3/	0-80±	Marine limestone, cream to tan and brown, granular, soft to firm, porous, highly fossiliferous; lower part at places is dolomite, gray and brown, crystalline, sacchoroidal, porous.	
					unconformity	
		Eocene, middle	Avon Park Limestone	200-400	Marine limestone, light brown to brown, finely fragmental, poor to good porosity, highly fossiliferous (mostly foraminifers); and dolomite, brown to dark brown, slightly porous to good porosity, crystalline, saccharoidal; both limestone and dolomite are carbonaceous or peaty; gypsum is present in small amounts.	
					unconformity	
			Lake City Limestone	600-700	Marine limestone, light brown to brown, fragmental, highly fossiliferous, slightly carbonaceous or peaty and cherty; and dolomite, brown to dark brown with very minor amounts of gypsum and anhydrite. Unit is slightly porous to porous.	
		Eocene, lower	Oldsmar Limestone	500-650	Marine limestone, light brown to chalky, white, porous, fossiliferous, with interbedded brown, porous, crystalline dolomite; minor amounts of anhydrite and gypsum.	
		Paleocene	Cedar Keys Limestone	400-700	Marine dolomite, light gray, hard, slightly porous to porous, crystalline, in part fossiliferous, with considerable anhydrite and gypsum, some limestone.	
					unconformity	
MESOZOIC	UPPER and LOWER CRETACEOUS			1500-2500	Mostly marine Upper Cretaceous carbonates, evaporites, sands and shales; thin Lower Cretaceous clastic section in some of area, but absent over crest of Peninsular Arch in northeast part of area.	
					Coastal Plain Floor	
PALEOZOIC and PRECAMBRIAN	DEVONIAN to PRECAMBRIAN(?)			Basement rocks	Marine Devonian, Silurian, and Ordovician quartzose sandstone and dark shale, except extreme east and southeast portion of Barge Canal area, where Early Paleozoic (?) or Precambrian (?) rhyolite, tuff, and agglomerate occur.	

1/ Ocala Group of Bureau of Geology, Florida Department of Natural Resources.
2/ Crystal River Formation of Ocala Group.
3/ Inglis Formation and Williston Formation (older to younger) of Ocala Group.

Figure 11. Description of stratigraphic section in area of Cross-Florida Barge Canal.

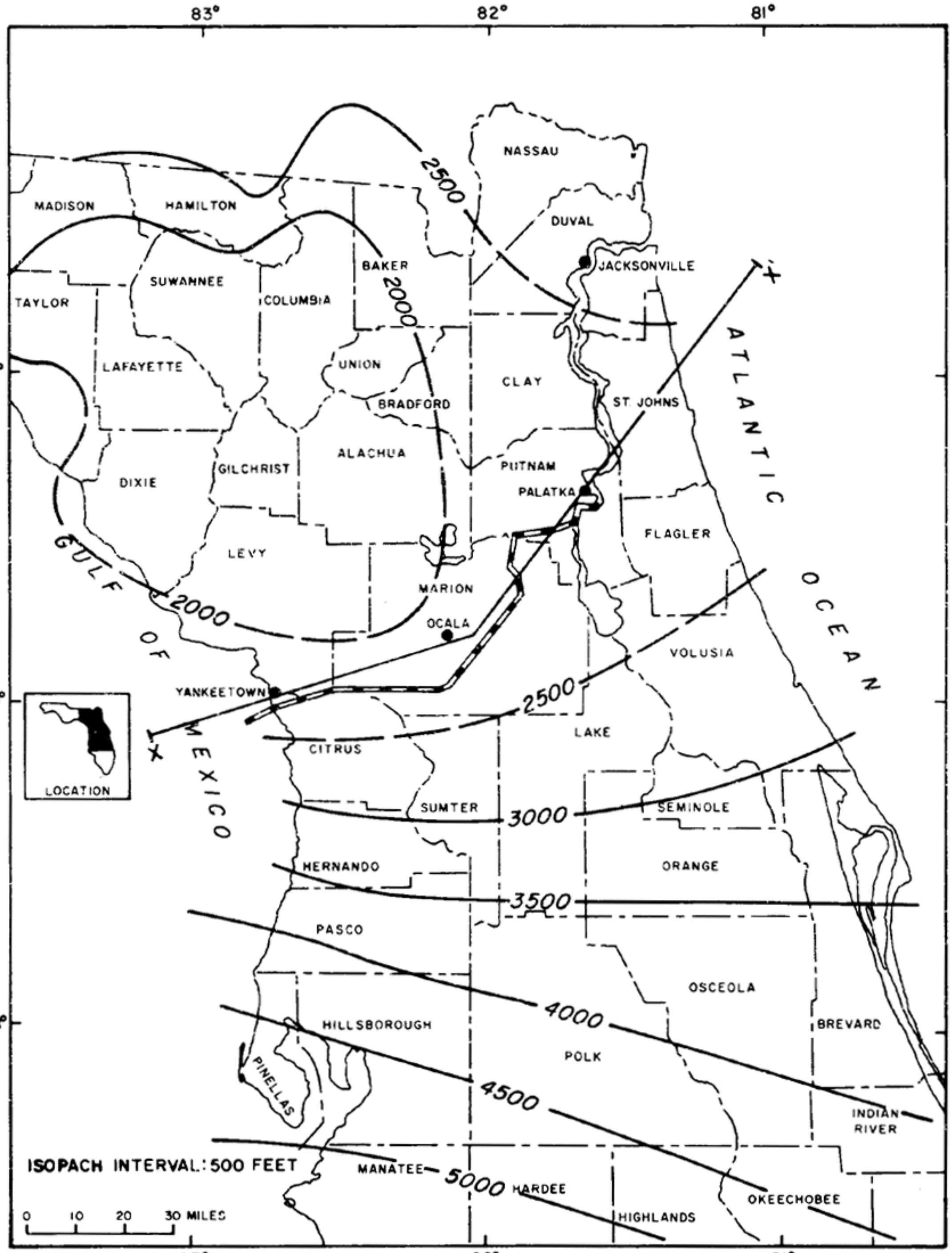


Figure 12. Isopach map of Cenozoic sediments in north Peninsular Florida.

thickness in the Barge Canal area, and ranges in age from middle Eocene to middle Miocene. The bottom of the aquifer is rather arbitrarily drawn at the base of the middle Eocene Lake City Limestone. The underlying Oldsmar Limestone contains salt water in many parts of Florida, but probably not in some parts of the canal area. Based on the height of the potentiometric surface, the fresh-water-salt-water interface is estimated to occur in the Cedar Keys Limestone in the Ocala vicinity. The youngest rocks comprising the aquifer are permeable limestones and' dolomites in the lower part of the lower and middle Miocene age Hawthorn Formation. In much of Florida where the aquifer is under artesian conditions, it is overlain by clastic beds of low permeability in the upper part of the Hawthorn Formation.

The Lake City Limestone, or basal unit of the Floridan Aquifer, is 400-700 feet thick in the area and consists of porous brown limestone and dolomite, sometimes peaty and cherty, and it contains minor amounts of gypsum and anhydrite. Dolomitization of the Lake City Limestone is widespread in north-central Florida (Chen, 1965, p. 60 and fig. 33).

The Avon Park Limestone of middle Eocene age unconformably overlies the Lake City Limestone, and is the oldest stratigraphic unit to be penetrated by the Barge Canal excavation. Rocks of the Avon Park Limestone are the oldest naturally exposed in the Barge Canal area, and for that matter in the whole state of Florida. The Avon Park is present at or near the surface on the crest of the Ocala Uplift in the Dunnellon area, near the west end of the canal route. The area of outcrop includes a small part of southwestern Marion County and a sizeable part of southeastern Levy County (Figure 13).

The Avon Park, in the area of the canal, consists of 200-400 feet of brown, finely fragmental limestone and dolomite with poor to good porosity. It is highly fossiliferous, containing mostly foraminifers, and is commonly carbonaceous or peaty. Usually gypsum is present only in small amounts

As in the case of the Lake City Limestone, Chen (1965, p. 60 and fig. 33) shows that dolomitization of the Avon Park Limestone is widespread in north-central Peninsular Florida. The Avon Park is separated from the overlying Ocala Limestone by an erosional unconformity (Vernon, 1951, p. 99), and the present writer thinks that regional dolomitization of at least the upper part of the Avon Park could have occurred, during a period of slight emergence in middle Eocene time, from geochemical processes similar to those described by Goodell and Garman (1969).

The upper Eocene Ocala Limestone overlies Avon Park Limestone throughout the canal area, except where the Avon Park occurs at or near the land surface. The absence of the Ocala in such places is probably due mostly to removal by erosion from the crest of tile Ocala Uplift. As may be seen on Figures 9 and 13, the Ocala occurs at or near the surface over most of the Barge Canal area from near Silver Springs west to the Gulf Coast.

The Ocala Limestone is one of the more productive formations of the Floridan Aquifer, and it is probably the most important single source of water for the Cross-Florida Barge Canal. Except for a few sand filled sinks and other erosional depressions, the canal will be excavated to depths below the

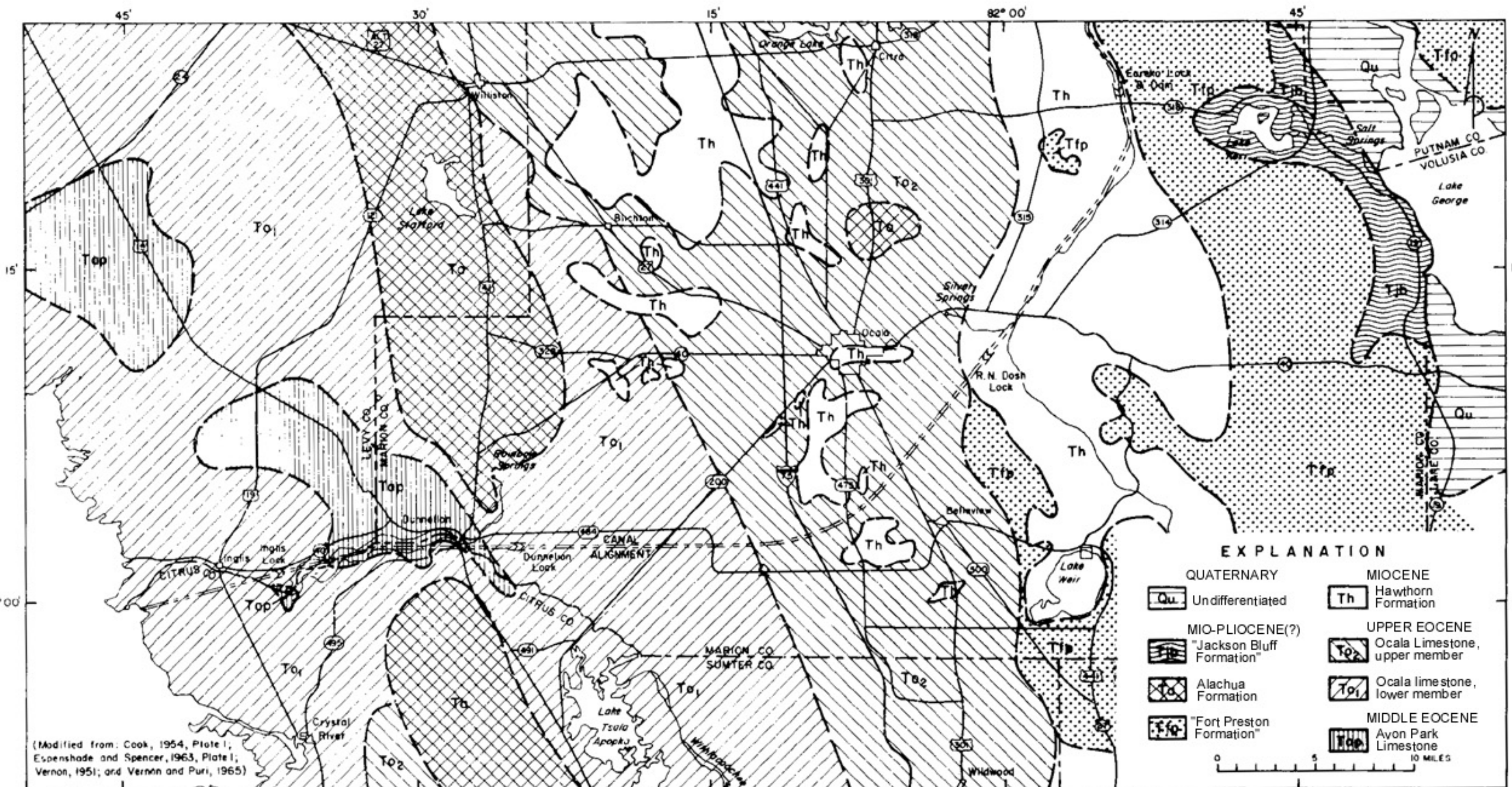


Figure 13. Geologic map showing stratigraphic units at or near land surface, Cross-Florida Barge Canal area.

water table in the Ocala Limestone, or the Avon Park Limestone, from a mile or two southwest of R. N. Dosh Lock to the Gulf (fig. 13). Also, water from the Ocala Limestone is the principal source of flow in surface streams during low flow.

In the Floridan Peninsula, the Ocala Limestone may be divided into an upper and a lower member based on lithologic and faunal differences (Applin and Applin, 1944, p. 1679 and Stringfield, 1966, p. 45). The Bureau of Geology, Florida Department Of Natural Resources, recognizes the Ocala as the Ocala Group consisting of three formations (Puri, 1957, p. 24-30). From older to younger, the Ocala Group includes the Inglis, Williston and Crystal River Formations. The lower member of the Ocala Limestone of Applin and Applin is equivalent to the Inglis and Williston Formations, and the upper member is equivalent to the Crystal River Formation.

The lower member of the Ocala Limestone, a few to about 80 feet think in the canal area, consists of granular, highly fossiliferous to coquinal, tan brown limestone, the lower part of which frequently consists of gray and brown dolomite. The upper member where present conformably overlies the lower member, and is a few feet to about 80 feet thick. It consists mostly of soft, granular, very fossiliferous, cream to white limestone. In places both the upper and lower members are coquinas consisting almost entirely of the tests of foraminifers. In places the Ocala is cherty, more commonly near its top, but erratic cherty zones may occur at any depth in the unit. The chert is not consistently present at any given horizon and is usually found in irregularly shaped masses, or as relatively thin layers of limited areal extent.

Late Tertiary

From the Silver Springs area west, Ocala limestone is shown on Figures 9 and 13 to be present at or near land surface. However, the Ocala, and for that matter the Avon Park Limestone, are, but for a few actual exposures of very limited areal extent, covered by at least a veneer of sand. In the Ocala vicinity, actual exposures are mostly on hilltops or upland flats at surface altitudes of about 100 feet above mean sea level, the average altitude at which the contact occurs between the Ocala and remnants of overlying Hawthorn Formation of Miocene age. Locally, the contact between the Miocene beds and the Eocene limestones is very irregular, as the top of the limestone is a post-Eocene and pre-Miocene erosional surface complicated by post-Miocene slumping as a result of ground-water solution and consequent removal of Eocene limestone. Limestone pinnacles sometimes protrude 10 to 20 feet or more above the 100-foot level into the overlying Miocene beds; and Miocene beds slump into sinkhole depressions to a few tens of feet below the 100-foot level.

Where the Ocala Limestone is not overlain by sandy limestone, clay, clayey sand and sand of Miocene and possible Pliocene age, most of the Eocene limestone surface is usually at an altitude well below 100 feet, and is generally covered by a few tens of feet of deposits of Quaternary age consisting of unconsolidated sand and clayey sand. These sediments are a combination of marine terrace deposits and of nonmarine residual deposits resulting from intensive leaching of Miocene-Pliocene(?) beds which probably at one time covered most of the report area.

Figures 9 and 13 show that the Miocene-Pliocene(?) cover is continuous east of an approximately north-south line between Silver Springs and the Oklawaha River. West of this line only outliers of the once continuous upper Tertiary cover remains.

On Figure 14, the formations present at or near the surface in the Ocala vicinity, especially the Mio-Pliocene (?) outliers, are shown in more detail than on Figure 13. Also, the thickness of material overlying the limestones and dolomites of the Floridan Aquifer in the Ocala vicinity is shown in detail on Figure 15.

Beds of Oligocene age are absent in the canal area. The oldest post-Eocene stratigraphic unit to occur is the Miocene Hawthorn Formation, which unconformably overlies the Ocala Limestone. The Hawthorn or some other Miocene-Pliocene(?) stratigraphic unit, such as the Fort Preston formation, is thought to be present in most places in the area east of the north-south line between Silver Springs and the Oklawaha River, but only scattered outliers of the Miocene-Pliocene(?) materials remain in the Ocala vicinity west of the line. (See figs. 9, 10, 13, and 14.)

The Hawthorn Formation ranges from a few feet to about 140 feet in thickness and consists of shallow marine and estuarine white to gray and cream phosphatic sand and clayey sand, green to gray and white phosphatic, often sandy, clay, and white and gray, sandy, sometimes clayey, phosphatic limestone and dolomite. The formation tends to be characteristically sandy in its upper part and limy and dolomitic in its lower part. A greater part of the Hawthorn consists of carbonate rocks, mostly dolomitic. In the eastern part of the area where the formation is thick, than where it is thinner, as in, the outliers in the central and western part of the Ocala vicinity. Here limestone beds a few feet thick are found only at or near the base of the formation; phosphatic sand, clayey sand and clay comprise the greater part of the unit (Espenshade and Spencer, 1963, p. 18-22). Permeable limestones and dolomites in the lower part of the Hawthorn constitute the uppermost part of the Floridan Aquifer. The clays and clayey sands of low permeability in the upper part of the Hawthorn confine the artesian part of the Floridan Aquifer.

In the Mount Dora Ridge area and in the vicinity of Lake Weir (figs. 5 and 13), the Hawthorn Formation is overlain and (or) perhaps interfingered with more than 100 feet, in places, of marginal marine coarse clastic sediments of middle Miocene age and younger. These sediments are varicolored white, pink, lavender, tan, orange, etc. and consist of poorly sorted quartz grains, ranging from fine grained to small pebble size, often present in a clay matrix. The name "Fort Preston formation" has been suggested for this unit by Pun and Vernon (1964, p. 185-186). These sediments had previously been correlated by Cooke (1945) with the Pliocene Citronelle Formation of the western panhandle of Florida. Sediments mapped as "Fort Preston formation" in this report are interpreted by some geologists as remnants of a large delta elongated by long shore currents that swept southward along a former peninsular east coast (White, 1958).

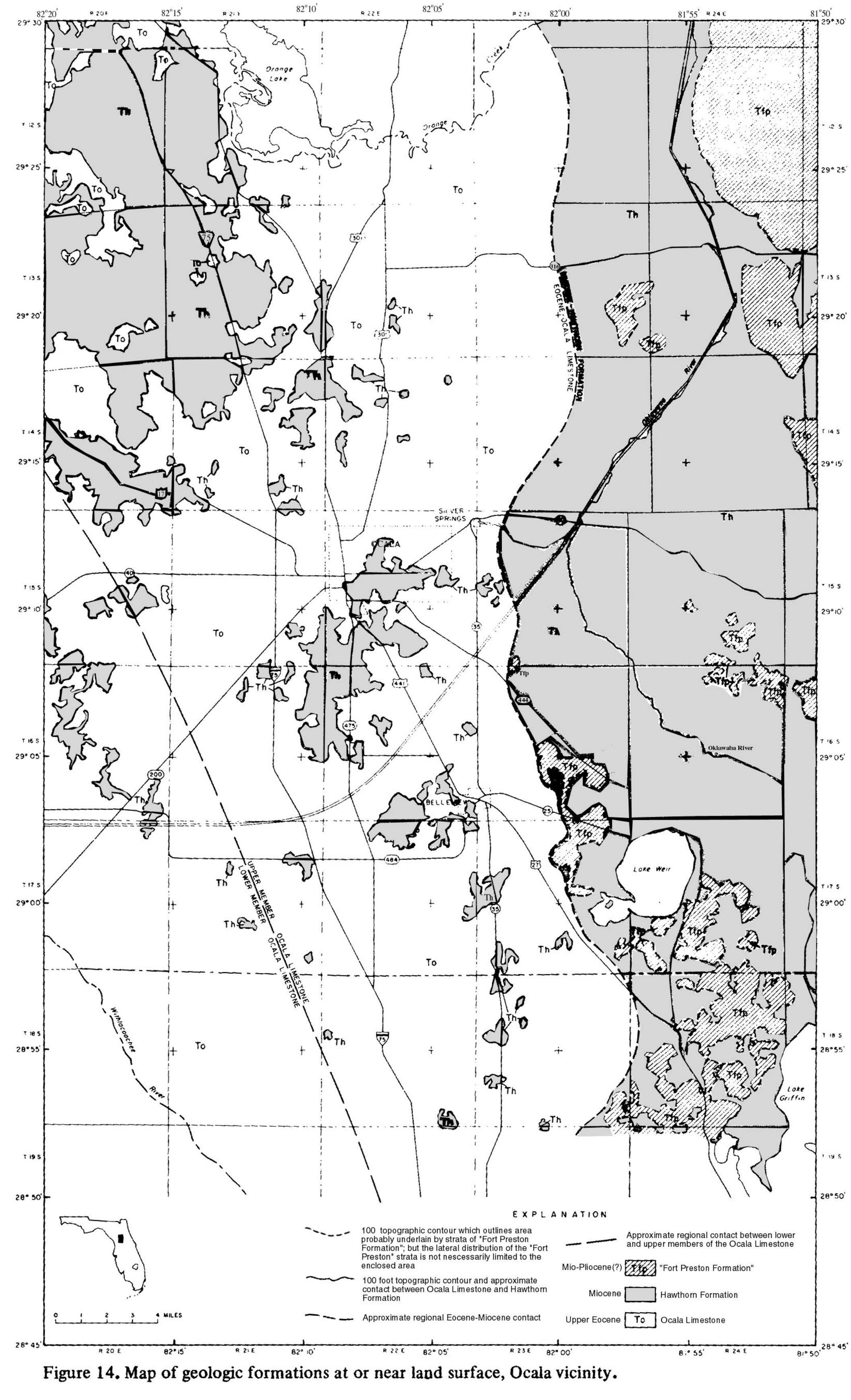


Figure 14. Map of geologic formations at or near land surface, Ocala vicinity.

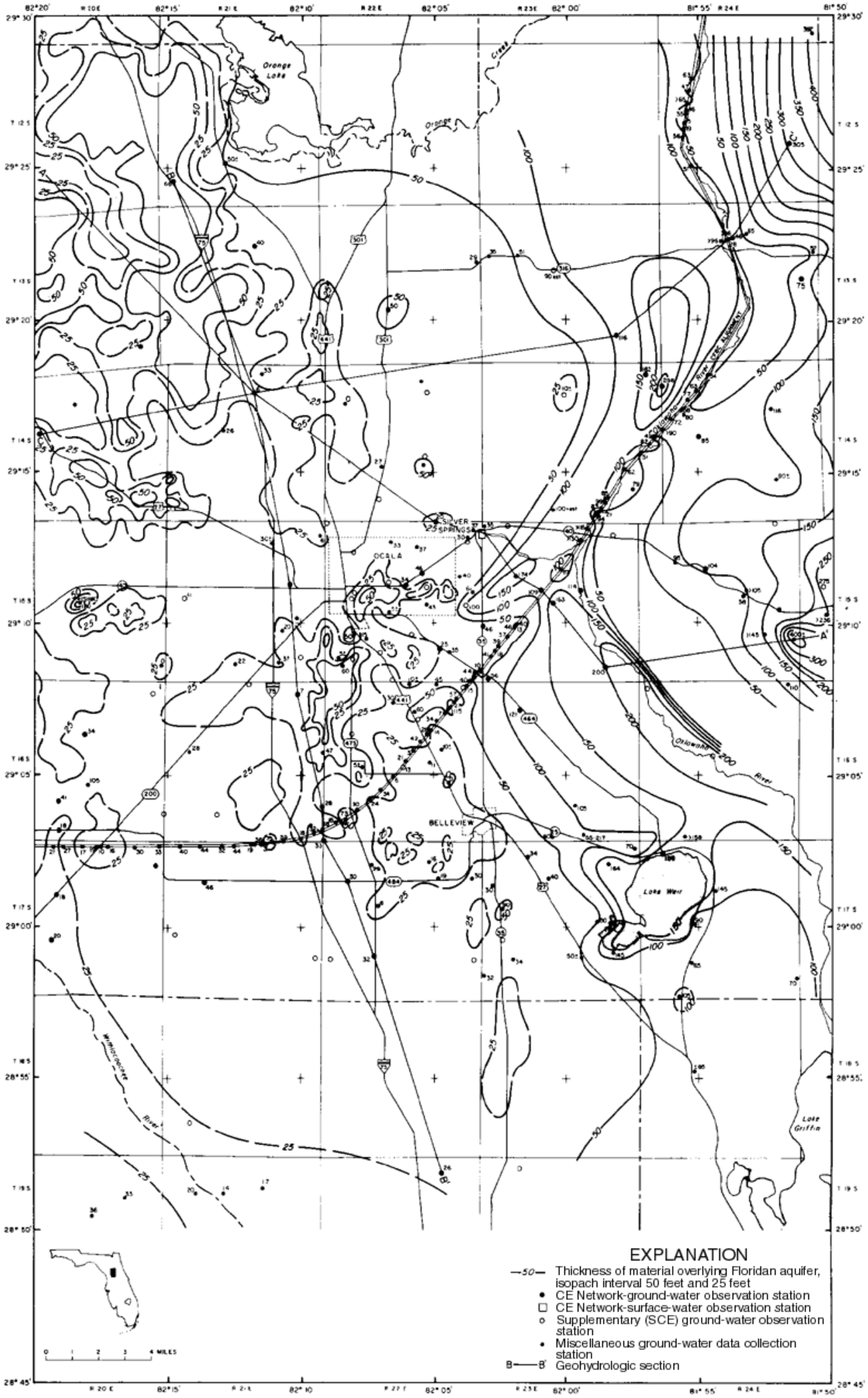


Figure 15. Thickness of material overlying Floridan aquifer, Ocala vicinity.

On the ridge east of the Eureka Dam, the Corps of Engineers found in one or two core holes possible Ocala Limestone directly underlying clayey sand of the Fort Preston formation. However, in the same general locality other test holes penetrated sandy dolomite of the Hawthorn Formation. Also, other sparse drill hole data support the probability that in most places east of the Oklawaha River, where the Fort Preston formation directly overlies the aquifer, it is in contact with sandy, phosphatic dolomites of the Hawthorn Formation, rather than with the Ocala Limestone.

Other Miocene-Pliocene(?) stratigraphic units mapped in the study area include the Alachua Formation and the informally named "Jackson Bluff formation." Both units are considered here to be of Miocene and possible Pliocene age and younger than lowermost Hawthorn, but they could be in part contemporaneous with later Hawthorn.

In the western part of the canal area, most of the northwest-southeast trending Brooksville Ridge (Figures 5 and 13), has been mapped as Alachua Formation (Cooke, 1945 and Puri and Vernon, 1964). The Alachua in the study area is estimated to be as much as 100 feet thick. The formation consists of irregularly interbedded deposits of clay, sand, and sandy clay. Much of the unit is highly phosphatic, and is characterized at its base by a rubble of phosphate rock and solicited limestone residuum which is the source of hard-rock phosphate mined in the area. Some workers consider the Alachua Formation to be residual from erosion of the marine Hawthorn Formation (Cooke, 1945, P1. 1 and Espenshade and Spencer, 1963, p. 8), but Vernon (1951) considers the Alachua to be terrestrial in origin, and that it was deposited over the crest of the Ocala Uplift at least in part contemporaneously with offshore deposition of Hawthorn beds.

The "Jackson Bluff formation" of Miocene-Pliocene(?) age is present in the extreme eastern end of the study area in the vicinity of Lake Kerr, Salt Springs and Juniper Springs (Figure 11). The unit name has been informally suggested by Puri and Vernon (1964) for beds previously mapped as Duplin Marl of Miocene age by Cooke (1945, p1. 1). The unit is considered younger than lower Hawthorn, and where present it probably overlies Hawthorn beds. The Jackson Bluff consists of argillaceous and carbonaceous sands and sandy shell marl with some phosphatic limestone or dolomite.

Quaternary

During Pleistocene time sea level alternately rose and fell in response to the effect on the world's water supply of the several glacial and interglacial periods. During periods of glaciation, sea level at times fell far below present day level as tremendous quantities of water were stored on the continents in the form of great glaciers. During warmer periods, when the glacial ice melted, the increased supply of water to the sea caused the sea to rise and inundate much of present day peninsular Florida. Various workers, including Cooke (1945, p. 248-312) and MacNeil (1950) have mapped what they believe to be remnants of several Pleistocene marine terraces developed on the Florida Peninsula during stands of the sea higher than the present shoreline. One classification of Pleistocene shorelines in Florida recognizes nine different stands of the sea ranging from the oldest at 270 feet to the youngest at

6 feet above present sea level (Stringfield, 1966, p. 68). However, other workers (Alt and Brooks, 1965) contend that the remnants of an old shoreline indicated at 90-100 feet was occupied during Pliocene time. Alt and Brooks believe that a shoreline at 25-30 feet is the highest of definite Pleistocene age, and that shoreline features at 45-55 and 70-80 feet are representative of either Pliocene or early Pleistocene interglacial periods. In any case, during several interglacial high stands of the sea in Plio-Pleistocene time, all or part of the Barge Canal area was inundated, and the Mio-Pliocene sediments were eroded and redeposited, or were reworked on the shallow sea bottom to form marine terraces.

In accordance with Cooke's classification (Stringfield, 1966, p. 68), the most recent time the sea would have made an appreciable advance into the canal area was during the Sangamon interglacial periods, when sea level was 25 feet above present level. At this time the sea probably crept up the Oklawaha Valley to near Eureka.

During high stands of the sea, deposits of well graded quartz sand accumulated several tens of feet thick to form terraces. Also, some fluviatile and lacustrine sand, clay, marl and peat were deposited in shoreward areas.

Since the close of Pleistocene time, some 10,000 years ago, the canal area has been continuously above sea level. Deposits of Holocene age in the study area consist of comparatively thin beds of alluvium, fresh water marl, peats and muds in stream and lake bottoms; and some windblown sand deposits occur in the area.

As indicated in Figure 11, the permeable sand, and in some cases shell beds, of Miocene to Holocene age in the Cross-Florida Barge Canal area constitute the "shallow aquifer" (Hyde, 1965). In most places in the area, the shallow aquifer either yields little water, or it is not practical to utilize the aquifer. But in some places, particularly east of the Oklawaha River, where the Floridan Aquifer is not easily accessible, domestic wells in the shallow aquifer often yield adequate quantities of water.

Structure

Folding

The route of the Cross-Florida Barge Canal passes over two long, doubly plunging, essentially parallel northwest trending anticlinal axes. The one of most direct significance to the canal is the Ocala Uplift; however, the Peninsular Arch, the axis of which is some 30 miles east the Ocala Uplift is the older of the two structures and is intimately involved in the structural history of the Barge Canal area.

In the general discussion of geology, was pointed out that the Peninsular Arch was the primary structural control in the area for the distribution of sedimentary rocks laid down during Cretaceous and early Tertiary time. Probably sometime in Oligocene or early Miocene time, but possibly early as late-middle Eocene time, crustal stresses caused a very gentle elongate upwarp of early Tertiary beds centering along a line parallel to, but located well down the west flank of the axis of the Peninsular Arch.

This new bulge or upwarp, the Ocala Uplift, may have resulted directly from shallow compressional forces, but it could have been related to differential subsidence due to sediment loading of the Floridan Plateau (fig. 8). That is, the subsidence of the Peninsular Arch area may have accelerated for some reason at the expense of the area of the Ocala Uplift. In any case, structurally speaking the Ocala Uplift has been relatively positive in its movements since Eocene time, while the Peninsular Arch has been relatively negative.

The axis of the Peninsular Arch as determined by drilling extends for about 300 miles southeast from the Georgia border through its apex in Union County to just east of Lake Okeechobee. The Barge Canal route probably crosses the axis of the Arch between Orange Springs and Ocala.

The axis of the Ocala Uplift parallels the Peninsular Arch for nearly 200 miles through the northern half of Florida. It extends from Madison County southward through its apex in Levy and Citrus Counties to Polk County. The Barge Canal route crosses the axis between the Dunnellon and Inglis Lock sites.

The Ocala Uplift is mappable in exposures of Eocene and younger rocks (fig. 16), but the axis of this broad gentle upwarp is rather poorly defined. The extremely low dip of its flanks of only a few feet per mile, its flat crest, and irregularities resulting from normal faulting and tension fracturing have tended to complicate, but in certain respects have aided (Vernon 1951, p. 47), detailed delineation of the structure. However, in spite of its low relief, the structure has had a pronounced effect upon the present distribution of Eocene and younger rocks at or near the surface in north Peninsular Florida.

Fracturing and Faulting

As the broad but elongate gentle arch of the Ocala Uplift developed, stretching of rock strata over the top and down the flanks of the structure resulted in tensional stresses sufficient to cause widespread, near vertical fracturing and sometimes normal faulting of the strata. Such fracturing is especially evident in near surface occurrences of the limestones and dolomites of the Floridan Aquifer in the Barge Canal area.

A two-fold system of fractures is expressed at the surface by various types of lineation which can be recognized and traced from aerial photographs with comparative ease. Vernon (1951, p. 47-52) was the first to map such a fracture and fault system in Florida. He found two sets of fractures (fig. 17) - a primary one with a generally northwest trend, and a secondary one trending northeast. The two sets of the system intersect at broad or nearly right angles. The primary set parallels the axis of the Ocala Uplift, whereas the secondary set trends in the approximate direction of dip of the flanks of the Uplift. Vernon found, mostly from test hole data along the centerline of the Cross-Florida Barge Canal, that many of the fractures paralleling the axis of the structure were actually faults with vertical displacements in some cases exceeding 100 feet. He had insufficient data to determine whether appreciable movement had taken place along the secondary or transverse set of fractures.

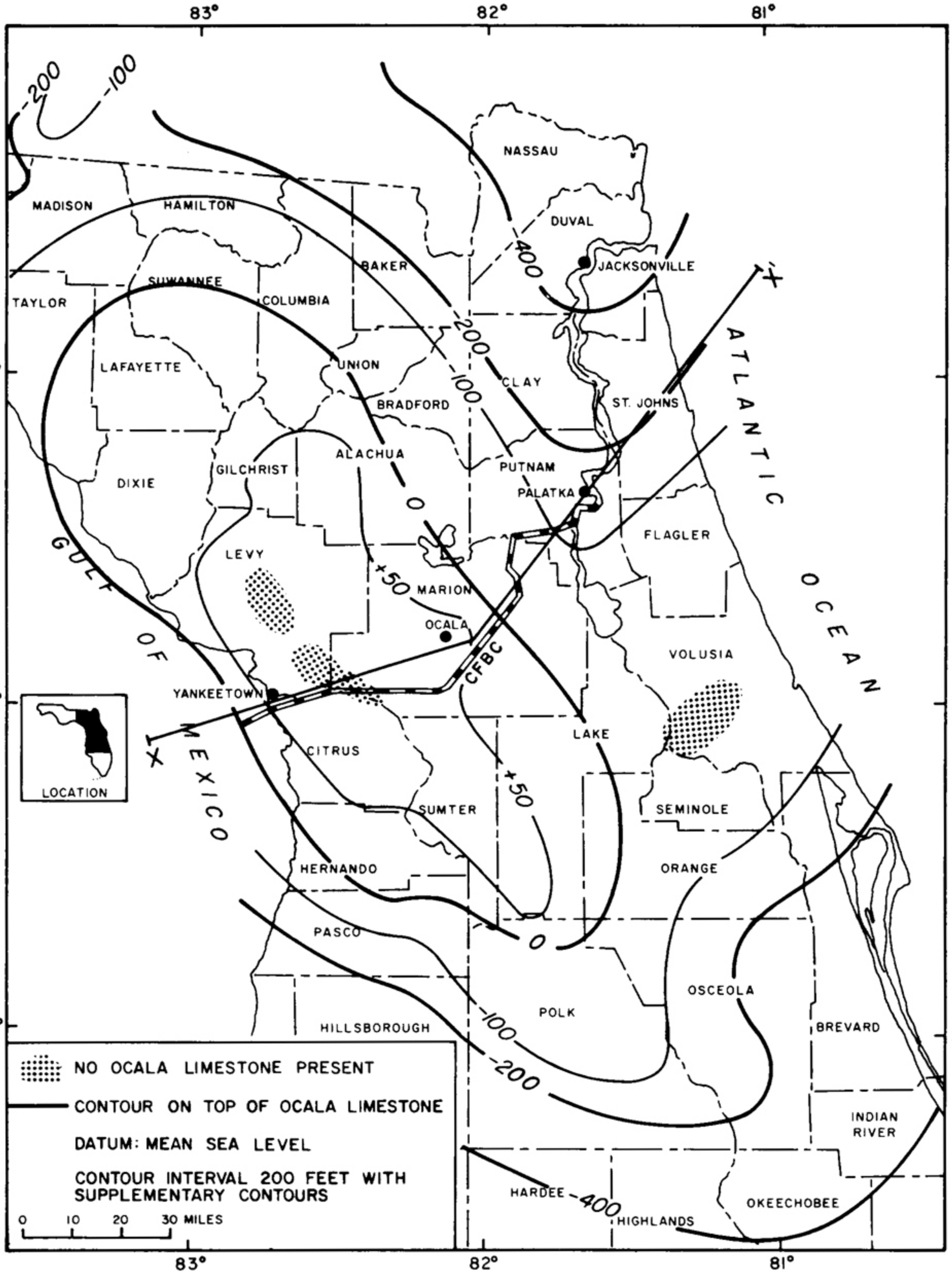


Figure 16. Contour map of top of Ocala Limestone, north Peninsular Florida.

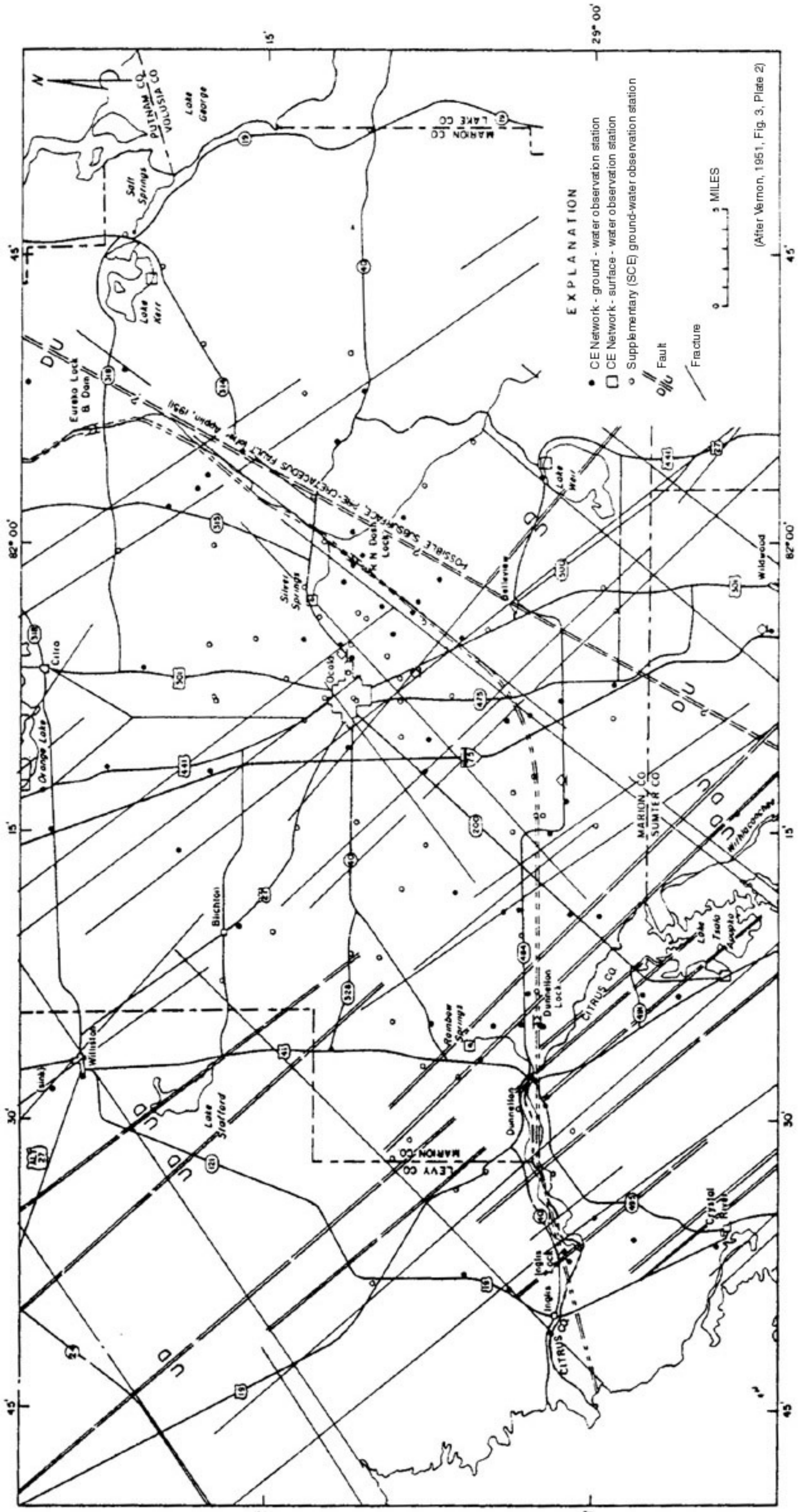


Figure 17. Fractures and faults mostly as interpreted from surface lineations and alignments observed on aerial photographs, Cross-Florida Barge Canal area.

A possible pre-Cretaceous, deep-seated, subsurface fault which is not necessarily expressed by surface lineation is shown on Figure 17. This possible fault was suggested by Applin (1951) to help explain differences in the lithologies of the Coastal Plain Floor, as shown in cross section X-X' (fig. 10).

In addition to those faults and fractures mapped by Vernon in the Barge Canal area and shown on Figure 17, the writer has examined aerial photographs of the Ocala vicinity and has shown on Figure 18 abundant liniations interpreted as surface traces of near vertical fractures. Appreciable movement may have occurred along some of these fractures.

Mapping the fracture lineations in the Ocala vicinity aids in determining principal directions and routes of ground-water in the Floridan Aquifer. The writer and others have observed that most caverns and solution channels in the limestone are oriented along near-vertical fractures having trends of fracture systems mapped at the surface. The logical inference is that water moving through the aquifer tends to follow the line of least resistance or greatest permeability, which in this case is along the fractures. In general, the greatest solution of limestone at shallow depths below the water table takes place where the greatest amount of water moves through. Thus cavities are developed as the walls of fractures are dissolved away by recently recharged ground water with a high carbon dioxide (CO_2) content. The relationship of the fracture and fault system to the subsurface drainage system, as well as to the past and present drainage system of the canal area will be discussed in detail in this report.

The top of the Floridan Aquifer, commonly referred to as the "top-of-rock", has been mapped by the writer in the Ocala vicinity (fig. 19). The map is not a map of the geologic structure, as it is not drawn on a consistent time-stratigraphic horizon, rather it depicts the top of materials that range from the lower member of the Ocala Limestone to sandy limestones and dolomites in the Hawthorn Formation. The map also indicates possible faults, a knowledge of which is important to the interpretation and understanding of the geohydrology of the area.

Using the top-of-rock map, the maps of fracture traces, thickness of aquifer cover, potentiometric surface, and a general knowledge of the geohydrology of the area, the writer has inferred a system of normal faults in and adjacent to the Oklawaha River valley east of Ocala and Silver Springs. More stratigraphic data and study will be necessary to fully substantiate and define details of the fault system, but the map as it stands helps to show the system as an important hydrologic control of both the ground-water and the surface-water regimens, as well as for the Barge Canal itself.

The fault system apparently controls the rectilinear course of the Oklawaha River, and it lowered the geologic section east of Ocala and Silver Springs. Poorly permeable strata in the Hawthorn are preserved in the structurally lowered river valley, but were removed by erosion from the higher areas to the west. Therefore, the Floridan Aquifer is confined in the down-faulted area, and is more or less unconfined in the structurally high area to the west (figs. 9, 13, and 14). Also in small areas east of the Oklawaha River valley, where upthrown fault blocks are covered by permeable sands of

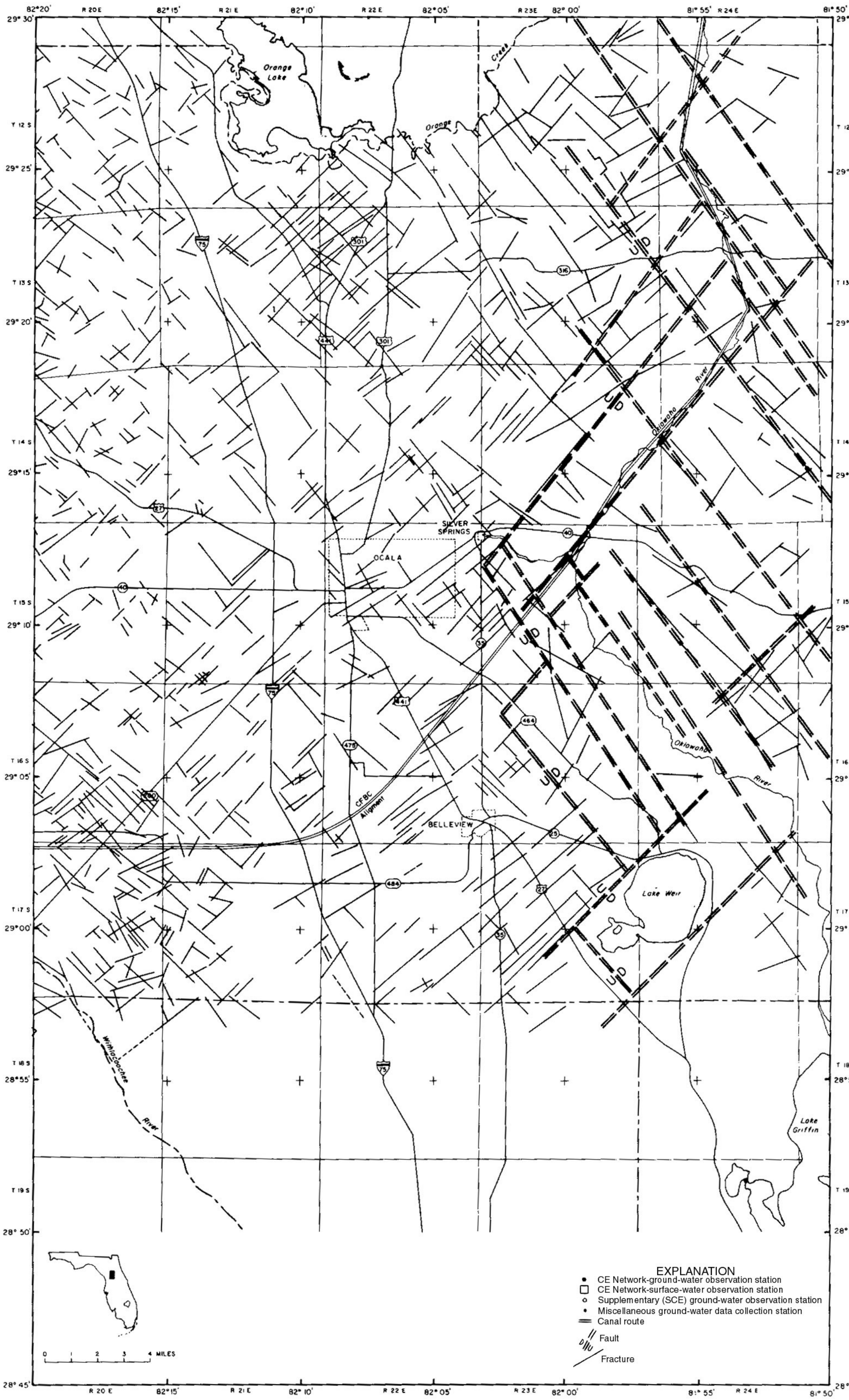


Figure 18. Fracture traces and some possible faults, Ocala vicinity.

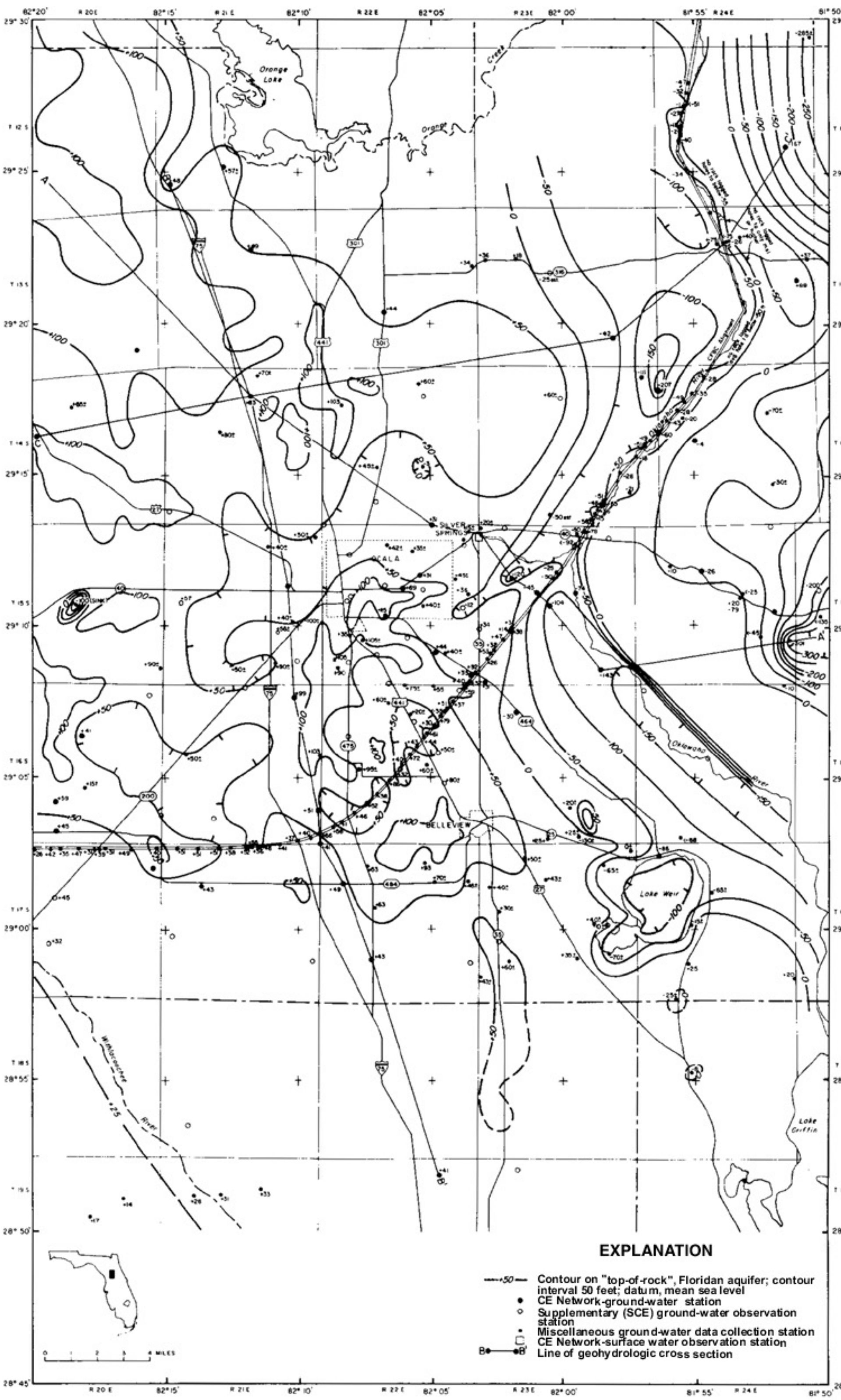


Figure 19. Top-of-rock - the Floridan Aquifer, Ocala vicinity.

the Fort Preston formation, the aquifer is unconfined. These aquifer-fault-confining layer relationships are shown by geohydrologic sections A-A’, B-B’, C-C’, and D-D’ (fig. 20).

Faults also occur in the highly fractured limestones of the structurally high area west of the Oklawaha River valley. Vernon (1951, fig. 14) recognized some faulting southwest of Ocala from studies of test holes along the centerline of the canal route. However, the writer believes that faults are more common and of considerably greater displacement in the eastern low area. As demonstrated by Vernon (1951) and shown on Figure 17, faults are again common along the crest of the Ocala Uplift in the Dunnellon area.

The multiple normal faulting and fracturing is believed to have resulted from tensional stresses as the Ocala Uplift developed during post Eocene through Miocene time. As evidenced by the more regular presence of Hawthorn beds in the structurally low area in the Oklawaha River valley, much of the faulting may have taken place in middle or late Miocene time and possibly as late as Pliocene time. Beds ranging in age from at least middle Eocene to apparently Miocene or Pliocene are involved, one place or another in the normal faulting in the Barge Canal area. Probably few faults have a vertical displacement much in excess of 100 feet, and as is typical of normal faults on anticlinal structures, the vertical displacement may be expected to greatest at the top. This is a result of greatest tensional stresses near the surface and, therefore, greatest crustal lengthening near the surface.

The greatest density of faults may be expected where the greatest amount of stretch occurs. The crest of the Ocala Uplift is one of these places, another is in the Oklawaha River valley, which is probably in a structural hinge area somehow related to negative Tertiary movements of the Peninsular Arch, the axis of which is believed to underlie the river valley east and northeast of Ocala and Silver Springs.

Structural dip of the strata in the Barge Canal area, even though very low, is thought to play an important role in the occurrence of Rainbow Springs. The Springs issue from points at or near the contact between the lower member of the Ocala Limestone and the underlying Avon Park Limestone just off the crest of the Ocala Uplift down the northeast flank of the structure (fig. 13). Although faulting may be significant among the reasons for the springs’ occurrence (fig. 17), the writer believes that the permeability of the upper part of the Avon Park is sufficiently low to produce a barrier at the base of the highly permeable Ocala Limestone. Thus, as the contact between the Avon Park and Ocala approaches the surface up the northeast flank of the Ocala Uplift, the large volume of ground water moving southward through the highly permeable Ocala Limestone, is too great to be conducted by the less permeable Avon Park as the overlying Ocala wedges out and the flow is discharged at the surface. On the other hand, faulting is one of the most important geologic factors causing Silver Springs. Apparent downfaulting east of the springs has placed poorly permeable beds of the Hawthorn Formation in position as a barrier to eastward flow in the Floridan Aquifer, thus maintaining a high enough potentiometric surface at the spring site to cause overflow from open limestone caverns and sinkholes. The geohydrologic relationships at the two springs sites are illustrated in part in section X-X’, Figure 10.

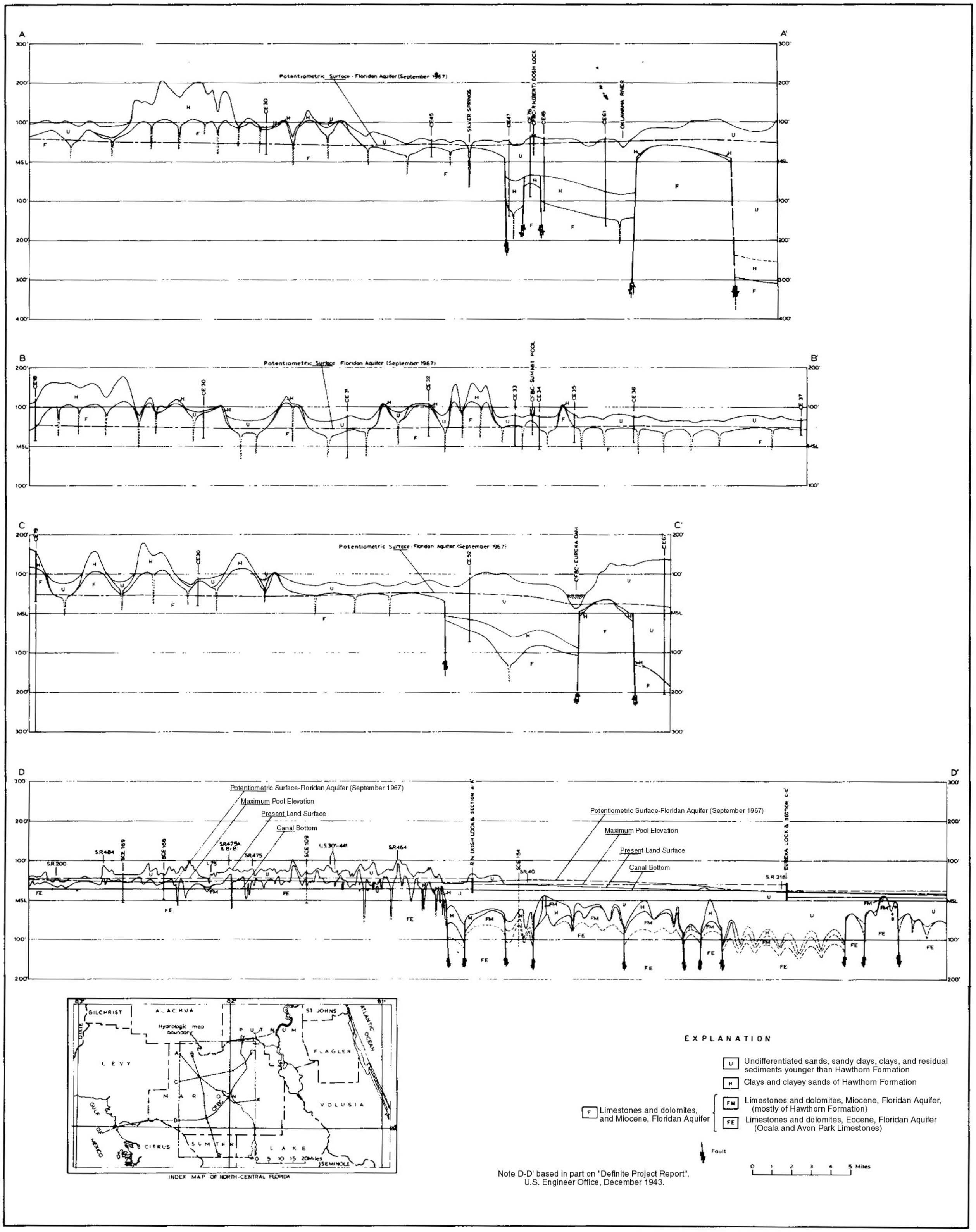


Figure 20. Geohydrologic sections A-A', B-B', C-C', D-D', Ocala vicinity.

Geomorphology

In Pliocene time, the sea receded sufficiently to permit subaerial erosion of the soft Miocene and early Pliocene(?) sediments covering most of north Peninsular Florida. Sea level was probably considerably lower than it is now, and a surface drainage system rapidly developed, its streams easily cutting their way into the soft, clayey sediments. Eventually the Miocene-Pliocene(?) cover began to be stripped off the permeable Tertiary limestones at points down the slopes of the topographically higher areas such as the crest and flank portions of the Ocala Uplift. In areas where the limestone became exposed, the surface drainage system was converted to a subsurface system. As the area of exposed limestone has increased, with the consequent increases in infiltration and limestone solution cavities, one of the best developed ground-water storage and drainage systems in the world has evolved.

As indicated by Figures 9 and 14, only the remnants of a once continuous cover of clayey and sandy strata of Miocene-Pliocene(?) age remains in that part of the canal area west of Ocala and Silver Springs. In this area no integrated surface drainage system now exists. Between the remnants of late Tertiary rocks, most of the limestone is, however, covered by a veneer of unconsolidated sand, clayey sand, and in places, Clay lenses. This cover is partly Miocene-Pliocene(?) residual material after removal of the finer constituents by stream transport or downward leaching by infiltrating ground water. Some is Miocene-Pliocene(?) material preserved as slumpage in old sinkholes, or preserved as originally deposited sediments in depressions on the erosional surface at the top of the Eocene limestone. However, most of the cover is considered to be reworked Miocene and younger clastic material transformed into marine terraces by seas covering the area during Pleistocene interglacial periods.

Generally, areas where the cover consists mostly of residual materials, the sand is relatively clean and unconsolidated within a few feet of the surface, but tends to become increasingly clayey with depth. This is thought to be due to downward leaching of the more soluble material by infiltrating rain water. In areas where the eroded Miocene-Pliocene(?) materials have been transported and redeposited, there often are distinct lithologic breaks, in contrast with the gradational change with depth in the residual materials.

West of the Oklawaha River, beds of Miocene-Pliocene(?) age tend to become more sandy and permeable as the crest of the Ocala Uplift is approached. This situation is thought to be partly responsible for the more continuous preservation of the Alachua Formation in the long, narrow, north trending belt of the Brooksville Ridge near the crest of the Ocala Uplift, as opposed to the more erratic distribution of Hawthorn outliers between the Brooksville Ridge and the Oklawaha River. The permeable sands of the Alachua hindered the development of a surface drainage system that could erode the material of that formation. Rather, rainfall infiltrated directly into the underlying Eocene limestones and then drained from the area through the subsurface. The Alachua outcrop area is expressed topographically in the Brooksville Ridge (fig. 5), where land surface altitudes average over 100 feet, and in places exceed 150 feet. The land surface is a rolling to broadly hummocky, subdued karst terrain with numerous closed sink depressions that have permeable bottoms which do not pond water. However, a few sink depressions with clayey bottoms do contain water. The vegetation growing on the sandy, well drained soil overlying the Alachua Formation tends to be and less lush than that on the Hawthorn outliers, which support large oak and whose clayey phosphatic soils yield good crops and pasture, partly because of their superior soil moisture retention.

The Hawthorn Formation outliers are probably the last vestiges of former surface drainage divides, now expressed as topographic highs--hilly areas such as the Fairfield Hills, Ocala Hill, and Belleview Hill (fig. 5) surrounded by the lower, nearly flat to gently rolling, thinly covered limestone terrain. The Hawthorn outliers are bounded approximately by the 100-foot topographic contour (fig. 14), as this is the average altitude of the very low dipping to nearly flat contact between the Hawthorn and the underlying Ocala Limestone. The condition has been noted by previous workers (Espenshade and Spencer, 1963, p. 24), and the writer's own field observations tend to confirm it. However, the Hawthorn overlies a karst surface of considerable local relief, so irregularities of the contact may range appreciably above or below the 100-foot level.

Altitude of the hilly terrain of the Hawthorn outliers is as much as 215 feet above sea level in the northwestern part of Marion County, but generally is around 150 feet or less. Numerous large depressions, some whose lowest points are at altitudes below 100 feet, have developed as a result of the collapse of limestone caverns at depth, thus producing a hilly, subdued karst topography. Many depressions, large and small, have clay bottoms and thus pond runoff. Drainage of the outliers is for the most part into the depressions. Some sinks are connected directly with the limestone aquifer, and many partially plugged depressions allow seepage into the aquifer. Drainage from the edges of the outliers infiltrates rapidly into the surrounding near surface limestones.

Limestone caverns oriented along either a northwest or a northeast trending fracture (fig. 18) are frequently found well above the water table and open to the surface at or near the 100-foot level in the Hawthorn outcrop areas. The caverns were formed below the water table in the limestone at a time when the water table stood at a higher altitude than at present. However, some vertical pipes were formed above the water table as a result of solution as recharge flowed to the aquifer through limestone fractures in the unsaturated zone.

Surrounding the highlands of remnant Miocene strata are the comparatively low, broad, nearly flat to gently rolling, thinly covered Eocene limestone terrains--the Central and Western Valleys and the Sumter Upland physiographic subdivisions (fig. 5) in the Barge Canal area. They average about 75 feet above sea level, although bottoms of characteristic shallow sink depressions sometimes are as low as 50 feet above sea level. Generally the limestone is overlain by 25 to 50 feet of sand and clayey sand which allows direct recharge to the aquifer, and seldom permits ponding of surface runoff in depressions whose bottoms are above the water table.

Figure 15, a map of the thickness of material overlying Floridan aquifer in the Ocala vicinity, shows that the cover is thinnest, usually less than 25 feet thick, at the edge of the Miocene outliers. Figure 19, shows that the altitude of the surface of the Ocala Limestone diminishes rapidly away from the Hawthorn Formation outliers. From the 100-foot contour, where the top of the Ocala is at land surface, the rock surface slopes rapidly downward to altitudes averaging about 50 feet above sea level. The Ocala covered by 25 feet or more of sand and clayey sand (fig. 15) within a short distance from the 100-foot contour. Comparatively rapid solution has caused a lowering of the limestone surface where it is not protected by the clayey cover of the Hawthorn Formation.

Sellards (1914, p. 121-124) estimated, through determinations of the amount of dissolved solids carried by ground water discharging from some springs in central Florida, including Silver Springs, that the limestone surface of the central peninsular section of Florida is being lowered at the rate of one foot in 5,000 or 6,000 years.

In Pliocene time, eroding streams removed enough poorly permeable material so that surface runoff began to infiltrate with increasing ease into the underlying, fractured, permeable limestone. After reaching the zone of saturation, the water then moved preferentially along the fractures in the direction of slope of the water table, producing solution channel systems oriented, of course, along the fractures and in a general way parallel to the old drainage valleys, as the valleys themselves tended to be oriented with the fracture systems. The old drainage valleys are reflected on the top-of-rock map (fig. 19) and are approximately coincident with the lowest parts of the present rock surface. The highest points of the present rock surface underlie existing outliers of the Hawthorn Formation west of the Oklawaha River valley, and the lowest points are intermediate between the Hawthorn outliers. It is apparent that the areas where the limestone surface is lowest, as shown on Figure 19, not only represent the alinements of old surface drainage valleys, but the lows also are the locations of the best developed solution channel systems where ground-water flow is concentrated today. The potentiometric surface of the Floridan Aquifer, as will be discussed later, substantiates this premise.

In the canal area, zones of concentrated ground-water flow such as just described terminate at Silver and Rainbow Springs, where tremendous volumes of ground water are discharged to what remains of the surface drainage system. The heads of the springs have migrated to their present positions partly because of a tendency of ground-water levels to decline as permeability in the aquifer is increased due to removal of limestone by solution, and because of mechanical erosion of the limestone in the vicinity of the spring heads. Also, points of principal spring discharge have shifted in the past due to changes in ground-water levels in response to changes in sea level. The subsurface drainage system is continuing to evolve today, as evidenced in part by frequent occurrence of new sinkholes and by the presence of significant amounts of calcium bicarbonate in the spring waters.

The frequency of reports of new sinkholes indicates that some of the most active sinkhole development in the study area today in areas underlain by zones of concentrated ground-water flow such as described above. The holes result from collapse and flow of the sandy and clayey cover into limestone solution cavities. If a sinkhole suddenly develops near a point of groundwater discharge, such as a well or spring, temporary turbidity in the discharging water may result. However, numerous sink occurrences apparently cause no noticeable turbidity in ground-water discharging at either near or far locations. It is possible also that a major sinkhole development near a spring could cause blocking of important flow channels and cause a shift in location of principal spring discharge, or otherwise alter the flow regimen of the spring.

East of Silver Springs, where an essentially continuous cover of Miocene-Pliocene(?) beds remains over the permeable limestones and dolomites of the Floridan aquifer (figs. 13 and 14), poorly permeable parts of this cover place the aquifer under confined or artesian conditions in much of the area. This is the case in the Oklawaha River valley, which is underlain by the Hawthorn Formation. However, in places east of the river, there is considerable evidence to indicate that the aquifer is unconfined and direct recharge occurs through permeable sand of the Fort Preston formation. In the Oklawaha River valley a surface drainage system has been preserved over the poorly permeable parts of the Miocene-Pliocene(?) strata. Silver River carries the discharge of Silver Springs eastward a few miles over these strata to empty into the northward flowing Oklawaha River, which continues to flow over these same strata to the St. Johns River with limited direct gain from the Floridan Aquifer.

Eaton and Orange Creeks are the only significant tributaries to the Oklawaha River in the reach downstream from Silver River before the Oklawaha turns sharply eastward near Orange Springs. Eaton Creek is a short sluggish stream which empties into the Oklawaha about three miles above Eureka Lock site. It drains a small chain of lakes including Eaton and Mud Lakes, all of which are located on the confining beds. Orange Creek heads at Orange Lake and flows eastward over the confining beds to discharge into the Oklawaha near Orange Springs. Wells drilled into the Floridan Aquifer near these streams will flow.

West of Ocala and Silver Springs, no surface drainage system exists in the Barge Canal area except for the Withlacoochee River and a tributary, Rainbow River, which empties into the Withlacoochee near the town of Dunnellon. The Withlacoochee River does not flow over discrete, poorly permeable strata, as does the Oklawaha, but seems to owe its existence to a high-water table maintained by a combination of normal faulting and permeability changes. The permeability changes are thought to be due both to fortuitous fault emplacement of poorly permeable dolomite of the Avon Park Limestone and to plugging of solution cavities by sand fill.

Test drilling along the centerline of the Barge Canal indicates that the valleys of both the Oklawaha and the Withlacoochee Rivers have been cut and filled, possibly several times, to considerable depths (more than 100 feet) below their present stream bottoms (Vernon, 1951, p. 31, fig. 14). Such down-cutting probably took place mostly at low stands of the sea during glacial periods, whereas the filling occurred as stream gradients were reduced when sea level rose again as the glacial ice melted. White (1958, p. 17) believed, partly because of their coast-parallel trends, that the valleys of both rivers originated as lagoons behind offshore sandbars, probably sometime during Pleistocene time, the bars now being represented by Mount Dora Ridge east of the Oklawaha River, and the Brooksville Ridge west of the Withlacoochee River (fig. 4). It is generally believed that the material making up the “bars” is older than Pleistocene, that the Mount Dora Ridge consists largely of clayey sands of the “Fort Preston formation,” and that sands of the Alachua Formation make up a large part of the Brooksville Ridge Coastal influences at the time of origin of the sediments in Miocene-Pliocene(?) time bad much to do with their present distribution. However, if Pleistocene seas rose as high as some believe, they most certainly would have had an influence on the present shape and size of the ridges, and to some extent the seas would have reworked at least the surficial sands of the ridges. In any case, the ridges are an important valley control, and neither stream could turn to the sea until a large gap was reached in the ridge so that the stream could pass through.

The tops of the ridges are hummocky and quite irregular in altitude, probably as a result of differential subsidence due to solution of underlying limestone (White, 1958). There are no tributaries from the ridges to the streams, as the sands allow rapid infiltration of rainfall as recharge to both the shallow and the Floridan Aquifers. Because of the insoluble character and greater resistance to physical erosion of the materials making up the ridges, they have remained above the surrounding terrain.

HYDROLOGY

General

The Cross-Florida Barge Canal, when completed, will be integral part of the hydrologic system of the area and is designed to conform to the natural hydrologic regime. Waters of the two principal streams and two major springs of the area, that is, the Oklawaha River and Silver Springs on the east and the Withlacoochee River and Rainbow Springs on the west, will be impounded by three dams for use in maintaining navigation depths in the lower reaches of the canal. In addition the dams and reservoirs will provide flow control for the area. The stage of the Summit Pool will vary with the natural fluctuations of the potentiometric surface of Floridan Aquifer. Five locks will maintain water levels in the several pools of the canal and produce a stairstep effect (fig. 2). Gated spillways in each of the three dams will control fluctuation of water levels in the reservoirs of the lower reaches of the canal, including Rodman and Eureka Pools in the Oklawaha River valley and Inglis Pool in the Withlacoochee River valley. A pumping station just below R. N. Dosh Lock and possibly another below Dunnellon Lock, will be used to pump water from Eureka Pool and (or) Inglis pool into the Summit Pool to replace losses due mostly to lockages. Thus it is possible for no net loss of ground water to occur from the aquifer in the area of the Summit Pool as result of canal operations.

Most of the length of the Summit Pool will be excavated into limestone or dolomite of the Floridan Aquifer; and aside from pumpage from lower pools, the source of water for this reach of the canal will be direct ground-water inflow from the aquifer. Surface runoff will be a negligible source of supply to the Summit Pool.

Surface streams will supply water to the three lower pools, although a major source for the streams is the discharge from large springs issuing from the Floridan Aquifer. During low-flow periods, streamflow consists almost entirely of ground-water inflow and spring discharge. During wet periods and floods direct surface runoff, partly from the overflow of lakes and swamp areas in the river valleys, may be the major contributor to streamflow.

Surface Water

Occurrence and Movement

In addition to that in the principal streams, the Withlacoochee and Oklawaha Rivers, surface water occurs in the Barge Canal area as a few fairly large lakes and numerous small lakes, ponds, and swampy areas, most of which are within the principal river valleys. Ponds in sinkhole depressions in the areas of Hawthorn Formation outcrop are also common, but the large sand covered limestone areas between have few ponds. Swampy prairies are in some low areas outside the river valleys where the water table fluctuates at or near the land surface.

The larger lakes of the area include: Orange Lake in southern Alachua County northwest of the canal, Lake Kerr in northeast Marion County about 5 miles east of the Eureka Lock and Dam site, Lake Weir about 10 miles southeast of the canal in southeastern Marion County, and Lake Tsala Apopka. This last is really an elongate group of interconnected, flooded sink depressions and potholes complexly involved with the channel of the Withlacoochee River some 6 to 20 miles southeast of Dunnellon Lock.

Stages in each of the four large lakes and in several other lakes farther from the canal are observed as a part of the Barge Canal ground-water monitoring program. Among those lakes being observed, differing hydrologic relationships exist between the lakes and the ground-water regime; and, although canal operations are not expected to affect the lake levels, collection of background data from the lakes is an important part of the ground-water monitoring program.

Since there is limited direct relationship between the canal and natural lakes and ponds in the area, no effort has been made to measure total surface area covered by ponded water. Many of the lakes are perched on materials of low permeability overlying the Floridan Aquifer, but there probably is some leakage to the aquifer from most of them. Some lakes, such as Orange and Tsala Apopka, have direct connections with the Floridan Aquifer, and not only do they receive water from the aquifer, but at times and in places they lose water directly to the aquifer. Evaporation losses in the area from open water surfaces are estimated to be about 51 inches per year based on studies made in the Orlando area, to the south (Anderson, Lichtler, and Joyner, 1965, p. 2).

No detailed examination and evaluation of stream discharge in the area has been made as a part of this investigation. However, general discharge information, based on published U. S. Geological Survey records (1963 and 1965), is discussed here in order that a clear and balanced account of the hydrologic system be presented. The 35-year record at the gaging station near Ocala (Sharpes Ferry) indicates the Oklawaha River will discharge into the upper end of the proposed Eureka Pool on the average, 420 cfs (cubic feet per second). The maximum discharge of record is 2,270 cfs, and the minimum is 7.2 cfs. During the period of record, flow was controlled by the old lock and dam at Moss Bluff, upstream from Sharpes Ferry. A new lock and dam completed in 1969 at the Moss Bluff site will control the flow of the Oklawaha River to Eureka Pool.

Added to the flow of the Oklawaha near the upper end of the Eureka Pool will be the flow of 5-mile-long Silver River, which heads in Silver Springs. The 36-year period of record (ending September 30, 1968) for Silver Springs indicates an average discharge of 822 cfs (531 mgd). Maximum discharge of record is 1,290 cfs and the minimum is 539 cfs. Silver Springs will be the most reliable and constant supply to the Eureka Pool.

Water for Rodman Pool will be supplied by flow through the Eureka Dam to the Eureka Dam spillway and by Orange Creek with a 20-year mean discharge of 188 cfs, a maximum record flow of 2,170 cfs, and a minimum record flow of 2.0 cfs. Additional small tributaries, the most important being Deep Creek, drain into Rodman Pool below Orange Creek.

The Withlacoochee River flows from the southeast into the canal at Dunnellon, and will be a principal source of water for the Inglis Pool. Mean discharge at the gage near Holder, about 8 miles south of Dunnellon, was 1,170 cfs for the 34-year period of record (1931-1965) with a maximum discharge of 8,660 cfs and a minimum of 112 cfs. At nearly the same point where the Withlacoochee joins the canal, the 5-mile-long Rainbow River, carrying the discharge of Rainbow Springs, enters the Inglis Pool from the northeast. Long-term, but intermittent, records (1898 to September 30, 1968) indicate the mean discharge for Rainbow Springs to be 724 cfs (468 mgd). The maximum discharge of record is 1,230 cfs and the minimum is 487 cfs.

Quality of Surface Water

The surface water of the Barge Canal area is commonly highly colored but contains less dissolved solids than the ground-water of the area. The concentration of dissolved solids in surface water is comparatively low because surface water is not in as close contact with soluble materials as is ground water, and because the rapid movement of surface water decreases the time for solution to take place. Also, the materials in contact with surface water in the area of investigation usually are less soluble than those forming the aquifers. During low flow when the water in the streams is derived almost entirely from ground-water inflow or spring discharge, its dissolved-solids content is appreciably higher than in wet periods when the water is derived mostly from direct surface runoff.

As streamflow diminishes, one of the more noticeable changes in the quality of the surface water in the area of investigation, except near major spring discharge, is the increase in total hardness as an increasing percentage of the flow is derived from ground-water from the Floridan Aquifer. Water from the Floridan Aquifer commonly has a high calcium bicarbonate content, and thus high total hardness.

The brown color of the surface waters of the Barge Canal area, and of those of Florida in general; results from the solution by surface runoff of decaying organic materials, especially leaves and wood debris, found in abundance on the land surface in Florida. When water enters the ground and percolates through soil and rock an appreciable distance, any bacteria or color that was present will usually be removed. This is why ground water generally has little or no color. During periods of low streamflow color is generally reduced.

The chloride content in both the Withlacoochee and Oklawaha Rivers is generally only a few milligrams per liter except near their mouths. In the Withlacoochee it usually increases rapidly from the vicinity of Yankeetown to the Gulf, the point of change being variable depending upon the magnitude of the tides and river discharge. A spot survey of chloride content was made in May 1968 in the Oklawaha River during the annual low-flow period. Between the mouth of Silver River and the mouth of the Oklawaha, chloride increased from about 10 mg/L (milligrams per liter) to more than 100 mg/L. Most of the overall gain occurred as abrupt increases at a few points along the reach. The chloride content of the St. Johns River opposite the mouth of the Oklawaha was over 350 mg/L at the same time, due at least in part to inflow of high chloride spring water into the St. Johns River upstream from the mouth of the Oklawaha River. Most other common constituents are present in only small amounts in the surface waters in the Barge Canal area.

During April-September 1967, the Federal Water Pollution Control Administration (1967, p. 3) made a study of the mineralogical and bacteriological quality of the waters of the Oklawaha and Withlacoochee Rivers in the area of the Barge Canal. The study indicated that, in general, the quality of surface water in the proposed impoundment areas along the canal is good, and is acceptable for planned uses if maintained at the pre-impoundment quality levels.

Ground Water

General

The ground-water system of the Cross-Florida Barge Canal area includes two fresh-water aquifers. By far the most important of the two as a source of water supply is the Floridan Aquifer (Hyde, 1965), which is composed almost entirely of limestone and dolomite ranging in age from middle Eocene to middle Miocene. This aquifer varies considerably in porosity and permeability both vertically and horizontally, but for practical reasons, it is usually treated as an essentially continuous hydraulic unit present in the subsurface throughout Florida and parts of adjacent states.

The Floridan Aquifer is overlain in parts of the area by a shallow aquifer (Hyde, 1965), which is more heterogeneous and less continuous than the Floridan Aquifer. This shallow aquifer consists of permeable sand and shell beds, often of limited horizontal and vertical extent, occurring within the stratigraphic section that overlies rocks of the Floridan Aquifer. Materials composing the shallow aquifer range in age from lower Miocene to Holocene.

The Shallow Aquifer

Distribution and Extent.--Geologic maps and a geologic section (figs. 9, 10, 13, and 14) show the distribution of Miocene and younger beds of the shallow aquifer. Most of the area mapped as near-surface limestone on the geologic maps is actually covered by sedimentary materials several tens of feet thick. When sand and clay lenses or beds are so associated with each other that the porous sand unit may collect and retain infiltrating water for a reasonable length of time as a perched water body, the shallow sedimentary materials constitute a shallow aquifer. Also, if no impermeable layer separates permeable sand from the underlying limestone of the Floridan Aquifer, but the water table extends up into the sand an appreciable distance, the sand may be arbitrarily treated as the shallow aquifer simply on the basis of differences in lithology.

The shallow aquifer in the study area is generally thought of as water table aquifer even though it is confined locally. It is not uncommon to find a sand bed confined above and below by clay which can maintain the water in the sand under artesian pressure.

Figure 15, a map showing the thickness of material overlying the Floridan Aquifer, and geohydrologic sections in Figure 20 aid in determining locations where the shallow aquifer might be utilized as a source of water.

Most of the better yielding shallow aquifer water-level monitoring wells in the Barge Canal area are east of Ocala and Silver Springs, where continuous Miocene-Pliocene(?) sedimentary cover is mapped (figs. 9, 13, and 14). Altitudes of water levels in the Floridan Aquifer and the shallow aquifer at most places are within 3 feet or less of each other; but in other places, as near Fore Lake in the Ocala National Forest, the water level in the shallow aquifer may consistently fluctuate within a range from 20 to 30 feet higher than that of the Floridan Aquifer. During periods of heavy rainfall and rapid recharge, water levels of the shallow aquifer wells often fluctuate more rapidly and over wider ranges than do those of the Floridan Aquifer.

The shallow aquifer is most frequently used in the general area of the Ocala National Forest, east of the Oklawaha River. Sand deposits as much as 300 feet thick underlie the area and result in comparatively high drilling costs for Floridan Aquifer limestone wells. Also, beneath the northern part of the Ocala National Forest near the Oklawaha River, the Floridan Aquifer contains salty water. Most wells in the shallow aquifer are for domestic use where only small supplies are needed. However, properly constructed wells in some areas may yield large quantities of water.

In hilly areas underlain by the Hawthorn Formation, the shallow aquifer absorbs large amounts of water from precipitation, and during wet periods helps to maintain water levels in small ponds and in short intermittent streams flowing to sinkholes. Also, during wet periods the shallow aquifer helps stabilize the flow of larger streams, such as the Oklawaha River, by temporarily retaining surface runoff.

Directly related to the storage and leakage characteristics of the shallow aquifer are the possible leakage conditions that could exist in the vicinity of the downstream part of the Eureka Pool where the canal water level will be 15 to 20 feet above present natural water levels of both the shallow aquifer and the Floridan Aquifer. The significance of this relationship is discussed later in the report.

Quality of water in the aquifer.--In general, the quality of water of the shallow aquifer is good--usually of considerably lower dissolved solids content and softer than the waters in the Floridan Aquifer. One of the most common problems of utilization of the shallow aquifer water is a high iron content in water from very shallow wells and from wells adjacent to ponds and lakes. This problem can sometimes be alleviated by drilling deeper or by relocating the well away the pond or depression. Also, another common problem is the presence of clay in suspension in the water. The sands of the shallow aquifer at many places contain some clay, but clay-free sand beds are present in the aquifer. Application of modern well completion practices can aid considerably in efficient production of water from sands of the shallow aquifer.

The Floridan Aquifer

Character and distribution.--The name “Floridan Aquifer" is commonly applied in Florida to the principal artesian aquifer of the southeastern United States. The aquifer consists most of limestones and dolomites, middle Eocene to middle Miocene in age, which act more or less as a hydrologic unit in most of Florida, in southeastern Georgia, and in parts of Alabama and South Carolina. The aquifer is, however, of variable porosity and permeability and consists in many places of well developed cavernous intervals separated by zones of low permeability which act as confining layers. Thus, the Floridan Aquifer may in places be thought of as a compound aquifer consisting of several subaquifers. It is one of the most expensive limestone aquifers in the United States (Stringfield, 1966, p. 95).

Parker and others (1955, p. 189) who first applied the name “Floridan,” defined the Floridan Aquifer in Florida as being limited to the following sequence: Lake City and Avon Park limestones of middle Eocene age, Ocala Limestone of late Eocene age, Suwannee Limestone of Oligocene age, Tampa Limestone of Miocene age, and permeable parts of the Hawthorn Formation of Miocene age that are in hydraulic contact with the rest of the aquifer. Although the aquifer is usually referred to as the principal artesian aquifer where it occurs in areas other than in Florida, the term “Floridan Aquifer” and the above described stratigraphic boundaries are now generally recognized in Florida, and in this report.

Down both the north and south plunge of the Ocala Uplift the aquifer attains a thickness of as much as 1,500 feet (Stringfield, 1966, p. 97), but in most of the Barge Canal area it ranges from about 1,000 feet to 1,200 feet thick (Chen, 1965). The Suwannee and Tampa limestones are missing in the Barge Canal area, so the Floridan Aquifer in this area includes only the Lake City, Avon Park and Ocala limestones, and parts of the Hawthorn Formation. As a source of ground water for the canal, the writer considers the Ocala Limestone the most important stratigraphic unit of the aquifer.

In much of Florida the aquifer is confined by overlying poorly permeable sediments of Miocene age and younger. These sediments are thickest in structurally low areas and thin toward structurally high areas. Only scattered remnants of the once continuous Miocene cover remain on the flanks of the Ocala Uplift. Thus, in the area of the Ocala Uplift much of the Floridan Aquifer is unconfined and, therefore, under water-table conditions. As discussed earlier, the aquifer is confined along the route of the Barge Canal northeast of Silver Springs by the thick sequence of poorly permeable deposits underlying the Oklawaha River valley. Near the river artesian flow occurs from wells that penetrate the Floridan Aquifer. However, southwest of Silver Springs, along the canal route to the Gulf, these relatively impermeable materials are essentially missing, and the aquifer is under water-table conditions. These hydrogeologic factors are of fundamental importance to the design and operation of the Cross-Florida Barge Canal.

Water in the aquifer in the Barge Canal area is derived from direct recharge by local rainfall and from ground-water inflow from potentiometrically high areas immediately to the north and south. Most recharge takes place where the aquifer is unconfined, although some recharge occurs where the poorly permeable overburden is thin or has been breached by sinkholes. Major discharge from the aquifer occurs from Silver and Rainbow Springs, two of the largest fresh-water springs in the United States. A comparison of hydrographs of rainfall, ground-water level, and spring discharge in the canal area demonstrates the close relationships among these three hydrologic parameters (fig. 21).

<u>Potentiometric surface of the aquifer</u>.--The potentiometric surface of the Floridan Aquifer is that imaginary surface connecting all points of equal altitude to which water will rise in tightly cased wells open to the aquifer, whether or not the aquifer is confined or unconfined. Where the aquifer is unconfined, the potentiometric surface is the level at which water will stand in uncased or screened wells.

Figure 22 is a map of the potentiometric surface of the upper part of the Floridan Aquifer in north-central Florida centered in the Barge Canal area. The contours are drawn at a 5-foot interval and are based on ground-water levels measured in May 1968, mostly in the upper 50 to 200 feet of the aquifer. Two major potentiometric highs appear on the map, one in Polk County at the south end of the map area, and the other in Clay and Bradford Counties at the north end of the area. These highs are separated by a saddle reflecting the large discharges at Rainbow and Silver Springs. The canal route passes through the area of this potentiometric saddle and within a few miles of both springs.

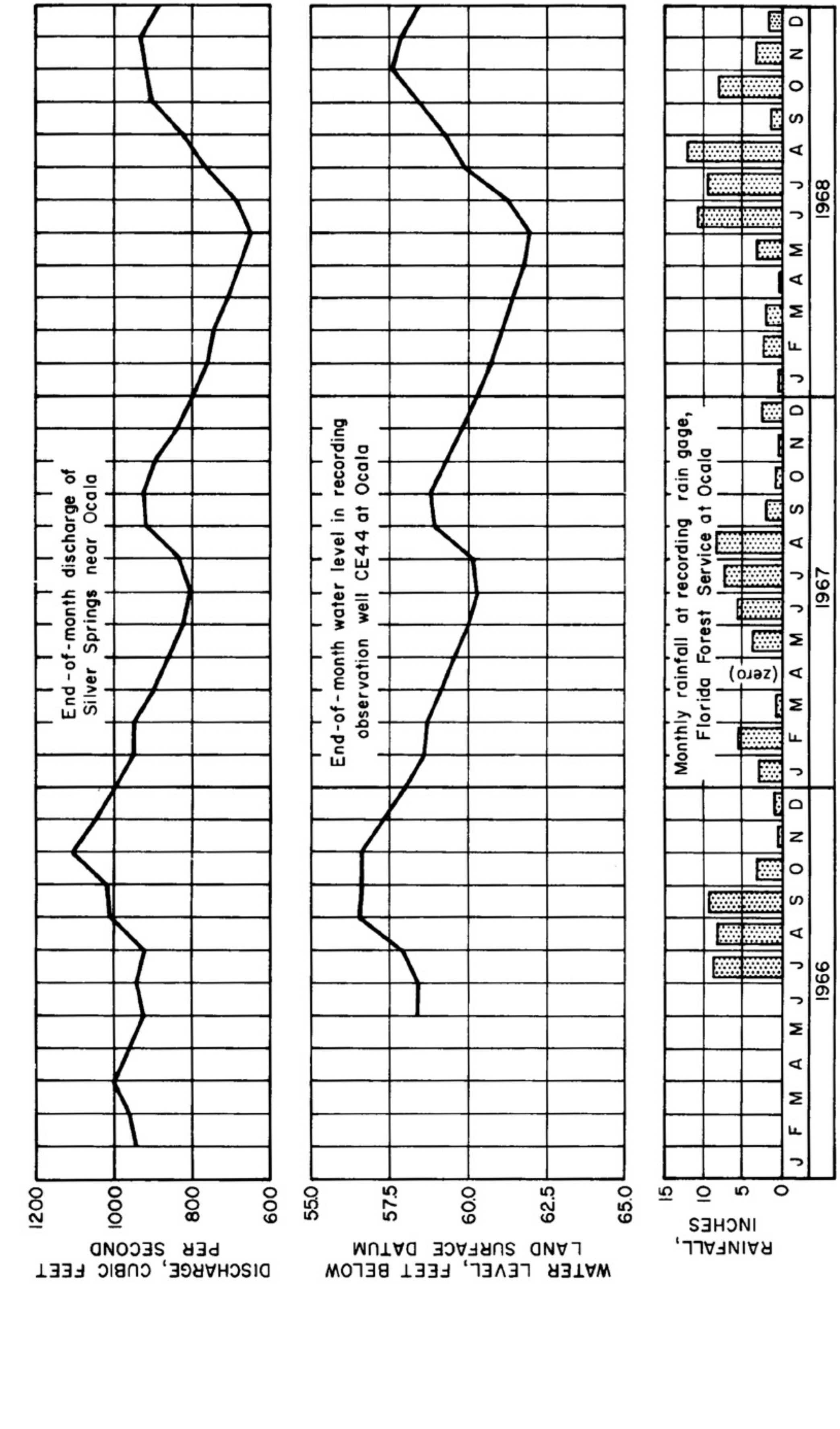


Figure 21. Hydrographs showing interrelationship of rainfall, water level in Floridan Aquifer, and discharge of Silver Springs.

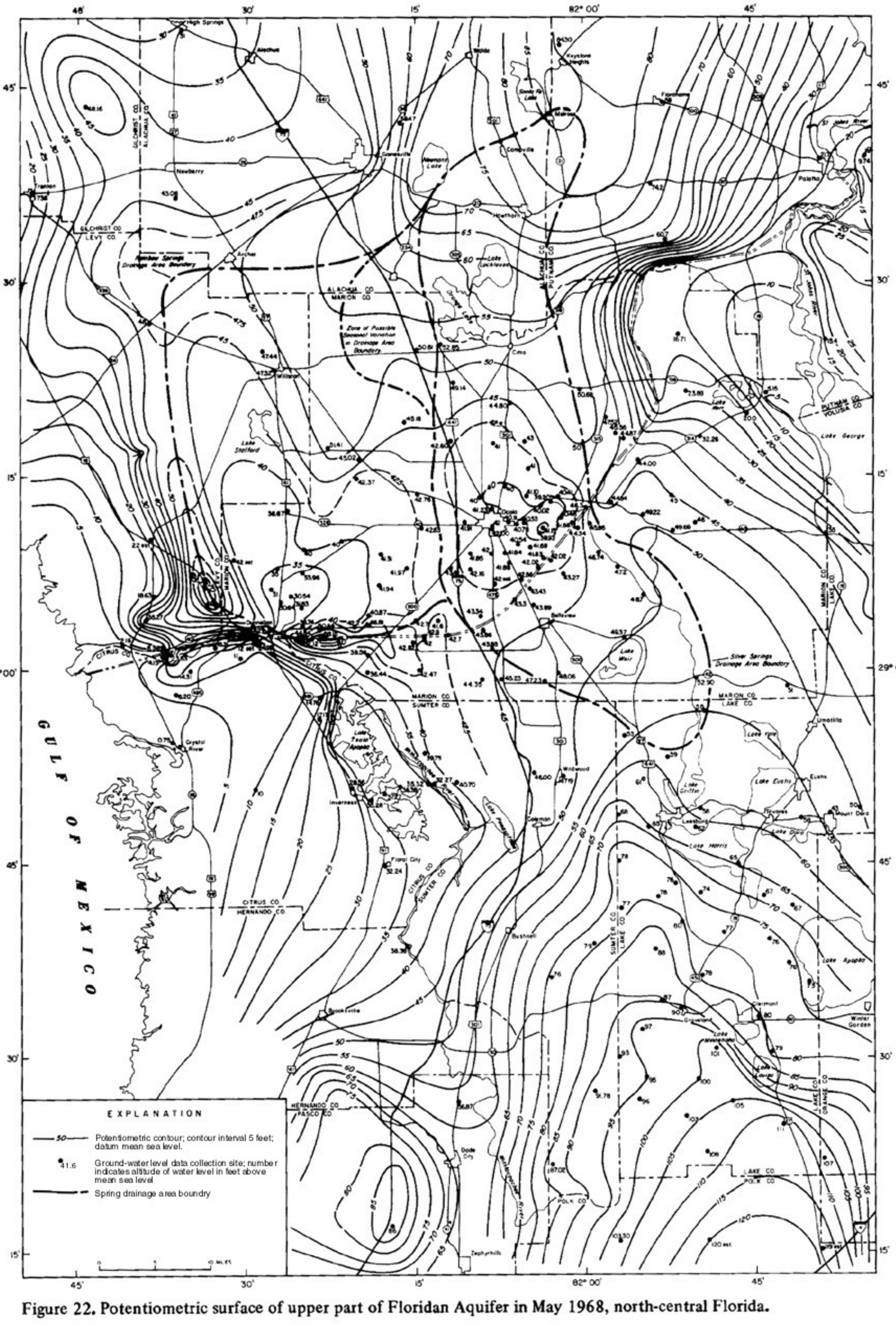


Figure 22. Potentiometric surface of upper part of Floridan Aquifer in May 1968, north-central Florida.

Ground water moves downgradient from the potentiometric highs toward the saddle and the peninsular coasts along flow paths approximately perpendicular to the equipotential lines. Ground-water drainage basins in the upper part of the aquifer may be outlined on the map by drawing lines along potentiometric divides. Thus, it is possible to demonstrate the spatial relationship of the canal route to the ground-water flow pattern and to delineate the drainage areas of Rainbow and Silver Springs. The mapped potentiometric surface is assumed to represent that part of the aquifer supplying the springs. A knowledge of the size and shape of the drainage areas of the two springs is essential to the quantitative analysis of aquifer characteristics along critical reaches of the canal.

In addition to the regional potentiometric surface map just described, maps of the Barge Canal area for May 1968 (annual low-water period) and September 1968 (annual high-water period) are presented as Figures 23 and 24. The purpose of the high- and low-water period maps is to show changes in the shape of the surface that occur from one seasonal extreme to the other. Larger scale detailed maps of the Silver Springs Ocala vicinity with a contour interval of 1 foot, for May and September 1968 (figs. 25 and 26), have been used for flow net analysis to determine hydrologic relationships between the Summit Pool and Silver Springs.

The general configuration of the potentiometric surface changes little from one season to the other, even though the surface is several feet higher and gradients tend to be somewhat steeper in the high-water period. The most interesting change is in an area northwest of Ocala where the ground-water drainage divide between Silver and Rainbow Springs appears to shift several miles westward during the change from the low- to the high-water period.

In addition to showing general directions and preferential routes of ground-water flow, the maps of the potentiometric surface help to demonstrate important geohydrologic relationships in the area. For instance, the configuration of the potentiometric surface tends to confirm routes of concentrated ground-water flow suggested by the fracture trace maps, thickness of cover map, and top-of-rock map (figs. 15, 18, and 19). Conversely, the geologic maps are useful interpretive tools in estimating the potentiometric surface in areas of sparse water-level control, and for interpreting the anomalies on the potentiometric surface. The foregoing relationships are discussed later in this report.

<u>Characteristics of recharge to the aquifer</u>.--Local rainfall recharges the Floridan Aquifer in the canal area. Most direct and heaviest recharge is on the flanks of the Ocala Uplift where the Ocala Limestone of Eocene age is at or near land surface (figs. 9, 10, and 14). Here recharge is by direct infiltration through the sand and clayey sand which covers much of the limestone and which is as much as several tens of feet thick (fig. 15). Where scattered remnants of Miocene-Pliocene(?) sediments of low permeability still cover the limestone on the flanks of the uplift, such as in the Fairfield and Ocala Hills, important recharge is concentrated at sinkholes that penetrate the cover. Also, there is some minor recharge to the aquifer through the poorly permeable material separating the sinks.

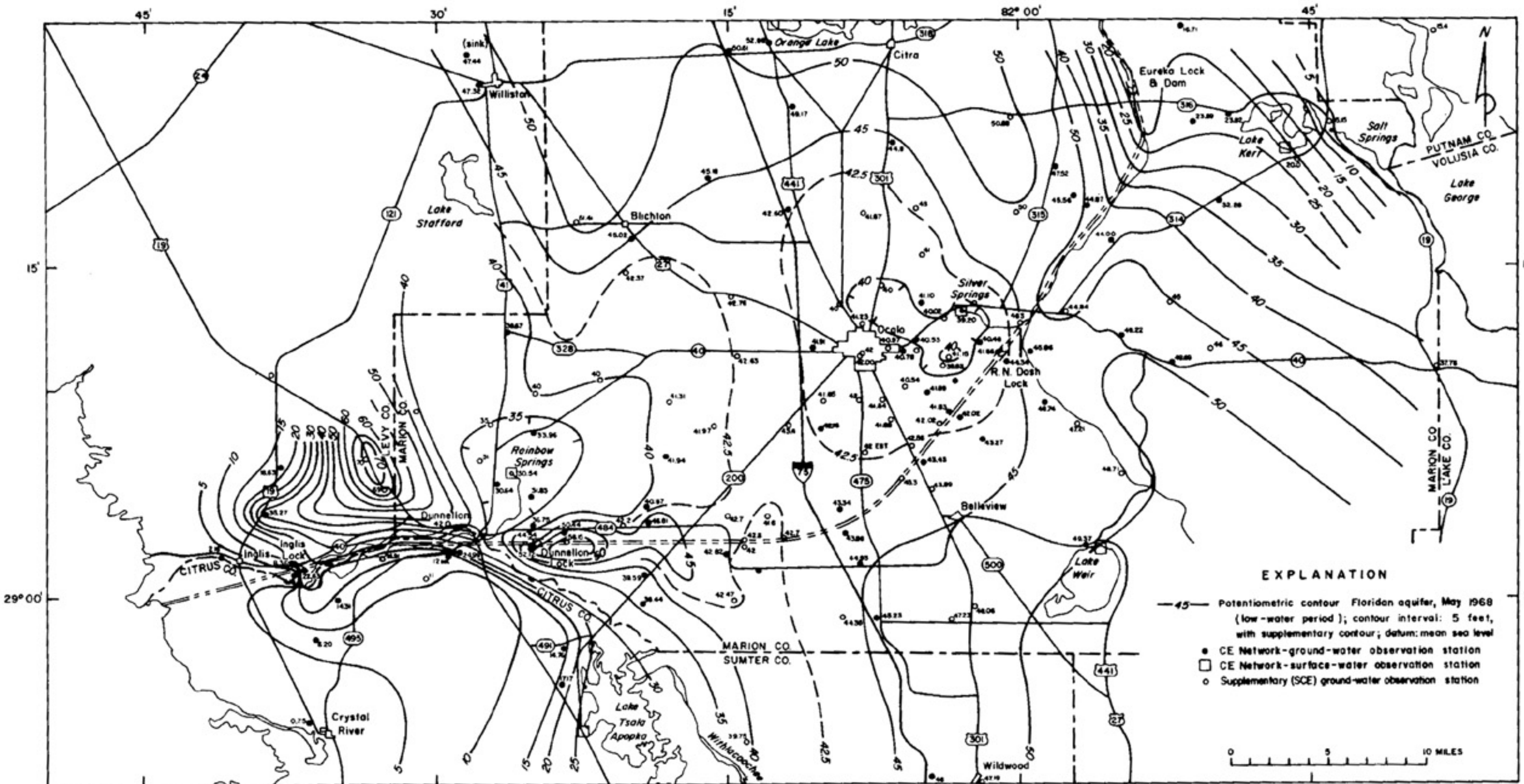


Figure 23. Potentiometric surface of upper part of Floridan Aquifer in May 1968 (low-water period), Cross-Florida Barge Canal area.

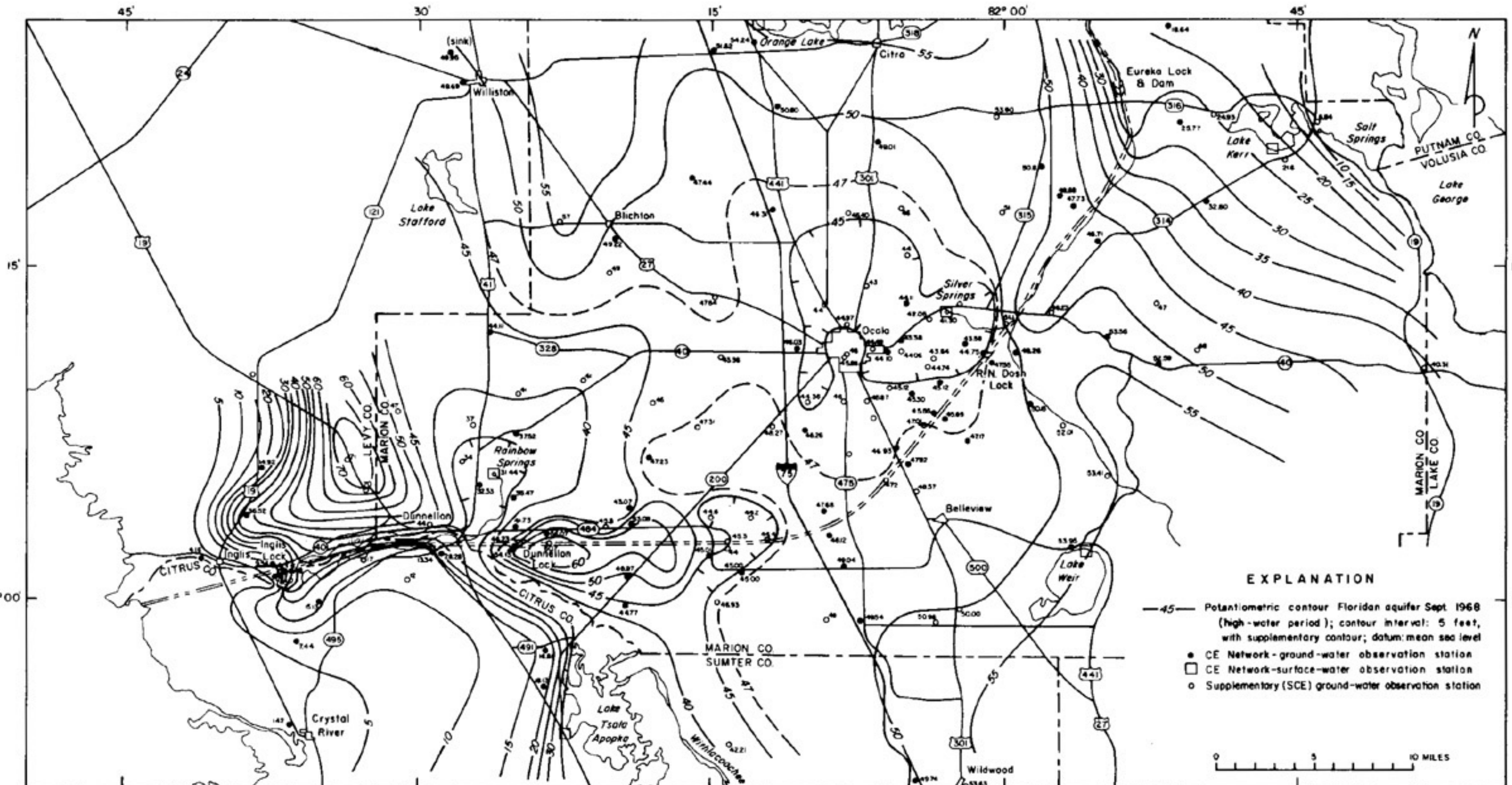


Figure 24. Potentiometric surface of upper part of Floridan Aquifer in September 1968 (high-water-period), Cross-Florida Barge Canal area.

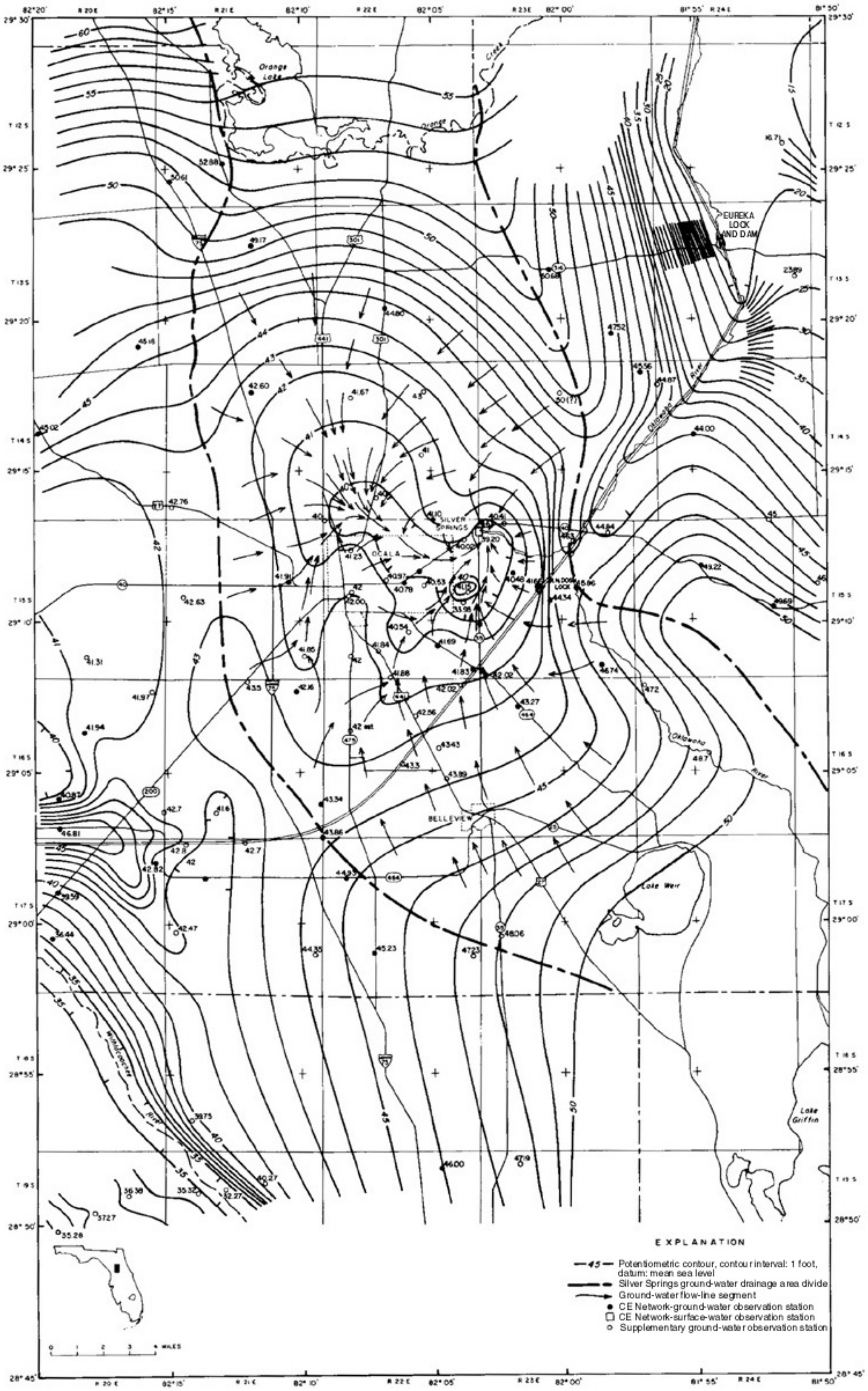


Figure 25. Potentiometric surface of upper part of Floridan Aquifer in May 1968 (low-water period), Ocala vicinity.

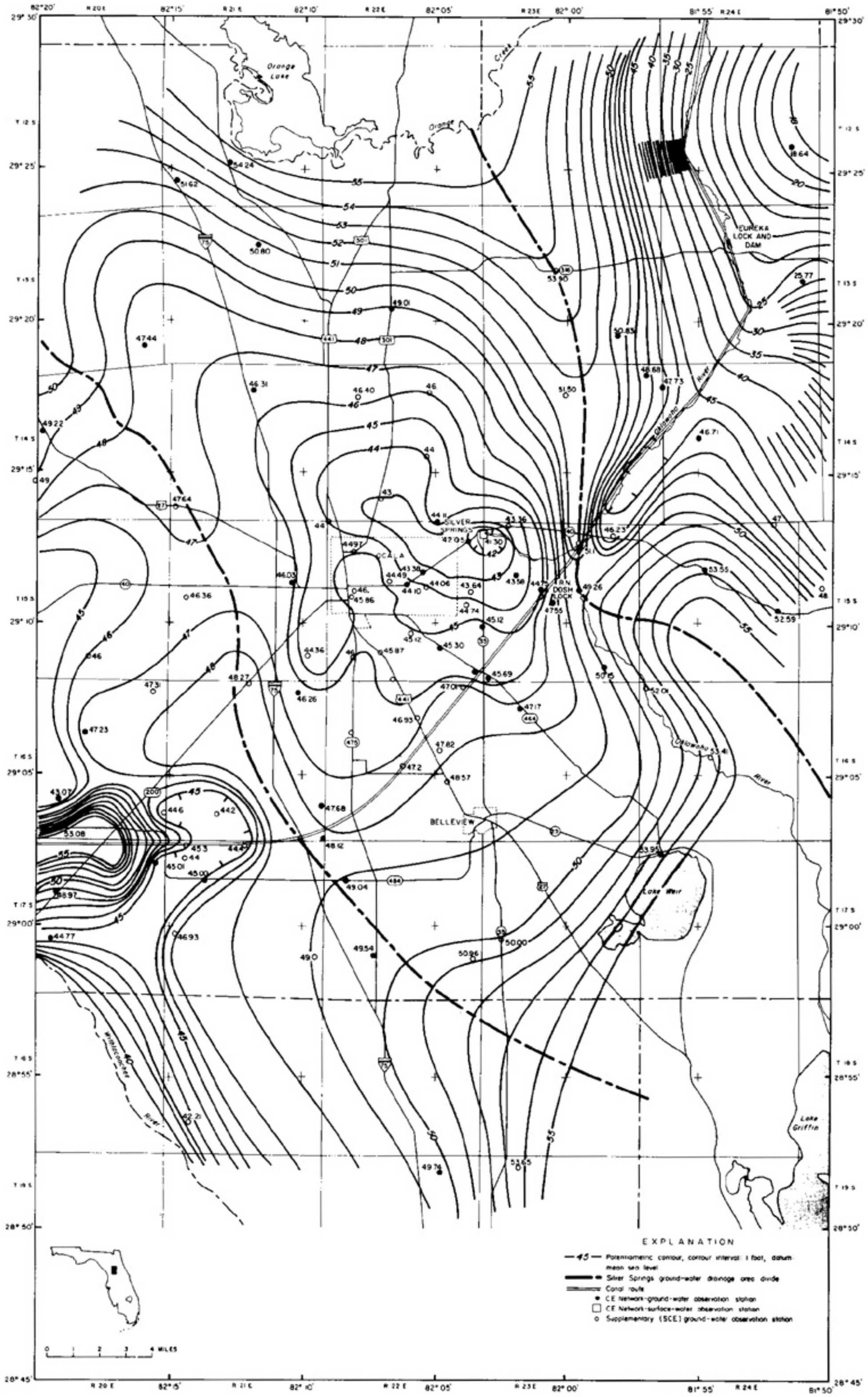


Figure 26. Potentiometric surface of upper part of Floridan Aquifer in September 1968 (high-water period), Ocala vicinity.

There is direct recharge to the aquifer in those areas west and south of Rainbow Springs where Avon Park Limestone is at or near the surface on the crest of the Ocala Uplift (fig. 13). However, the comparatively low permeability of the upper part of the Avon Park, a consequence of dolomitization and presence of sand and clay fill in solution cavities, has caused local potentiometric highs, water levels near the land surface, and some rejection of recharge during very wet periods.

East of Silver Springs, where the confining layer underlies the floor of the graben valley of the Oklawaha River, the flood plain of the Oklawaha is a ground-water discharge area where wells flow and some upward leakage apparently occurs in association with normal faults and thin parts of the confining layer.

East of the river valley, along the Mount Dora Ridge, there is good evidence of recharge to the aquifer through the sands and clayey sands of the Fort Preston formation. Here the top of the aquifer is higher than immediately to the west due to uplift along normal faults bordering the east side of the graben valley of the river (fig. 18).

West of the Oklawaha River, in the Lake Weir vicinity the believes that important recharge to the Floridan Aquifer takes place through thick, sandy deposits shown as Fort Preston formation in and 14.

Recharge from local rainfall occurs through out the Barge Canal area wherever the materials covering the aquifer permit infiltration. Aside from the principal river flood plains, there is no integrated surface drainage system, and rainfall must either infiltrate the land surface and percolate to the aquifer or be lost to evapotranspiration. Recharge not only occurs in the high slope and ridge areas of the potentiometric surface, but also in the lower slope areas near the spring discharge points, provided the aquifer extends to the surface or is covered with permeable materials. As indicated earlier, some of the highest points on the potentiometric surface, such as those local highs southeast and west of Rainbow Springs, actually are areas of low permeability in the aquifer, and although they are recharge areas the rates are low.

Figure 27 is a map of the water level changes from May 1968 (low-water period) to September 1968 (high-water period). The map shows differences in range of water-level fluctuations from place to place and an approximation of the normal annual change resulting from recharge during the rainy season. It also may be used in conjunction with the geohydrologic maps for evaluating recharge-discharge relationships of individual localities. In general, the change is greatest in areas of recharge and least in areas of discharge. In localities far removed from discharge areas where water-level changes tend to be minimal, recharge is restricted by the poorly permeable Miocene-Pliocene(?) cover. On the other hand, in the locality showing the greatest change in water level, recharge occurs, but is somewhat restricted due to limited permeability of the aquifer. In localities surrounding important discharge areas, significant recharge commonly occurs, even though this recharge is not reflected by marked water-level fluctuations, owing to the short storage time and subsequent discharge nearby.

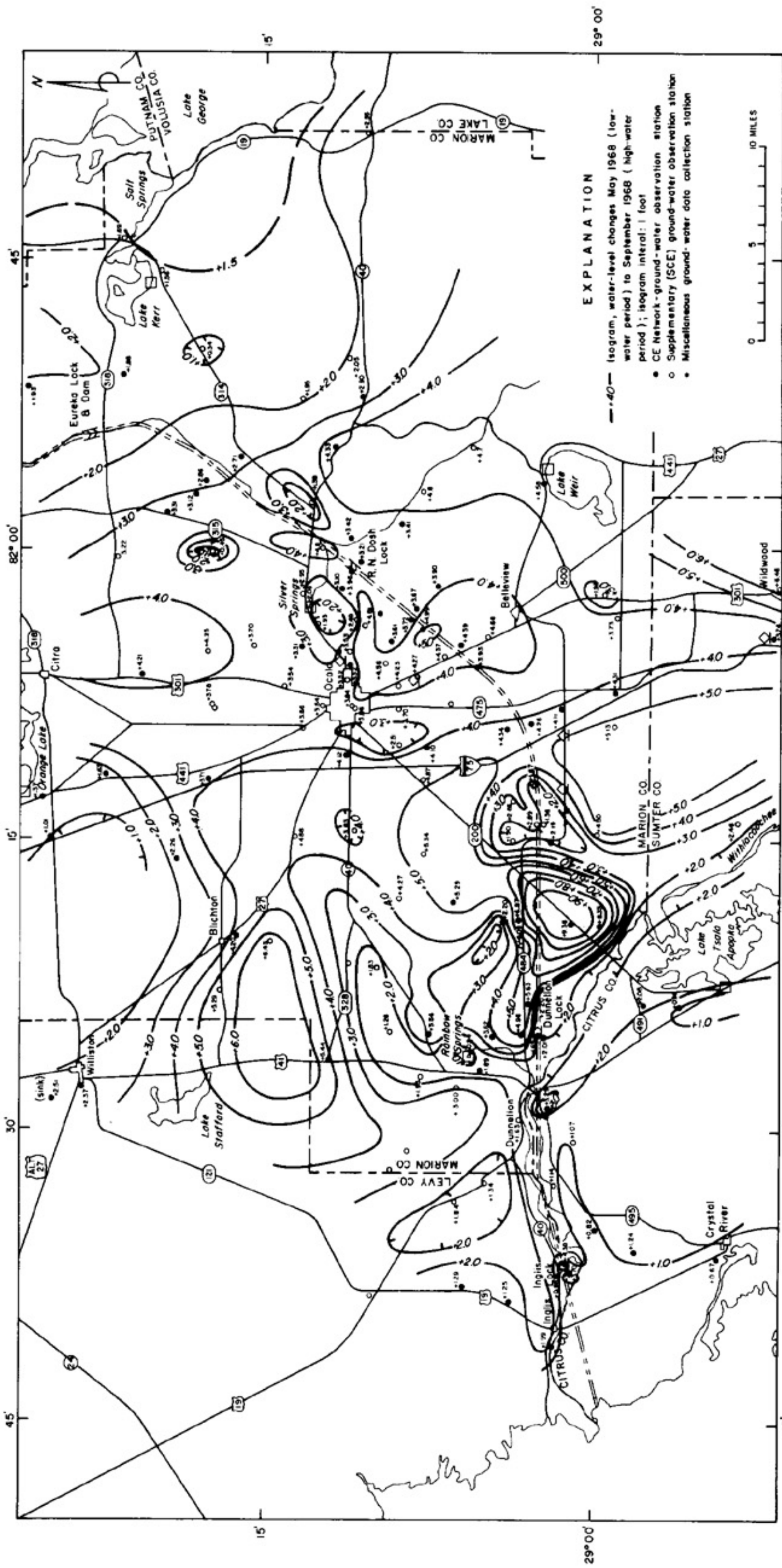


Figure 27. Water level changes in upper part of Floridan Aquifer from May 1968 (low-water period) to September 1968 (high-water period), Cross-Florida Barge Canal area.

Data on natural tritium concentrations in water in the upper part of the Floridan Aquifer were collected from December 1966 to November 1968 to evaluate tritium as a radioactive tracer. Some useful information on recharge characteristics resulted. The scope of this report precludes a detailed discussion of the tritium data, but some remarks pertaining to relationships between recharge and tritium concentration may be made.

Tritium is a radioactive isotope of hydrogen with a half-life of 12.26 years, and as such its presence in water is sometimes useful in age determinations. In very general terms, an appreciable concentration of tritium is an indicator of young water.

Natural tritium is a product of the interaction of cosmic rays with the atmosphere. Incorporated in water molecules in precipitation, it ultimately reaches the land surface and thence the ground-water reservoir. Tritium concentration in water is generally measured and recorded in tritium units (TU), where one tritium unit is defined as equal to one tritium atom in 10^{18} ordinary hydrogen (protium) atoms. The concentration of natural tritium is so small that it does not interfere with the use of the water.

The natural production rate of tritium in the atmosphere is considered constant, but nuclear weapons tests as late as the early 1960's raised the tritium level to many times the normal level. The level declined after the tests but has not returned to normal, estimated by L. L. Thatcher as about 6 TU along the Atlantic Coast (Stringfield, 1966, p. 150).

During the 2-year sampling period in the study area, one or more samples were collected at each of 26 different observation stations. Bimonthly samples were collected and analyzed from ten of the stations, including Rainbow and Silver Springs, Ocala Caverns, and seven wells.

Samples collected irregularly from four wells in the confined part of the aquifer, mostly in the Oklawaha River valley, contained zero to only traces of tritium, indicating complete or nearly complete decay of originally contained tritium. This supports interpretations from other geologic and hydrologic data that the water was old and not subject to direct recharge. On the other hand, all samples collected west of the river valley in the unconfined part of the aquifer had appreciable concentrations of tritium, thus supporting the belief that much of that area is subject to direct recharge. Each of the samples had more than twice the presumed normal concentration of 6 TU in rainfall, but far below peak concentrations measured in rainfall in Ocala in 1963 after the large thermonuclear tests in late 1962, thus suggesting that the high concentrations in the ground water from 1963 recharge had already passed.

The weighted average tritium concentration for 1963 precipitation in Ocala was 620 TU (Stewart and Farnsworth, 1968, p. 281). Concentrations in ground-water samples collected west of the Oklawaha valley during the 1966-1968 sampling period ranged from 13 to 174 TU while rainfall in Ocala ranged from 20 to 158 TU. The concentration in eleven samples collected at Rainbow Springs averaged about 49 TU with a high of 85 TU in May 1967 and a low of 38 TU in March 1968. Silver Springs water during the same period averaged about 47 TU with a high of 150 TU in May 1967 and a low of 25 TU in July 1967. This compares with a count of only 4.2 TU in Silver Springs water in January 1961 (Stringfield, 1966, p.150). The average concentration in the seven wells sampled bimonthly during 1966-68 was about 80 TU, and that in rainfall at Ocala for the same period was about 60 TU.

Ever since the record highs in precipitation in 1963, there has been a general decline in tritium concentration, although seasonal peaks and troughs continue to occur as they would under normal conditions. A similar decline apparently occurs at most ground-water observation stations. A correlation between many of the concentration curves for the ground-water stations and the concentration curve for precipitation is suggested. A lag of a month or two is apparent from the rainfall peaks to the peaks in several of the ground-water curves. Correlation of the curves is very much complicated by variable recharge characteristics and by differential rates of ground-water movement in the area. Much remains to be done to properly integrate and fully evaluate the tritium data, but in the meantime the tritium data provide a useful tool for recognition of some differences in recharge conditions from one place to another within the study area.

The calculated average annual recharge is over 15 inches for that part of the Barge Canal area west of the Oklawaha River. Based on an average spring discharge of 531 mgd (million gallons per day) and an area of 730 square miles, and average recharge of 15.3 inches per year has been calculated for the Silver Springs ground-water drainage area. The average discharge of 468 mgd drained from an area of 645 square miles indicates an average annual recharge of 15.2 inches for the Rainbow Springs drainage area. The combined drainage areas for the two springs cover most of the canal area west of the Oklawaha River, and their recharge characteristics may be considered typical of that area. The Floridan Aquifer receives no recharge in much of the Oklawaha River valley because of the thick, overlying confining layer and the valley is an area of artesian flow.

At localities just east of the river valley, where permeable materials directly overlie the Floridan Aquifer, average recharge may be comparable to that west of the valley.

Characteristics of water movement through the aquifer.--Recharging water moves in essentially vertical paths above the capillary fringe of the unsaturated zone, but after entering the capillary fringe and thence into the zone of saturation the water assumes a horizontal component of movement in a down-gradient direction approximately normal to the contours on the potentiometric surface (fig. 25). In a highly permeable aquifer such as the cavernous limestone of the Floridan Aquifer, the horizontal component is usually dominant. As illustrated by flow line segments on Figure 25, individual flow paths converge into gently sloping troughs in the potentiometric surface as the paths approach points of concentrated discharge. The troughs of preferential or concentrated flow in a limestone aquifer, such as the Floridan, may be indicative of solution channel systems which become increasingly well developed as large springs such as Silver and Rainbow Springs are approached. Without an increase in hydraulic gradient, such a progressive increase in permeability is essential in order that the aquifer be able to accommodate the increasing volume of flow converging on the area of discharge.

Intervening parts of the aquifer that separate zones of preferential flow are represented by gentle ridges on the potentiometric surface (fig. 25), and transmissivities in these ridge zones may be many times less than those represented by the potentiometric troughs. Flow in the ridge areas tends towards the troughs of preferential flow rather than along independent direct routes to the area of discharge.

Maps of the potentiometric surface are valuable indicators of the two dimensional or plan-view pattern of ground-water flow. The pattern of flow in the Ocala vicinity is illustrated in Fig. 25 by the flow-path segments plotted on the potentiometric surface. However, the two-dimensional flow-pattern is only a part of the total flow picture, and flow in the third dimension, or vertical plane of flow, may be illustrated by the vertical sections drawn along flow paths plotted on the potentiometric surface map. Figure 28 shows two such sections (X-Y and X-Z) which have been drawn leading to Silver Springs across the canal alinement. The locations of the lines of section are shown on Figures 32 and 34.

No actual measurements are available, along the lines of section, of changes in hydraulic potential with depth in the aquifer. If detailed information were available on the vertical head distribution, it would be possible to draw actual paths of flow normal to lines of equal potential. However, through knowledge of other hydrologic and geologic data it is possible to sketch presumed paths of flow based on assumptions drawn from various theoretical analyses of ground-water flow such as discussed by Freeze and Witherspoon (1966 and 1967), Thrailkill (1968), and Toth (1963).

A ground-water basin is a three-dimensional closed system that contains all of the flow paths followed by all of the water recharging the basin. Thus, the Silver Springs and Rainbow Springs drainage areas or basins may be described as bounded by vertical flow lines at the potentiometric divides, by a nearly horizontal “impermeable” bottom (possibly the upper part of the Avon Park Limestone), and by an irregular top (the potentiometric surface).

It is the change in ratio of permeability that is needed for depth limitation of the basin rather than the presence of a truly impermeable layer. Also, the presence of comparatively shallow subregional basins, as the Silver and Rainbow Springs basins may be considered, does not preclude movement of ground water in an underlying regional basin not directly involved in the subregional or local recharge-discharge relationships active in the shallow basin area.

It is thought by the writer that upper part of the Avon Park Lime-stone, although not impermeable in the true sense, is sufficiently less permeable than the overlying Ocala Limestone that it may constitute the bottom to important ground-water basins in the Barge Canal area. The average thickness of the Ocala Limestone in the Ocala-Silver Springs area is estimated to be about 100 feet. Therefore, most of the flow to the Silver Springs may

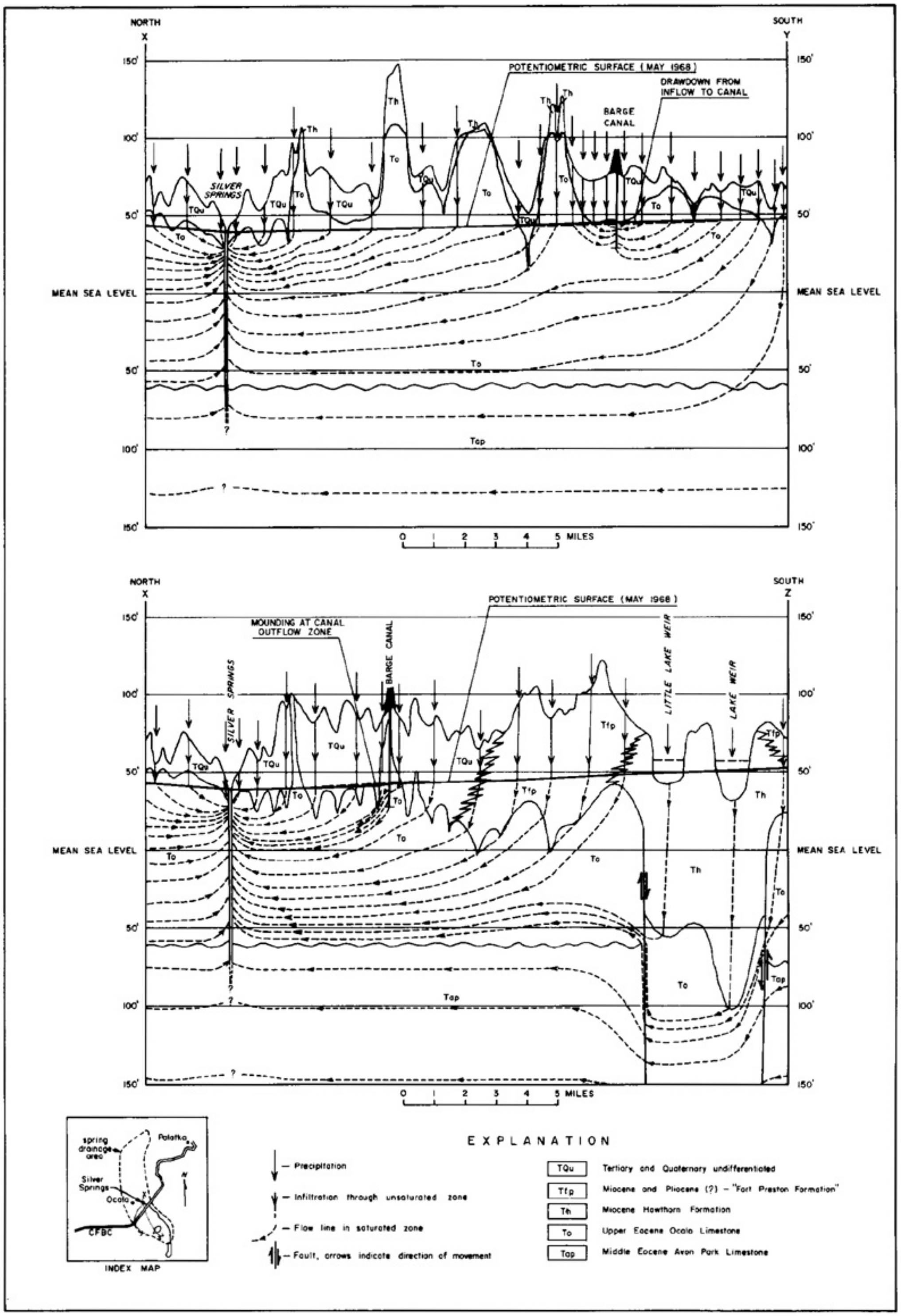


Figure 28. Ground-water flow sections X-Y and X-Z, Ocala vicinity.

take place through the upper 100 feet or so of the aquifer. If such a situation does exist, it is complicated in places by the normal faulting described earlier. That is, subordinate amounts of flow probably pass through or adjacent to older rocks of the aquifer which are in fault association with the Ocala Limestone.

The data used to map the potentiometric surface in this study are mostly from wells penetrating only the upper part of the aquifer, that is, in most wells to depths less than 75 feet below the top of the aquifer. Water-level data from deeper zones of the aquifer in the Silver Springs-Ocala vicinity are limited to several deep wells in the city of Ocala. These data indicate no appreciable change in head with depth. Three of the wells, all city of Ocala public-supply wells in the south-central part of town, bottom in the Avon Park Limestone, and are open to the aquifer between about 120 and 350 feet below land surface. A storm drainage well in east Ocala bottoms at 500 feet below land surface at a depth near the top of the Lake City Limestone. An industrial well at the Libby-McNeil citrus plant on the north edge of the city is cased to 850 feet and is completed open hole to 1,083 feet near the bottom of the aquifer. During the drilling of the well in late 1951 and early 1952, static water-level measurements were made and posted on the driller's log at different depth intervals in the cased zone as well as in the present open hole. No appreciable changes in head with depth were indicated.

Various hydrogeologic factors suggest the regional presence of an appreciable reduction in permeability from the Ocala Limestone to the Avon Park Limestone. Potentiometric-surface maps indicate pronounced highs in areas where the Avon Park Limestone crops out or where the Avon Park is close to the surface, southeast and west of the Rainbow Springs area (figs. 13, 23, and 24). The highs are the result of the relatively low permeability of the Avon Park Limestone. The primary permeability of the Avon Park was reduced later by dolomitization and it is thought, still further by sand-filling of solution cavities, some of which cavities may have been developed prior to dolomitization. Not only were effective intergranular porosity and permeability reduced in the Avon Park Limestone in north-central Florida by recrystallization during dolomitization but, because the dolomite is much less soluble than limestone, development of solution channel porosity and permeability from ground-water circulation also were reduced. Dewatering operations at the Inglis Lock construction site, pumping tests at the future site of the Dunnellon Lock, and data from observation wells in the vicinity of the Dunnellon Lock site all indicate relatively low permeability in the upper part of the Avon Park Limestone, as compared with the high permeability of the more or less undolomitized Ocala Limestone.

A two-dimensional section through a ground-water basin is representative of the three-dimensional basin if it is taken parallel to the direction of dip of the water-table slope. Freeze and Witherspoon (1966 and 1967) investigated details of steady flow in regional ground-water basins by using digital computer solutions of numerous mathematical models. By machine plotting of head distributions, they were able to determine ground-water flow patterns resulting from the effects of many different types and combinations of types of geologic and hydrologic conditions. By using their results and that of others in conjunction with a knowledge of certain geologic and hydrologic factors in the study area, sections X-Y and X-Z (fig. 29) were prepared as conceptual models of the third dimension of flow in that part of the Silver Springs drainage basin through which the Barge Canal will pass. Flow lines are sketched in the sections in accordance with the kind of head distribution the writer would expect from the set of recharge, discharge, and geologic conditions illustrated in the sections. It was necessary to use a highly exaggerated vertical scale; therefore, the reader is cautioned to take cognizance of some avoidable distortions of the true scale model.

Head distribution is dependent on the locations of recharge and discharge areas. In a recharge area the flow is downward, so that the head decreases with depth. Conversely, in a discharge area flow is upward so the head increases with depth. Between a recharge area and a discharge area the flow is nearly horizontal so that the change in head with depth is slight.

Changes in head with depth may be represented in vertical section by equipotential lines. In a recharge area the water table in the upgradient direction of the water table. In a discharge area they slope in the downgradient direction. In intermediate areas of no recharge or discharge, the equipotential lines are approximately perpendicular to the water table. If the water table is perfectly flat, no horizontal component of flow exists, any flow is vertical, equipotential lines are horizontal, and the water table is an equipotential line.

Recharge moves downward through the unsaturated zone along approximately vertical lines, but when it passes into the capillary fringe of the zone of saturation, a horizontal component of movement is applied parallel to the down-gradient direction of the water table. As long as there is additional recharge in the downstream direction there will remain a vertical component of flow resulting in a gradual deepening of flow paths, so that recharge entering at the divide will have to travel deepest in the aquifer to reach the discharge area. Where there is little or no recharge, the equipotential lines will tend to become vertical, and the flow paths horizontal. However, if ground water then flows from under a covered area and enters an area downgradient where recharge occurs, a downward component of flow will again result because of the higher head at shallower depths resulting from the recharge. As flow reaches the discharge area, the flow paths turn upward and the hydraulic head lessens toward the water table.

Section X-Y (fig. 28) represents the flow situation just described. Under natural conditions, because of low topography and a very gently sloping potentiometric surface, no small subbasins or local basins, as described by Toth (1963), exist along the line of section. However, as shown on the section, a subbasin will exist when the Barge Canal is completed, because water will discharge from the aquifer to the canal. Consequently the canal will be subordinate discharge area within the Silver Springs drainage basin. Along section X-Z (fig. 28) no subbasin exists now, nor will one when the canal is completed. In this area water will discharge from the canal to the aquifer.

In section X-Y, Ocala Limestone of the Floridan Aquifer is present at or above the potentiometric surface over most of the length of the section. However, along more than half the length of section X-Z, the potentiometric surface is well above the top of the Ocala Limestone, within the Miocene and younger sands and clays which have much lower permeability than the Ocala. In addition to this vertical change in permeability, normal faults present in the section place Ocala Limestone and Avon Pan Limestone in fault contact with each other, thus producing lateral changes in permeability. These permeability variations have very important effects upon the shape of the flow pattern.

If a formation of comparatively low permeability separates the water table from a formation of higher permeability, the more permeable formation is in effect an aquifer with essentially horizontal flow that is being recharged from above. Consequently, the vertical component of flow in the upper layer, such as in the sandy Fort Preston formation in section X-Z, is much more pronounced than in the case where the water table is at or below the top of the much more permeable limestone (Freeze and Witherspoon, 1967, figs. 2A-D). The greater the contrast between the permeability of the upper and lower layers, the steeper the flow lines in the upper layer. In the case of a formation of high permeability overlying one of comparatively low permeability, such as the Ocala overlying the Avon Park, there is no appreciable change in potential, and flow remains essentially horizontal (Freeze and Witherspoon, 1967, fig. 2E).

On section X-Z the down-faulted block of Ocala Limestone would receive recharge along the fault from the structurally higher Ocala Limestone upstream, and possibly a small amount of recharge from the overlying Hawthorn Formation.Where flow in the Ocala meets the less permeable Avon Park, an upward vertical component of flow along the fault might be expected. Although some water would flow laterally into the Avon Park, appreciable flow would be directed into the structurally higher Ocala Limestone downstream from the fault, and from here the flow attitude would change from upward to approximately horizontal because of additional recharge from above. This type of flow condition may account for ground-water movement under the confining layer from recharge area east of the Oklawaha River valley (fig. 20, section A-A'), northwestward to Silver Springs.

Circulation in a typical limestone aquifer under water-table conditions tends to concentrate in a moderately thin zone just below the water table. Deep lying limestones having impeded circulation may be almost hydrologically inert (Stringfield and LeGrand, 1966, p. 8). Also, work by others as noted by Toth, (1963, p. 4811) indicates that only a small percentage of the total amount of water occupying most ground-water basins actively participates in the hydrologic cycle, and that the deeper the basin, the smaller is this percentage. This is not to say that there is no circulation in deep zones of good permeability in the Floridan Aquifer where solution channel systems have been developed in the geologic past, but merely that such circulation very likely is not involved to a great extent in the discharge at Silver and Rainbow Springs. The deeper circulation takes much longer to move through the aquifer than does the shallow, and most of the deep flow probably bypasses these inland spring discharge areas and moves toward coastal and offshore discharge areas. In fact, a zone of stagnation could exist at depth, centered in the regional potentiometric saddle area occupied by the two springs. This could result where regional flow systems on the north and south meet, and deep flow is diverted toward the coasts.

The foregoing data and discussion indicate that in the Silver and Rainbow Springs drainage basins, the Ocala Limestone may be treated as a moderately thin zone of circulation (100± feet in the Ocala vicinity) in the top of the Floridan Aquifer. In places this zone of circulation may extend downward to moderate depths into old unplugged solution channel systems in the upper part of the Avon Park Limestone, but circulation actively involved in the flow to the springs should diminish rapidly with depth in the formation (fig. 29). In places, such as areas of concentrated solution channeling indicated by troughs on the top-of-rock map (fig. 19), the Ocala Limestone may be considerably thinner than 100 feet, and flow may be concentrated at the base of the Ocala and in solution channels not far below the top of the less permeable Avon Park Limestone where downward movement in the aquifer is restricted.

As discussed by Toth (1963), theoretically three types of flow systems may occur in a ground-water basin: local, intermediate, and regional. The different types of systems are separated by subhorizontal boundaries, with local systems at the top, underlain in turn by intermediate and regional systems. The local systems are separated from each other by subvertical boundaries resulting mostly from topographic relief. The higher the topographic relief, the more important are the local systems. Since topographic relief is so subdued, and permeability of the aquifer is so great, virtually none of the so-called local basins exist in the Floridan Aquifer in the canal area. However, the Rainbow and Silver Springs ground-water drainage areas may be categorized as intermediate flow systems, the two basins being separated from the deeper regional flow system of the Floridan Aquifer in Peninsular Florida by the upper part of the Avon Park Limestone, which may be treated as the sub-horizontal boundary.

Certain water-quality factors support the premise that most of the flow of both Rainbow and Silver Springs is from near the top of the aquifer, although the more highly mineralized character of Silver Springs water indicates a longer residence time and a somewhat deeper source than that of water issuing from Rainbow Springs. Dissolved solids concentrations in the spring waters compare favorably with ranges of concentrations found in water from wells in the shallow part of the aquifer, and sulfate (SO_4) concentrations in the spring waters indicate the upper part of the aquifer is the major source of spring discharge. Like dissolved solids, sulfate increases markedly with depth, probably to a large extent the result of the more common occurrence of gypsum and anhydrite (calcium sulfate) in rocks deeper in the aquifer. In Ocala a concentration of about 260 mg/L of sulfate occurs in water from a well open to the aquifer from 850 to 1,083 feet. Concentrations of about 150 mg/L are found in the city of Ocala public-supply wells open to the interval 120 to 350 feet, and concentrations ranging between 30 and 50 mg/L occur in Silver Springs water.

In the Ocala-Silver Springs vicinity an average sulfate concentration of 22 mg/L has been calculated for water samples collected from 18 wells in the upper part of the aquifer. The well waters range in sulfate concentrations from 0.0 to 92 mg/L and the wells range in depth from 40 to 200 feet. The average well depth is 102 feet with an average aquifer penetration of 40 feet. Twelve of the wells have total depths less than 100 feet and waters averaging 14.3 mg/L sulfate. The 12 wells have an average depth of 74 feet with an average aquifer penetration of 42 feet.

Calculations based on average concentrations of 22 mg/L of sulfate near the top of the aquifer, 260 mg/L near the bottom of the aquifer and 40 mg/L in Silver Springs water indicate that the spring water is a mixture of about 92 percent shallow water and 8 percent deep water. If the foregoing sulfate concentrations for the upper part of the aquifer and for the springs are compared with 150 mg/L in the lower part of the Avon Park, a Silver Springs water mixture consisting of about 86 percent upper aquifer water and 14 percent lower Avon Park water is indicated.

Specific conductance of water from Silver Springs is about three times that of water issuing from the head of Rainbow Springs. Sulfate concentration of water at the head of Rainbow Springs is only 3 to 4 mg/L. But water in Rainbow River near Dunnellon, about 4 miles below the head of the springs, has a specific conductance of nearly twice that at the spring head and a sulfate concentration of 13 to 16 mg/L. The river gains ground water from numerous spring vents in its bed between the head of the spring and Dunnellon, and presumably much of the water gained is from deeper in the aquifer and of longer residence time than that issuing from the head of the spring.

Judging from the difference in quality, the average residence time and depth of travel in the aquifer for Rainbow Springs water is probably less than that for Silver Springs water.

One reason may be the fact that the Rainbow Springs drainage area is higher on the flank of the Ocala Uplift. Consequently the average thickness of the Ocala Limestone is less and, therefore, rapid flow may be concentrated at comparatively shallow depths in thin zones near the Ocala-Avon Park contact, where somewhat reduced opportunity for accumulation of dissolved solids may be expected.

Also, the low mineralization at the head of Rainbow Springs is believed at least partly due to nearby abundant recharge from the Alachua Formation, which covers the aquifer immediately north and west of the spring head (fig. 13). Water infiltrating to the aquifer has little opportunity to dissolve materials of the sandy Alachua Formation, largely because it has only a short distance to travel to the spring, and then through what is probably a well developed solution channel system near the Ocala-Avon Park contact.

The temperature of water in both Silver and Rainbow Springs indicates that it had percolated only to moderate depth before being discharged at the spring orifices. Collins (1925, p. 98) states, from a study of over 3,000 records of ground-water temperatures, that under normal conditions the temperature of ground water obtained at a depth of 30 to 60 feet will generally exceed by 2° or 3°F (Fahrenheit) the mean annual air temperature. It is also recognized that the most probable average rate of increase in temperature of ground water with depth is 1° F for each 64 feet of depth.

The mean annual air temperature for the north-central Florida climatic division, as determined by the U. S. Weather Bureau, is 71.7° F. Annually the temperature of the water discharging from both Silver and Rainbow Springs ranges only between 73° and 74° F. On the basis of Collins (1925) discussion, most of the spring water comes from depths ranging not much in excess of 100 feet. This interpretation compares favorably with temperatures measured in wells in the area.

The increase in temperature with depth in wells in and around Ocala appears to follow the normal gradient. The temperature of water in the city of Ocala wells open to the approximate interval 120-350 feet below land surface is a consistent 76° F. According to the normal gradient, 76 degrees indicates a depth of about 260 feet. The deep Libby-McNeill well in Ocala is open to the aquifer between 850 and 1,083 feet and the water has a temperature of a t 86° F. This temperature indicates a water depth of about 900 feet. The temperature of water from most wells in the Barge Canal area that are less than 125 feet deep is 74° F or less, with an average of about 73° F. Water from some of the most shallow wells (40± feet) has a temperature of about 71°F

An average transmissivity needed by a 100-foot thick aquifer in the Ocala vicinity to transmit the flow to Silver Springs has been calculated by treating Silver Springs as a discharging well and evaluating its effect upon the potentiometric surface. As expected, average transmissivity is greatest in the area immediately surrounding the springs and diminishes as the distance from the springs increases. Through use of Darcy's law expressed by the following formula, average transmissivity based on flow across a closed potential contour has been calculated at three separate closed potential contours surrounding Silver Springs:

$$T = (Q-R)/(IL)$$

where T = transmissivity, in gpd/ft (gallons per day per foot width of aquifer)

Q = spring discharge, in gpd

R = recharge in gpd in area enclosed by potential contour at which average transmissivity is calculated

I = average hydraulic gradient across potential contour, in ft. per ml.

L = width of flow channel or length of closed potential contour, in miles

The map of the potentiometric surface in September 1968 (fig. 26) was used, and average transmissivities of 10, 20, and 35 mgd/ft, respectively, were calculated for the 45-, 44-, and 43-foot potential contours surrounding Silver Springs. If the effective thickness of the aquifer supplying water to Silver Springs is 100 feet, average permeability at the 45-, 44-, and 43-foot contours is calculated to be 100,000, 200,000 and 350,000 gpd/ft^2.

The coefficient of storage has not been calculated for the aquifer in the Ocala-Silver Springs area, although the aquifer obviously contains tremendous quantities of water as evidenced by the good intergranular cavernous nature of the limestone, and the high flow at the springs. The dynamic conditions in the system caused by continuous discharge and variable, but probably nearly continuous, recharge preclude calculating a storage coefficient by means of a analysis of the relationship between spring discharge and changes in the potentiometric surface.

Neither transmissivity nor storage coefficient of the aquifer in the vicinity of Rainbow Springs Has been calculated. The lack of closed potentiometric contours and the wider distribution of individual discharge points limit the use of the methods used in the Silver Springs area for calculation of average transmissivity. However, from known geologic and hydrologic facts, it is reasonable to expect that the aquifer characteristics in the Rainbow Springs area are comparable with those in the Silver Springs area.

Flow-net analysis of the upper part of the Floridan Aquifer in the vicinity of Silver Springs and the Summit Pool of the canal is discussed later.

Quality of water in the aquifer.--Water in the Floridan Aquifer in most of the Barge Canal area is typically a hard calcium bicarbonate type and generally of very good chemical quality. It is commonly free of bacteriological contamination and is considered excellent for public and domestic supplies. The mineral content of the water usually increases with depth.

Except in the lower Oklawaha River valley below Eureka, salt-water is not a problem in wells. Iron is sometimes bothersome in water from wells drilled near ponds, lakes, and old plugged sinkholes, and dissolved hydrogen sulfide gas frequently causes odor problems in water supplied from wells located in the river valleys and in some extraordinarily deep wells. Iron may often be avoided by drilling deeper, but the hydrogen sulfide content is likely to increase with depth. In most parts of the area abundant supplies of potable water are obtainable at shallow depths, so most wells are drilled to depths considerably less than 200 feet.

Figures 29, 30, and 31 are contour maps of specific conductance, chloride concentration and total hardness, respectively, of water in the upper part of the Floridan Aquifer. In general, high specific conductance and hardness indicate a high dissolved-solids concentration, and suggest a comparatively long residence time in the aquifer, and perhaps that the water has traveled farther and deeper through the aquifer. Chloride concentrations in the upper part of the aquifer in the study area are for the most part very low, under 10 mg/L. A chloride concentration of 24 mg/L has been reported in water from a deep well in Ocala open to the aquifer from 850 to 1,083 feet. The altitude of the fresh water head in the Ocala vicinity indicates that the fresh water-salt water interface, based on the Ghyben-Herzberg principle, should be more than 1,500 feet below land surface. A definite trend toward higher chloride concentration is noted in wells in the Oklawaha River valley, downstream from the Silver River, as the potentiometric surface of the Floridan Aquifer declines northeastward. Also, at least moderate quantities of ground water of comparatively high chloride content are apparently leaking upward into the Oklawaha River channel downstream from Eureka.

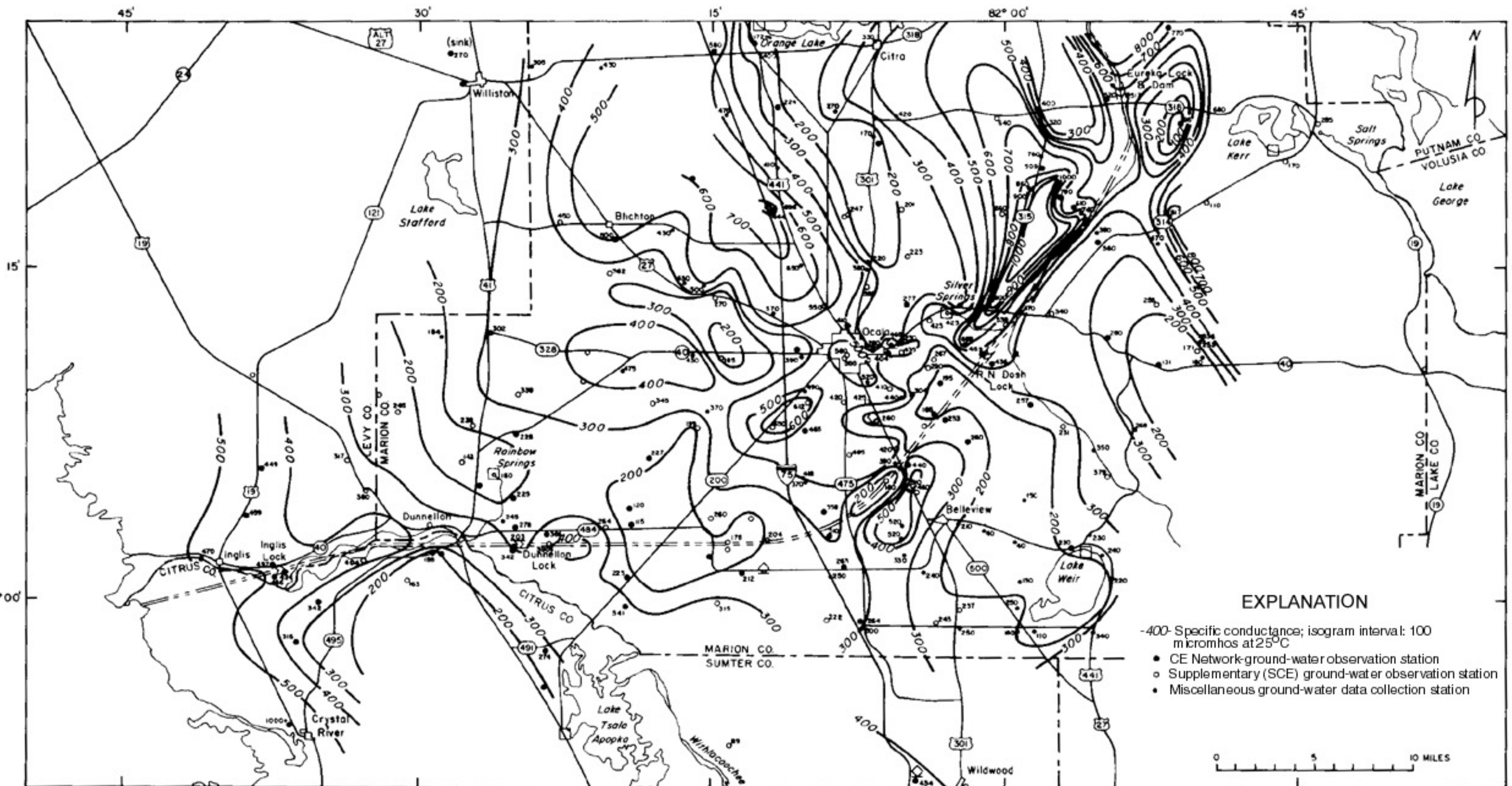


Figure 29. Specific conductivity of ground-water in upper part of Floridan Aquifer, Cross-Florida Barge Canal area.

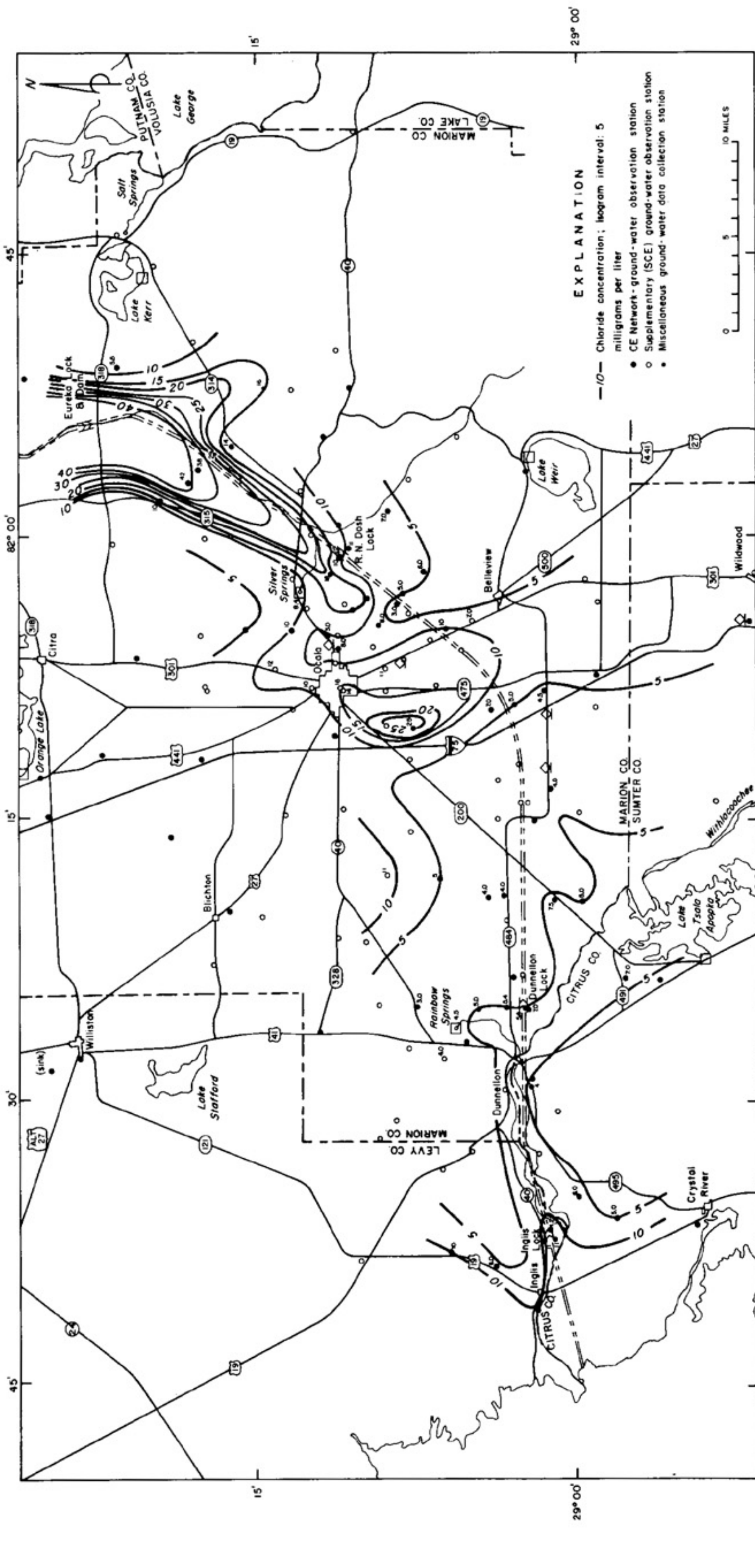


Figure 30. Chloride content of ground water in upper part of Floridan Aquifer, Cross-Florida Barge Canal area.

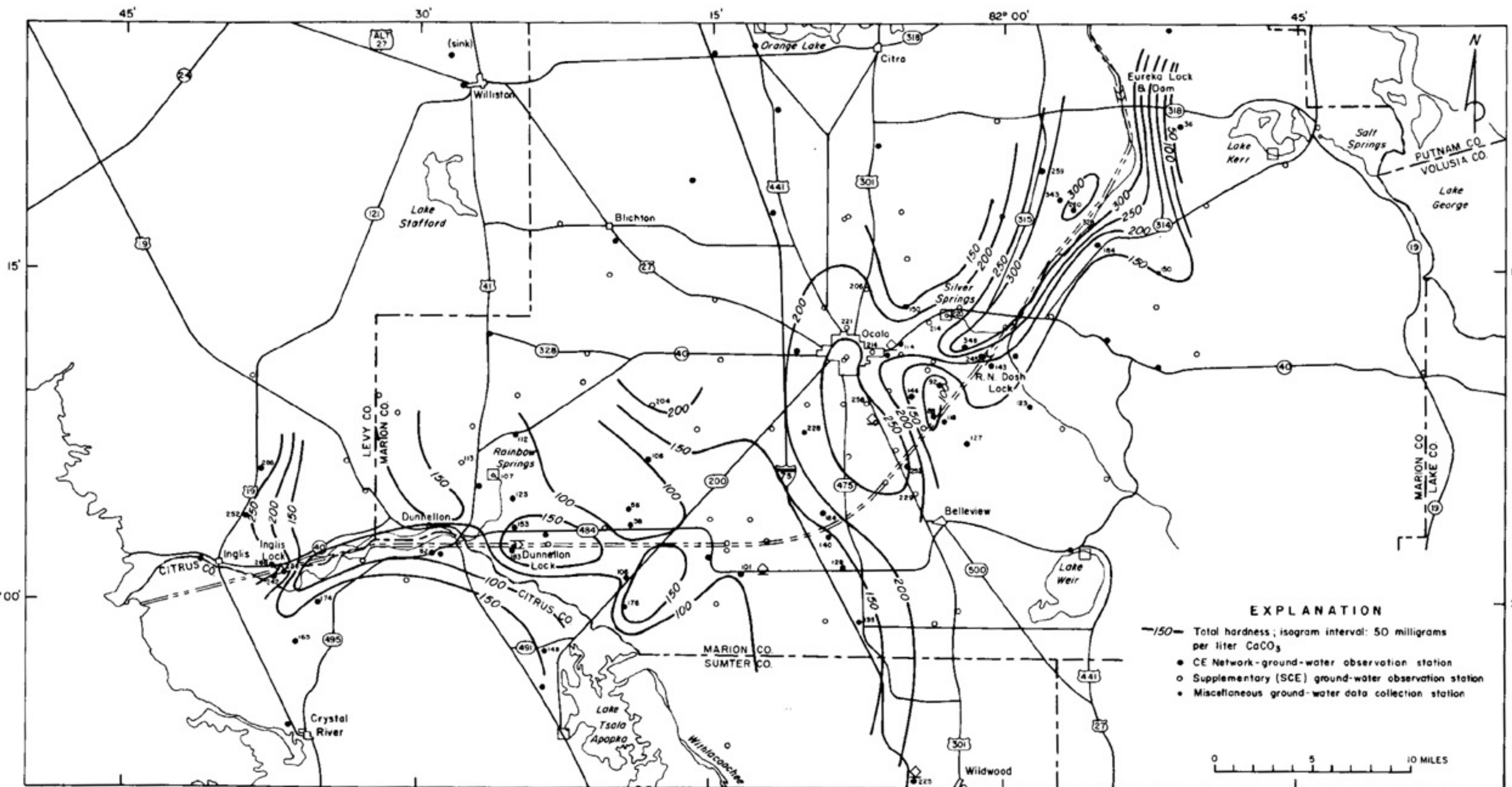


Figure 31. Total hardness of ground water in upper part of Floridan Aquifer, Cross-Florida Barge Canal area.

Specific conductance and temperature are the two most common types of ground-water quality data collected in the investigation. Temperatures increase slowly with depth, and for the depth range of most observation wells most water temperatures are in the range of 72° to 74° Fahrenheit, with little, if any seasonal change. Values of specific conductance ranger just under 100 micromhos to as high as 1,000 micromhos at different Floridan Aquifer well sites in the study area. The seasonal range of specific conductance is quite small, but areal differences are sufficient to be diagnostic of important hydrologic and geologic relationships.

The specific conductance map aids in identification of areas major important recharge. Water of low specific conductance occurs where direct recharge might be expected. Water of high specific conductance is found under and adjacent to the remnants of poorly permeable materials of the Hawthorn Formation where direct recharge might not expected. In areas of abundant direct recharge, older water, including ground water prom sources more distant from the discharge area, is forced to travel deeper flow paths as additional recharge enters the zone of saturation. Thus water with low dissolved-solids concentration is found near the top of the saturated zone. On the other hand, ground water moving beneath areas of rest restricted recharge from a nearby recharge area tends to remain in flow paths near the of the zone of saturation because of a lack of additional re to force it downward. In this case, because water remains high in the a aquifer as dissolves soluble minerals, water high in the aquifer acquired comparatively high specific conductance.

A comparison of the conductance and the potentiometric surface maps shows some interesting relationships between troughs of principal ground-water flow and low specific conductance. A good correlation between the areas of most active recharge and the zones of major ground-water movement is indicated. Northwest and southeast of Silver Springs, elongate bands of low specific conductance coincide approximately with major flow zones to the spring. Specific conductance of the spring water is higher than that of the water in the high flow zones, because the spring flow is a mixture of water from the major flow zones and more mineralized water moving to the spring from areas of restricted recharge and from deeper parts of the aquifer, especially along the edge of the confining layer east of Silver Springs.

Similar flow zones containing water of low specific conductance leading to Rainbow Springs are evident. The low specific conductance of water discharging at the head of Rainbow Springs compares favorably with the specific conductance of well water collected in the potentiometric trough leading to the head of the springs from the northwest. It was noted earlier that the specific conductance of Rainbow River water near Dunnellon is about twice that of water at the head of the springs, thus indicating an even higher specific conductance for some ground water entering the stream channel between the spring head and Dunnellon.

The highest specific conductance values appearing on Figure 29 occur in the area northeast of Silver Springs where geologic maps (figs. 15 and 19) show the top of the aquifer to be as much as 200 feet below sea level and covered with as much as 250 feet of poorly permeable materials. The high specific conductance tends to indicate old, very slowly moving ground water not involved in the dominant ground-water circulation system active at shallower depths.

The aquifer in most areas is protected from bacterial contamination and other types of pollution by a sand, clayey sand, or clay cover at least several tens of feet in thickness (fig. 15), which acts as an excellent natural filter. Where the Floridan Aquifer is exposed, or is recharged directly through natural sinkholes or drainage wells, risks of contamination exist. However, unless contaminated recharge flows directly into a well developed system of flow channels, the limestone of the aquifer may adequately filter out the contaminants near the point of recharge. Thus, the contamination of the ground water may be a very localized problem. However, the risk of widespread contamination increases with proximity of polluted recharge to a major discharge area, because of the highly developed solution channel system common to the vicinity of major discharge.

Qualitive Relationships between the Cross-Florida Barge Canal and the Ground-water Regime

General.--From the west edge of the Oklawaha River valley, about a mile and a half southwest of R. N. Dosh Lock, to the Gulf, most of the canal will be excavated into the rocks of the Floridan to depths ranging from 12 feet to 27 feet below the water table (fig. 2). Northeastward from that point, a mile and a half southwest of Dosh Lock, the canal will no penetrate the Floridan Aquifer. Most of the way to the St. Johns River, the canal and its reservoirs will be located in the Oklawaha River valley, and will be separated from the Floridan Aquifer by sediments of comparatively low permeability which underlie the valley floor. In places, the Floridan Aquifer through these poorly permeable materials into the river channel probably occurs. Also, minor leakage occurs in association with some fault zones in and along the edges of the Oklawaha River valley, particularly near the east edge. This is indicated by small rather abrupt increases in either stream discharge or in specific conductance and chloride content of the river water, as well as observed relationships between two possibly fault emplaced outcrops of the Hawthorn Formation.

Water levels in all pools of the canal, except the Summit Pool, will be controlled directly by dams and spillways. The stage of the Summit Pool will fluctuate in accordance with the natural rise and fall of the potentiometric surface of the Floridan Aquifer, if all lockage, leakage and evapotranspiration losses are replaced with pumpage from the lower pools. However, the controlled stage of Eureka Pool will exert a partial indirect control on the stage of the Summit Pool.

Rodman Pool.--The planned operating stage of 20 feet above mean sea level for the Rodman Pool approximates the natural potentiometric surface of the Floridan Aquifer in the area. Under pre-canal conditions the potentiometric surface is higher than the river stage, and most wells drilled in the river valley flow. In the upper end of Rodman Pool, the stage should not exceed the potentiometric surface, and little or no change from normal conditions should occur, whereas the lower reaches of the pool stage will at times be higher than the potentiometric surface and some downward seepage through the confining layer could occur, possibly causing a moderate rise in the potentiometric surface. A rather shallow fresh water-salt water interface exists in the Floridan Aquifer in the area (Bermes and others, 1963, fig. 34), and a rise in the potentiometric surface would tend to lower the interface. Any infiltration of water from the reservoir downward through the confining layer would be slow.

The stage of that reach of the canal which connects Rodman Pool at St. Johns Lock to the St. Johns River will be naturally controlled by the stage of the river, and the canal should have little or no effect on the ground-water regime of the area. The approximately 20-foot stage differential between Rodman Pool and the lower reach of the canal will tend to minimize the locking up of possibly slightly brackish water from the St. Johns River into the pool.

Eureka Pool.--At the Eureka Lock and Dam there will be a possible stage differential of 16 to 20 feet between the Eureka Pool and the Rodman Pool. The present design for Eureka Pool calls for a range in stage of 38 to 40 feet above mean sea level, but if necessary the pool channel may be dredged 2 feet deeper to permit a range of 36 to 40 feet in pool stage. The Eureka Pool level will be about 15 feet higher than the natural potentiometric surface in the vicinity of the downstream part of the pool. Therefore, when Eureka Pool reaches its operating level, any leakage through the confining layer will be downward instead of upward, and the potentiometric surface in the vicinity of the north, or downstream, end of the Eureka Pool will tend to rise to a point of equilibrium with the stage of the pool. Due to the steep slope of the natural potentiometric surface from the west down to the river, thought caused in part by the presence of a permeability barrier resulting from graben faulting of the river valley, changes in the potentiometric surface more than a mile or two west of the lower end of the pool should be small. However, if hydraulic connection between the pool and the artesian aquifer is adequate, it is possible that in the Eureka-Fort McCoy area where the Floridan Aquifer is still confined, ground-water levels may be raised sufficiently to reduce downward seepage from wet lands, and thus possibly result in maintenance of higher water levels in ponds and marshy areas than would be the case under natural conditions. East of the dam site and under the Mount Dora Ridge, the natural potentiometric surface rises only 3 or 4 feet before it flattens into a broad ridge and then slopes northeastward toward the St. Johns River valley. In this area east of the dam, more leakage should take place into both the shallow and the Floridan Aquifers than is likely to occur to the west.

Beneath Mount Dora Ridge the Floridan Aquifer stands high, probably at least partly due to faulting, and is covered by sands and clayey sands of the Fort Preston formation. Here the water levels in the shallow aquifer and in the Floridan Aquifer seem to coincide, and direct recharge to the Floridan Aquifer apparently takes place.

Leakage through the confining layer beneath the Eureka Pool and directly into sands of the shallow aquifer along the pool 's east side should tend to raise ground-water levels for some distance to the east. However, because the aquifers just east of the pool's lower end are essentially unconfined and because loss of head through the clayey sands making up much of the Fort Preston formation should cause the water table to slope steeply eastward, a significant rise in ground-water levels should extend little more than a mile or two east of the reservoir, and any rise in levels east of Mount Dora Ridge should be small (fig. 5). Loss of head through the clayey sands making up much of the Fort Preston Formation should cause the water table to slope steeply, and any rise in ground-water levels east of Mount Dora Ridge should be small (fig. 5). The low ridge in the potentiometric surface now separating the Eureka area and Lake Kerr will diminish because of the rise in water level on the west side of the ridge; but the effect on ground-water levels in the vicinity of Lake Kerr should be negligible. Any rise in ground-water level under Mount Dora Ridge should cause no significant increase in wet lands, as the land surface, for the most part, will still be far above the water table.

Figure 32 shows the writer's concept of the possible effect of the Eureka Pool on the potentiometric surface of the Floridan Aquifer. Any significant adjustment of the potentiometric surface should extend southwestward little farther than to the vicinity of Gores Landing, which is about midway between Eureka and Dosh Locks. Upstream from this area the potentiometric surface exceeds the planned maximum stage for the Eureka Pool. Also, in much of the area surrounding the upper reaches of the pool, a confining layer separates the shallow aquifer from the Floridan Aquifer, with the water table in the shallow aquifer frequently above the planned stage for the Eureka Pool.

Where water is expected to leak into the aquifers from the Eureka Pool, the sands and clayey sands through which the pool water will seep will tend to filter out any particulate materials that might contaminate the pool water. The natural filter, of course, will not filter out dissolved contaminants if present in the pool waters.

The controlled water level in Eureka Pool will have an important stabilizing effect on the water level at the head of Silver Springs. It will also, in turn, limit the range in stage of the Summit Pool, although there is no direct hydraulic connection through the aquifer between Eureka Pool and Silver Springs or between Eureka Pool and Summit Pool. The water level at the springs is naturally controlled by the interrelationship of the ground-water pressure at the head of the springs and the surface-backwater effect extending up the Silver River from the Oklawaha River. A similar relationship will exist between the Eureka Pool and Silver Springs. On the other hand, there will be a direct hydraulic connection through the aquifer between the head of Silver Springs and the Summit Pool. The stage at the head of the Springs is directly related to ground-water levels in the Summit Pool area; therefore, the controlled stage of Eureka Pool will act indirectly as an important partial control on the stage of the Summit Pool, and it will tend to reduce the range of natural ground-water level fluctuation in the area.

Water discharged from Silver Springs will continue to flow over the confining layer down the Silver River to empty into the Eureka Pool, as it now empties into the Oklawaha River. Regulation of the stage of the Eureka Pool within a range of 4 feet (36-40 feet above mean sea level), will in turn control the stage of Silver Springs within an estimated range of about 5 feet and

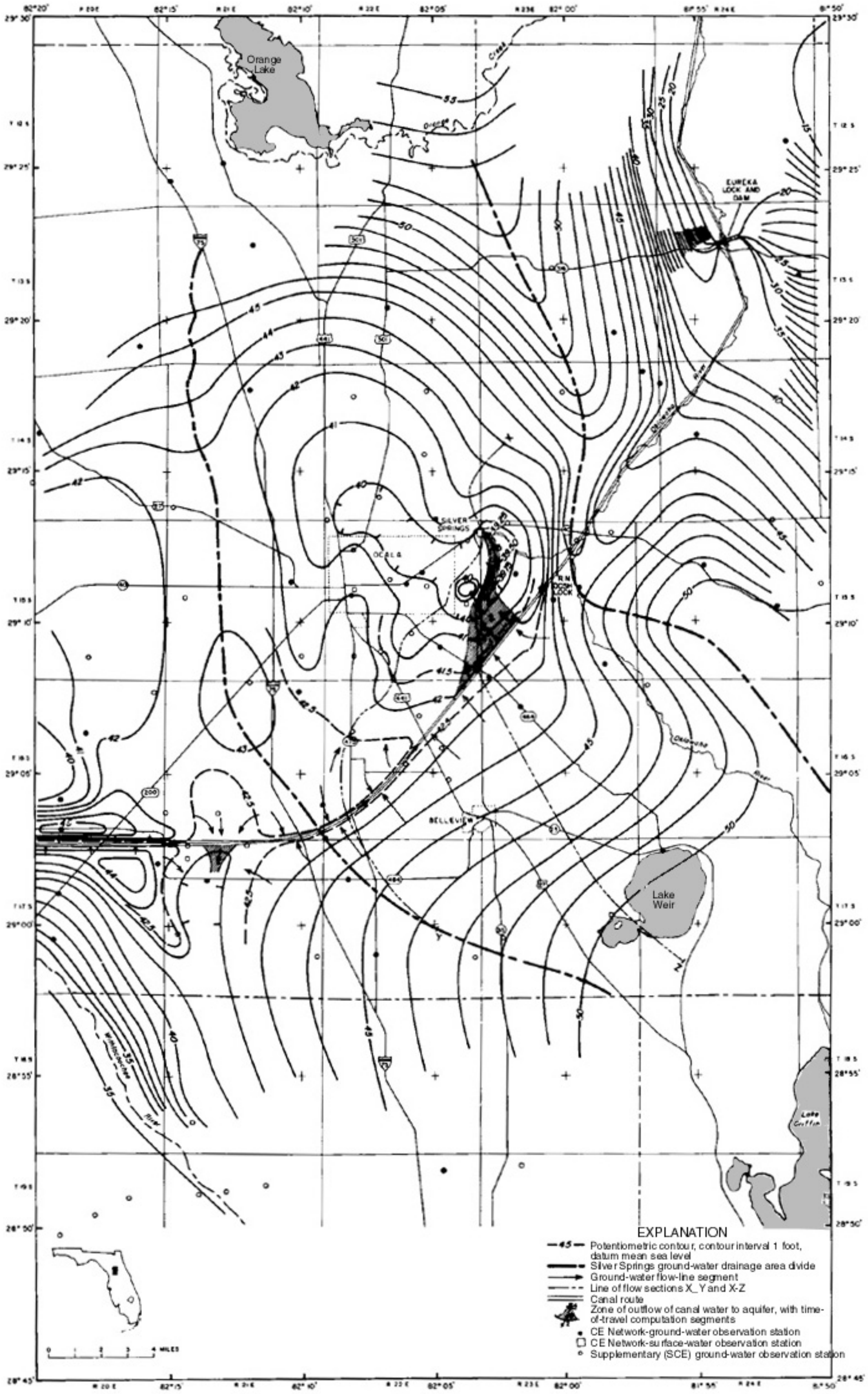


Figure 32. Conceptual model of potentiometric surface of upper part of Floridan Aquifer, Ocala vicinity, had the canal been in May 1968 and had the stage of Eureka Pool been held at 36 feet above mean sea level at the time.

the stage of Summit Pool within a range about 10.5 feet. In contrast, the stage range of record for the Oklawaha River Connor, near the upstream end of Eureka Pool, is 6.26 feet, and at the head of Silver Springs it is 6.49 feet. Also, if it were possible for the Eureka Pool stage to fluctuate as has the stage of the Oklawaha River near Connor during the period of record, it is estimated that the stage of the Summit Pool would have a range of about 12.3 feet.

Shown in the form of line graphs in Figure 33 are the estimated ranges in stage at the head of Silver Springs and in the Summit Pool as governed by the controlled stage in Eureka Pool. Also included in Figure 33 are bar graphs showing the maximum and minimum stages of record at the head of Silver Springs and in the Oklawaha River near Connor, as well as the estimated range of the Summit Pool were it not to be limited by the stage of Eureka Pool. The graphs are based on the long-term stage records, 36 and 22 years, respectively, for the head of Silver Springs and for the Oklawaha River near Connor in addition to considerable ground-water level data including the 36-year record for the artesian well (CE50) near the Sharpes Ferry bridge over the Oklawaha River.

For a given Eureka Pool stage, the estimated range in stage at both the head of Silver Springs and in the Summit Pool may be read from the line graphs. For instance, it may be seen that should the Eureka stage be held at 38 feet throughout the range of record for natural ground-water level fluctuation, the stage at the head of Silver Springs may be expected to range from 41 to 42 feet, while the stage in the Summit Pool should range from 43 to 49.5 feet, provided that lockage losses are replaced regularly by pumpage from the lower pools. Should the stage of Eureka pool instead be deliberately varied between 36 and 40 feet in accordance with natural ground-water level changes, the ranges in stage at Silver Springs and in the Summit Pool will be considerably greater, and it follows that ground-water levels in the area will be allowed to fluctuate more naturally and closer to, natural ranges of record.

At times a high Eureka Pool stage may hold the stage at the head of Silver Springs somewhat higher than it would be under natural conditions. Should this occur, the effect on spring flow will be that the increased hydraulic head at the springs will cause a temporary reduction in the volume of flow as the stage is raised above what it normally would be. This will in turn cause a rise in the water level in the aquifer, a rise which will tend to compensate for the increased head at the springs, and which will ultimately return the spring flow approximately to normal.

Should the stage of Eureka Pool ever be lowered sufficiently to draw the water level at the head of Silver Springs lower than it would be under natural conditions, the reduced stage at the spring head will result in an increased flow rate at the spring. The higher flow rate will tend to lower ground-water levels in the area, a change that will, in turn, cause a reduction of the flow rate approximately to normal.

Inglis Pool.--The Inglis Pool will consist essentially of what presently is called the Withlacoochee backwater or Lake Rousseau, the impoundment on the Withlacoochee River maintained by the old Inglis Dam, although the backwater

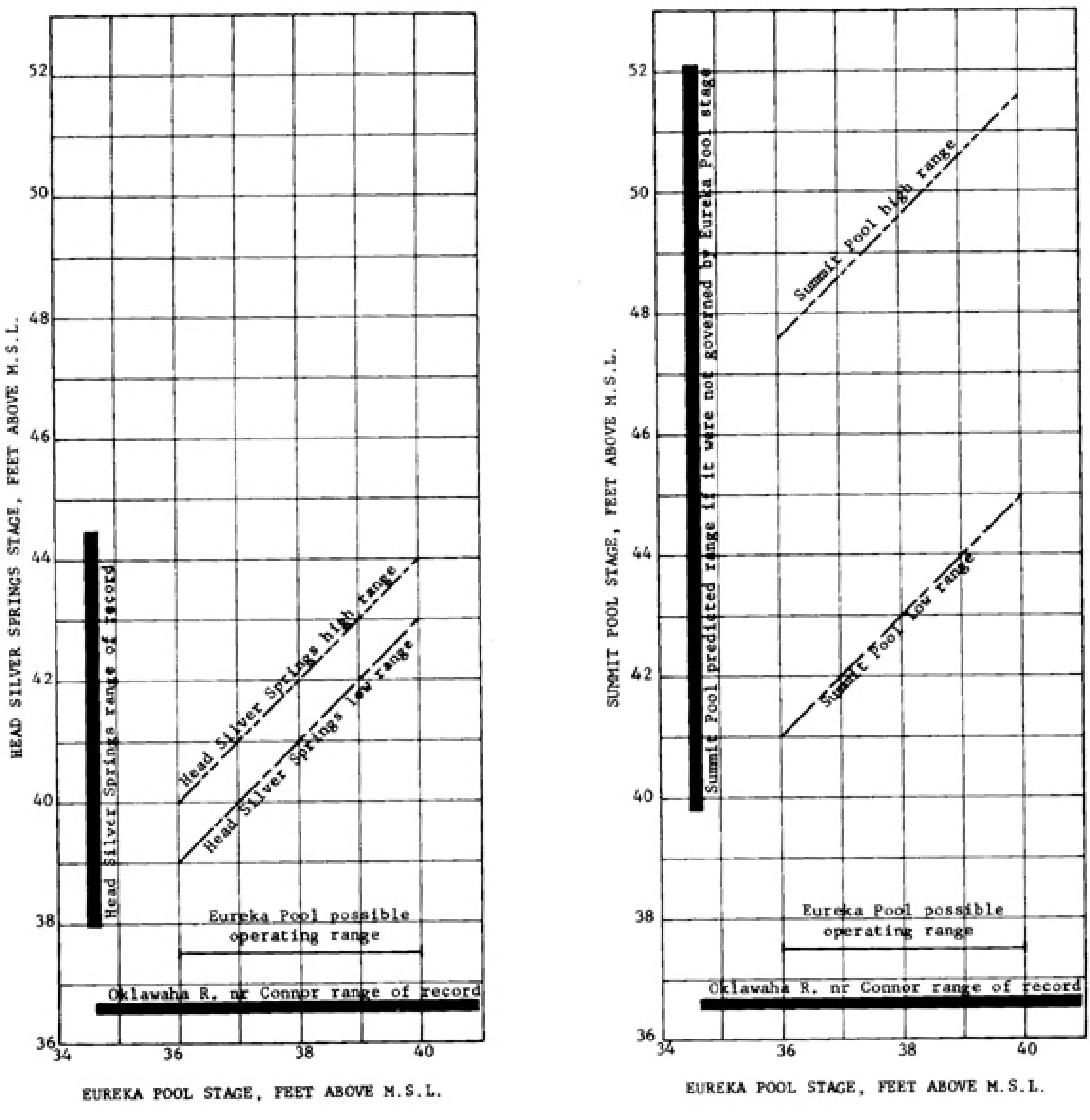


Figure 33. Estimated range in stage at head of Silver Springs and in Summit Pool as governed by Eureka Pool controlled stage.

has been in existence for approximately 55 years, no adverse effects on the ground-water system are known. The canal pool will be operated within about the same stage range as the present impoundment, so there is no reason to expect canal operations to cause significant changes in the ground-water levels. Near the pool area is Rainbow Springs, which will discharge into the Inglis Pool by way of Rainbow River. The head of the springs is located about 4 miles north of the pool, and the water level at the spring head is expected to be 3 or 4 feet higher than the stage of the pool most of the time. Since the Inglis Pool stage will be maintained at comparable levels to those at which the old backwater has been held, no significant changes in stage at Rainbow Springs are expected as a result of canal operations.

Leakage through solution channel systems of the Floridan Aquifer in the vicinity of the old dam has been excessive through the years, although at times it has been temporarily checked by grouting. There is likely to be similar leakage around the new spillway and dam structure which will ultimately replace the old dam. Although the nearby Inglis Lock has a similar geologic and hydrologic setting, leakage may be less of a problem. The considerable length of the lock structure, in comparison to the relatively thin dam structure, somewhat reduces the probability of a well connected set of solution channels of large diameter spanning the full length of the lock. Construction of Inglis Lock is already completed, and during excavation of the lock site, poorly connected and comparatively isolated cavities were found. Dewatering conditions during construction indicated comparatively low transmissivity in the area, although there was some steady ground-water flow from sparsely distributed fissures in the floor of the excavation. If the relatively low transmissivity is partly due to unconsolidated deposits filling some cavities, transmissivity could increase as a result of flushing of the cavities by leakage.

Some direct exchange of water between Inglis Pool and the Floridan Aquifer may occur in a few places where the dredged canal channel is expected to penetrate rocks of the aquifer. However, the potentiometric surface in most of these places is believed close to the planned operating stage for the pool, so the net volume of water exchanged may be small. At the head of Inglis Pool, some ground-water inflow to the canal should occur for a mile or two below Dunnellon Lock, because here the natural potentiometric surface is appreciably higher than the pool's planned operating stage.

The water surface of the already completed canal reach connecting the Inglis Pool with the Gulf slopes very gently to the Gulf, but over most of its length is nearly at the level of the Gulf waters and fluctuates with the tide. The expected moderate lowering of the water table adjacent to the canal occurred, although insufficient data are available (1970) to fully document the change. The natural water table just seaward of the Inglis Lock is a few feet above sea level, and it slopes toward the canal and the coast. The saltwater-fresh-water interface slopes downward in the aquifer laterally away from the seaward reach of the canal, and little salting of the aquifer should occur a short distance from the canal.

The difference in stage of the Gulf reach of the canal and the Inglis Pool will be about 25 feet. This considerable difference in elevation is expected to aid in minimizing the lockage of Gulf salt water into the upper pools of the canal. Over the long term of canal operation, possibility exists for the "locking up" of significant amounts of salt water into the Inglis Pool as a result of salt water and fresh water intermingling during lock filling operations. Although the high stage differential at the lock reduces the probability of high volume movement of salt water into the upper pool, remedies for control of migration should be developed. The subject is discussed in more detail later in the report.

Summit Pool.--The Summit Pool will extend about 28 miles from the R. N. Dosh Lock, in the Ocala-Silver Springs vicinity, southwest to the Dunnellon Lock in the Dunnellon-Rainbow Springs vicinity. The Dosh Lock will be excavated into relatively impervious material comprising the confining layer overlying the Floridan Aquifer. Also, the Summit Pool channel will be bottomed in the confining layer for a distance of a mile and half southwest of the lock. From here to Dunnellon Lock and beyond, the canal will be excavated into rocks of the aquifer to depths ranging from 12 to 27 feet below the water table. Ground-water will flow into the canal channel along the entire length of the pool, and outflow from the pool is expected through one side of the canal in only two comparatively short reaches, one of which is of secondary importance. Interchange of water between the canal and the aquifer will be a natural occurrence directly related to normal movement of water through the aquifer down the potentiometric slope to areas of discharge. An important source of water for the pool will be ground-water inflow, with direct surface runoff a very minor and intermittent source. The stage of the Summit Pool will be partly controlled by the stage in Eureka Pool, but will fluctuate within limits with normal ground-water level changes. Water lost from the Summit Pool and the aquifer by lockage will be replaced with pumpage from the lower pools.

Although natural interchange of water will take place between the pool and the aquifer, the ground-water flow pattern in the upper part of the aquifer will be altered because the canal will function as a flow tube in the upper 12 to 27 feet of the aquifer. The canal will cut across natural shallow flow lines and will, in effect, "short circuit" some of the ground-water flow by diverting water from one zone of preferential flow to another.

Figure 32 is a conceptual model representing the May 1968 potentiometric surface in the Ocala-Silver Springs vicinity had the canal been in existence and had the stage of Eureka Pool been held at 36 feet above mean sea level at that time. Superposed on the model are flow-line segments depicting the plan-view flow pattern as modified by the canal. Geohydrologic sections X-Y and X-Z (fig. 28), drawn along flow lines leading from south to north across the canal to Silver Springs, illustrate the third dimension of the flow pattern. Section X-Y crosses the canal at a point of ground-water inflow; section X-Z crosses at a point of inflow and outflow.

Only two comparatively narrow zones of outflow are indicated on the conceptual model. The zone located about five and a half miles south of Silver Springs, which is considered the more important, would permit flow from the canal toward the springs. The other outflow zone, located to the west between Interstate 75 and State Road 200, may permit minor flow from the canal southwestward toward the Withlacoochee River. Whether or nor significant outflow would actually occur at this secondary location would be, to a large extent, dependent upon the pool stage as determined by outflow conditions at the more important eastern zone of outflow.

Although the model does not include the entire Summit Pool area, examination of the potentiometric surface map showing the entire summit reach (fig. 23) reveals that only these two outflow areas are possible. In the western part of the pool, not shown on the conceptual model, the potentiometric surface is considerably higher than the western zone of ground-water outflow, and therefore eastward flow is anticipated from Dunnellon Lock to this lesser outflow zone, and thence through the aquifer toward the Withlacoochee River. Possibly some water will flow from Dunnellon Lock to the eastern zone of outflow leading toward Silver Springs. Whether such flow is possible will be dependent on several factors, among which are included the effectiveness of the western outflow zone and the relationships between lockage losses at Dunnellon Lock and amounts of ground-water inflow to the canal from the Floridan Aquifer in the Dunnellon Lock vicinity.

From an examination of sections X-Y and X-Z (fig. 28), it is apparent that the canal excavation will have its greatest effect in the shallow part of the aquifer flow system and for some distance below the bottom of the canal (28 feet above mean sea level). In areas where the canal will gain ground water, there will be some inflow through the canal bottom as well as through the sides of the canal (Section X-Y). Once in the canal, water will move toward a point of outflow (fig. 32). At some undetermined depth below the canal bottom, flow will pass beneath the canal on its way to a discharge area without entering the canal. In areas where the canal loses water to the aquifer, the water will flow from the canal through a side and the bottom (Section X-Z). Thus, water flowing beneath the canal towards a discharge area will be forced deeper in the aquifer by the water that is lost from the canal.

Leakage conditions around Dunnellon Lock should be comparable to those at Inglis Lock. The Dunnellon Lock excavation will be below the water table in the upper part of the Avon Park Limestone, where comparatively low aquifer permeability resulting from dolomitization and a partly sand- and clay-filled cavity system is expected.

Test borings drilled on and near the Dosh Lock construction site show that about 110 feet of confining layer rest on the limestone of the Floridan Aquifer. Significant leakage problems are not expected as the lock excavation will penetrate only about 40 feet of the confining layer, leaving about 70 feet between the aquifer and the bottom of the excavation. The artesian head at the lock averages about 45 feet above mean sea level, or about 25 feet above the bottom of the excavation. However, although some upward seepage may be expected, chances appear small for blowouts through the bottom of the excavation or for flotation of the lock, as the remaining overburden below the excavation should be sufficiently thick to counteract the artesian head and maintain a seal over the aquifer. The specific gravity of the saturated silty and sandy clay is estimated at 2.0. On this basis, the 70 feet of confining layer between the aquifer and the bottom of the excavation would provide a counter pressure equivalent to nearly 150 percent of the artesian head.

Quantitative Flow Relationships Between the Summit Pool and Silver Springs Drainage Area

General.--When the canal is completed, a dynamic inflow-outflow relationship will exist between the Summit Pool and the Floridan Aquifer. An important source of water in the Summit Pool will be ground- water inflow from the Floridan Aquifer in potentiometrically high areas, which will be balanced by outflow into the aquifer in potentiometrically low areas. The water surface of the canal will slope slightly toward zones of outflow, and although the total length of inflow reaches is expected to exceed the total length of outflow reaches by several times, the pool stage is expected to reach equilibrium at a level closer to the lowest level on the natural potentiometric surface intersected by the Summit Pool than to the highest level. This is because of the appreciably higher average transmissivity of the aquifer in the zone of outflow.

In addition to outflow to the aquifer, water will be lost from the Summit Pool by lockages, evaporation and leakage, lockage being the most important. Average daily evaporation losses have been estimated at about one percent of maximum daily lockage losses. No estimate has been made of possible losses due to leakage around locks. The design of the canal provides for pumping facilities below Dosh Lock and possibly below Dunnellon Lock which can replace losses from the Summit Pool with water from Eureka Pool and (or) Inglis Pool.

It is important for several reasons that reliable estimates be made of the quantities and rates of movement of water in and out of the aquifer along the Summit Pool under a given set of conditions. These estimates can be useful for, among other things, estimating pumpage requirements needed to avoid excessive drawdown at any time by lockages, determining the time it would take water to flow through the aquifer from the canal to the spring, and predicting the stage of the pool at a given time.

The transmissivity of the aquifer must be known in order to make the foregoing determinations. As discussed earlier, computed average transmissivity of the aquifer along various closed potentiometric contours surrounding Silver Springs (fig. 27) varied from 10 mgd/ft at the 45-foot contour to 35 mgd/ft at the 43-foot contour. Although these values appear reasonable enough as averages for the area, they are too generalized to permit specific analysis of the ground-water flow relationships between the aquifer and the Summit Pool. Therefore, flow-net analysis was used by the writer for computing a first approximation of the variation in transmissivity from one aquifer flow tube to another in the closed ground-water drainage area of Silver Springs. As may be seen, it was possible through application of flow-net analyses to compare differences in transmissivity within the drainage area, and thereby further distinguish the major zones of flow to the springs. The result was a more accurate evaluation of the relationships of the Summit Pool to the total Silver Springs flow system. Also, it has been possible to make useful estimates of inflow volumes to the aquifer, and thereby make predictions of various parameters including stage in the pool under a given set of ground-water level conditions, stage duration, and rates of ground-water flow from points of outflow from the pool to points of natural discharge such as Silver Springs.

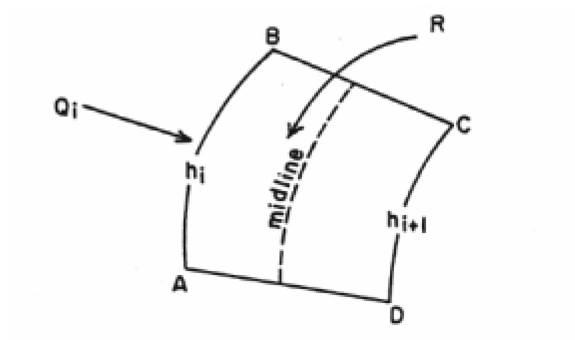
Flow-net analysis.--In the method of flow-net analysis used, it is assumed that flow in the aquifer is lateral (horizontal), the component of vertical flow being negligible. Flow in the leaky beds overlying the aquifer is considered vertical, and when the water gets to the aquifer the direction of flow abruptly changes to lateral. Flow lines in the aquifer in the Silver and Rainbow Springs drainage areas tend to be nearly horizontal except at the discharge point where flow lines are vertical. The gradient of the potentiometric metric surface is so low in most places that it is nearly horizontal. Further, the aquifer in the study area is considered heterogeneous and isotropic with respect to its permeability; that is, the method assumes that the permeability differs from one point to another given point in the aquifer, permeability is the same in all horizontal directions. From descriptions presented of the quantitative aquifer characteristics, it is reasonable to consider that the Ocala Limestone part (about 100 feet thick) of the aquifer is heterogeneous and isotropic.

In a heterogeneous and isotropic aquifer the equipotential and stream surfaces are orthogonal (that is, equipotential lines and flow lines intersecting at right angles). In the analyses made the aquifer is assumed artesian and flow steady. It is further assumed that the blah of equipotential and stream surfaces holds true when the aquifer receives recharge vertically. Also, as the variation in saturated thickness of the aquifer is small compared to the estimated total thickness where the aquifer is under water-table conditions, the assumption of an artesian condition should in large errors.

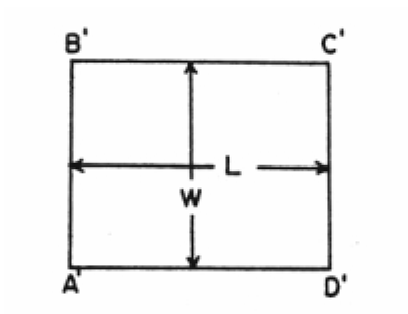
Silver Springs is considered as a well which fully penetrates the Ocala Limestone and it is assumed that the discharge from the springs is derived entirely from this part of the aquifer (100± feet thick). By drawing flow lines perpendicular to the equipotential lines from the potentiometric divide to Silver Springs, a pattern of distorted rectangles is produced, each bounded by two flow lines and two equipotential lines. Then, by use of the law of continuity and Darcy's law, an estimate of the average transmissivity in each rectangle is obtained.

In order to calculate transmissivity, it is first necessary to estimate how much water flows across the middle of the rectangular area in question. This flow equals the sum of the recharge in all the rectangles bounded by the two flow lines, the potentiometric divide, and the upstream end of the subject rectangle plus the recharge in the upstream half of the subject rectangle. Average recharge to the spring drainage area is considered equal to the average discharge of the spring, as other withdrawal from the aquifer is negligible. With the aid of geohydrologic maps showing variations in thickness and type of cover over the aquifer, estimates may be made of the areal variation in recharge.

The method is illustrated by the following diagram:



where BC and AD are flow lines, and AB and CD are equipotential lines. Q_i is the flow into the region ABCD, R is the vertical recharge per unit area, and h_i and h_{i+1} are potential values. Rather than the distorted rectangle ABCD, it is convenient to consider a rectangular approximation A'B'C'D', where A'D' and B'C' are flow lines representing AD and BC, and A'B' and C'D' are equipotential lines having the same values as AB and CD respectively:



This transformed flow element has the properties where

A'B' = C'D' = 1/2 (AB + CD) = W, and

A'D' = B'C' = 1/2 (AD + BC) = L.

With all other values known, transmissivity, T, may now be calculated for any given locality within the area ABCD by use of the following equation:

$$T = \frac{(L(Q_i + (1/2)WLR))}{(W(h_i + 1 - h_i))}$$

The closer the shape of ABCD to a rectangle, the more accurate will be the determination of transmissivity. Another factor limiting the accuracy of the method is the difficulty in estimating recharge within the individual rectangles. Average recharge for the entire drainage area may be readily calculated, but that for the individual rectangles must be estimated on the basis of the amount and types of surficial materials and other factors which control recharge. With care, however, the recharge distribution in terms of percentages of the whole can be estimated with acceptable accuracy. From this it is possible to estimate the recharge in each rectangle.

Because steady flow is assumed in the flow-net method, averages of the potentiometric gradient and rates of recharge should be used. The mean discharge of Silver Springs was used to determine the average recharge. However, a map of the average potentiometric surface is not available, and the map of the more gently sloping potentiometric surface in May 1968, was used. The discharge of Silver Springs in mid-May 1968, was about 80 percent of the mean discharge for the period of record. Therefore the calculated transmissivities for each flow cell were adjusted by the coefficient 0.8. Theoretically, had a potentiometric surface representative of the average spring discharge been used rather than the May 1968 potentiometric surface, the calculated transmissivities would equal the adjusted values.

Figure 34 is a map of the May 1968 potentiometric surface in the Ocala vicinity, on which have been drawn 25 flow tubes converging from all directions on the discharge area at Silver Springs. Average transmissivity within one rectangle or flow cell in each of the 25 flow tubes has been calculated. The cells either abut or are diagonally connected in a continuous belt around the springs, and they are positioned on the potentiometric surface at altitudes of 1 to 4 feet upgradient from the springs.

All flow from upgradient of the cells must pass through the belt to reach the springs. Thus, the sum of the flow through all cells of the belt is equal to the flow from the springs less the recharge downgradient from the belt.

Also shown on Figure 34 is the distribution of the Miocene and younger beds which form the confining layer in the Oklawaha River valley, and restrict recharge in places west of Silver Springs. This information aids in estimating the recharge to the individual flow tubes.

On figure 34 the flow cells for which transmissivity is calculated are numbered clockwise 1 through 25. The computed transmissivity and the data used to compute it are listed in Table 2 for each of the flow cells.

Transmissivity ranges from 80,000 gpd (gallons per day) per foot (10,700 ft^2/day) in cell-1, which is located under the confining layer east of the springs, to 190,000,000 gpd/ft (25,500,000 ft^2/day) in cell-15, through which flow passes from a major part of the north half of the Silver Springs drainage area. The average transmissivity along the belt of cells is 15,600,000 gpd/ft (2,090,000 ft^2/day). This compares favorably with the 14,700,000 gpd/ft (1,970,000 ft^2/day) average determined along the 41.5-foot potential contour on the May 1968 map by the method of analysis based on the total flow across a closed contour. The 41.5-foot contour passes through the centers of 19 of the 25 flow cells.

The general accuracy of the transmissivity values was checked by comparing calculated flow through the aquifer with flow at Silver Springs. Cumulative daily flow calculated from the transmissivities of each of the flow cells, plus the estimated daily recharge in the area downgradient from the belt of cells, differs less than three percent from the daily discharge of Silver Springs.

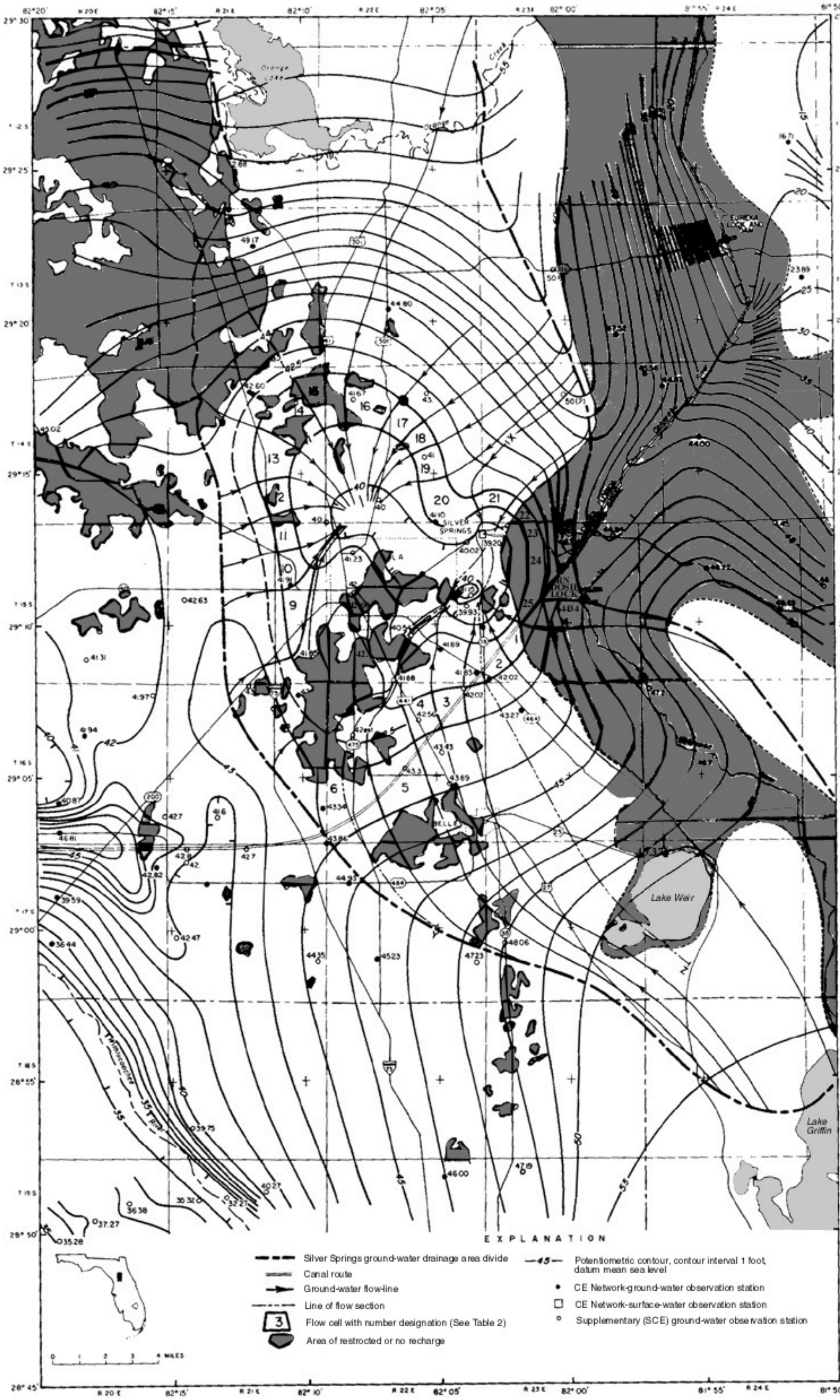


Figure 34. Flow-net, upper part of Floridan Aquifer in May 1968, Ocala vicinity.

Table 2.--Flow-net analysis data, upper part of Floridan aquifer, Ocala vicinity.

Cell No.	Length L - miles	Width W - miles	Inflow area - miles^2	Aver. annual recharge inches	Inflow Q - mgd*	Recharge R to cell - mgd/mi^2	WLR - mgd	Q&WLR - mgd	h_i+1^{-h}i - feet	T_1 ** mgd/ft	Percent flow of Silver Springs	Trans. T Adj. w/coef. 0.8*** mgd/ft
1	1.0	1.4	94.9	9.0	40.7	0.686	0.96	41.62	1	29.2	7.8	23.4
2	1.4	1.5	82.4	15.0	58.8	0.857	1.80	60.60	1	54.9.	11.4	43.9
3	1.4	1.1	14.5	16.0	11.0	0.857	1.32	12.32	1	14.8	2.3	11.8
4	1.4	1.2	14.7	16.0	11.2	0.857	1.49	12.69	1	14.3	2.4	11.4
5	2.0	2.0	14.9	15.0	10.6	0.714	2.86	13.46	1	12.0	2.5	9.6
6	2.6	2.1	8.0	15.0	5.7	0.714	3.90	9.60	1	9.5	1.8	7.6
7	1.0	1.2	4.0	17.0	3.2	0.381	0.48	3.68	0.5	5.8	0.7	4.6
8	2.2	2.1	4.2	14.0	2.8	0.714	3.34	6.14	1	4.8	1.2	3.8
9	2.7	1.4	6.6	18.0	5.7	0.857	3.26	8.96	1	13.8	1.7	11.0
10	1.3	1.2	2.7	18.0	2.3	0.857	1.26	3.56	1	3.3	0.7	2.6
11	1.0	1.4	2.5	18.0	2.2	0.714	0.96	3.16	1	1.8	0.6	1.4
12	1.1	1.3	2.1	18.0	1.8	0.714	0.99	2.79	1	1.9	0.5	1.5
13	1.5	1.6	6.4	16.0	4.9	0.809	1.93	6.83	1	5.2	1.3	4.2
14	1.8	1.7	17.7	12.0	10.1	0.571	1.78	11.88	1	12.1	2.2	9.7
15	1.9	1.4	247.0	15.0	176.4	0.714	1.93	178.33	1	237.0	33.6	189.6
16	1.6	1.8	38.6	20.0	36.8	0.952	2.78	39.58	1	34.4	7.5	27.5
17	1.3	1.1	11.2	18.0	9.5	0.809	1.12	10.62	1	11.9	2.0	9.5
18	1.1	1.0	7.4	18.0	6.4	0.857	0.94	7.34	1	7.5	1.4	6.0
19	1.0	1.0	5.7	18.0	4.9	0.857	0.89	5.79	1	5.3	1.1	4.2
20	0.8	2.7	3.3	18.0	2.8	0.857	2.23	5.03	1	1.2	0.9	1.0
21	0.6	1.3	7.5	18.0	6.4	0.857	0.67	7.07	1	3.3	1.3	2.6
22	0.6	1.2	4.7	5.0	1.1	0.286	0.20	1.30	1	0.6	0.2	0.5
23	0.5	1.2	1.6	2.0	0.2	0.048	0.03	0.23	1	0.1	0.04	0.08
24	0.6	1.3	3.6	5.0	0.8	0.095	0.08	0.88	1	0.4	0.2	0.3
25	0.8	1.1	5.0	8.0	1.9	0.143	0.12	2.04	1	1.4	0.4	1.1
92 mi^2 -area below cells				15.0			65.70	65.70			12.3	
Totals								521.20			98.04	Av.15.6

***million gallons per day.**

****T_1 computed from the equation below; unadjusted for the nonaverage potentiometric surface.**

$$T = \frac{L(Q + (1/2)WLR)}{W^h i + 1^{-h} i}$$

*****Coefficient 0.8 corrects T_1 to T, as Silver Springs discharge in May 1968 was only 80% of average.**

Since the primary purpose in determining areal variations in transmissivity is to evaluate the ground-water inflow-outflow relationships in the Summit Pool reach, which passes through the Silver Springs drainage area, cells 1 through 6 are the most important of the 25 cells, as the canal will be excavated through them. However, the complete belt of cells surrounding Silver Springs is an important indicator of the accuracy of estimates and calculations made in any part of the flow-net. Also, the belt of cells is a valuable tool for determining the contribution to the total spring flow made by any given part of the drainage area.

Static water level in Summit Pool.--With both locks closed and the water level stabilized in the Summit Pool, ground-water inflow and outflow will be equal. The static water level will then depend upon, among other things, the variation in transmissivity of the aquifer along the reach. The variation in transmissivity will also control the length of the zone or zones of ground-water outflow. As illustrated on the conceptual model in Figure 32, it is assumed that slight potentiometric divide will continue to exist across and nearly normal to the canal near the present southwestern boundary of the Silver Springs drainage area. There will be a saddle in the potentiometric divide at the canal because of drawdown resulting from inflow into the canal. The water surface of the canal will tend to slope very gently in both directions from the divide. However, it is probable that the divide in the canal's static water surface will be shifted far to the west of the natural ground-water divide, or clear to Dunnellon Lock, as most of the water lost to the aquifer from the Summit Pool will be in the primary zone of outflow toward Silver Springs.

Flow in the canal will tend toward zones of ground-water outflow where the water level of the canal will be slightly higher than the potentiometric surface in the adjacent part of the aquifer. Actually, some inflow will occur from the upgradient side of the canal in these areas, but outflow will occur also through the downgradient side and through the bottom of the canal. Elsewhere in the Summit Pool, only ground-water inflow will occur. The distinction between the inflow and outflow zones on Figure 32 is emphasized by flow-direction arrows plotted on the model and by shading in the outflow zones.

During locking operations and periods when lockage losses are being returned to Summit Pool, the direction or directions of flow within the pool may of course be quite different from those that will prevail under the state of equilibrium discussed above. For instance, frequent operation of Dunnellon Lock and (or) large volume pumping from Eureka Pool can reverse the direction of flow in Summit Pool and result in the drainage of appreciable volumes of Eureka Pool water into Inglis Pool.

An approximation of the boundaries of the Summit Pool's primary zone of outflow, centered about five miles south of Silver Springs (fig. 32), and a prediction of the approximate static water level in the pool for a given natural ground-water level condition was made by a method of trial and error through the aid of a form of the Darcy equation expressed as follows:

$$Q = TIW$$

where Q = inflow or outflow, gpd

T = transmissivity, gpd/ft

H = hydraulic gradient, feet

W = length of inflow or outflow reach, feet

Computation of the actual volume of inflow, and thus the actual outflow, is difficult because of limitations in the capability for estimating the effective depth of influence that the canal will have on the flow system of the aquifer. However, it is known that when the canal water level is static, inflow and outflow will be equal. If the water level required to produce a volume of outflow equal to the volume of inflow is known, the static level in the pool is therefore known. Also, once static water level is known and the potentiometric surface is adjusted in the manner illustrated by the conceptual model (fig. 32), the boundaries and, therefore, the length of the outflow reach, are known because outflow will occur at all points where the stage of the pool is higher than the potentiometric surface. Conversely, inflow will occur at all points where the canal stage is lower than the potentiometric surface.

From the foregoing, it is seen that through a process of trial and error in estimating static water level in the canal, and by using transmissivity values approximated through flow-net analysis, the above equation may be used to estimate equal inflow and outflow. The static water level, and therefore, the length of the outflow reach, may be adjusted until inflow and outflow balance. Since, in the method used, the canal is treated as a fully penetrating well into which flow occurs from the full effective thickness of the aquifer, the quantity of inflow or outflow determined is not its actual volume of inflow or outflow. The reason is that the canal is really only partially penetrating, and only a part of the total flow in the aquifer enters the canal. The rest passes under the canal on its way to the discharge area. A crude approximation of the actual volume of inflow or outflow may be determined by estimating the percentage of the calculated quantity that enters or leaves the canal.

Inflow and outflow were first calculated on the May 1968 potentiometric surface for an assumed canal stage of 42.0 feet at the zone of northward outflow toward Silver Springs. It was found that inflow was 126 percent of outflow. When calculated for an assumed stage of 42.2 feet, outflow exceeded inflow by 25 percent. Therefore, the stage of equilibrium was estimated to be 42.1 feet. This means that the maximum rise above the natural potentiometric surface in the Summit Pool would have been about 0.9 foot had the canal been in existence in May 1968. The slope in the canal surface from inflow zone down to outflow zone should be very slight; therefore, the decline in the potentiometric surface where the canal crosses the western divide of the Silver Springs drainage area should have been about 1.6 feet in May 1968. The ratio of rise to fall then is about 1 to 2 or the decline in the potentiometric divide area is nearly 2 times the rise in the lowest area.

The total length of the reach of the Summit Pool from the west edge of the Miocene confining layer to the southwest edge of the drainage area is about 11.5 miles. As shown on Figure 33, it has been determined that the zone of northward outflow extends a distance of about 4 miles southwest from the western limit of the Miocene confining layer. Straight-line distance of the outflow zone from Silver Springs ranges from about 4 to nearly 6.5 miles.

With a change in natural ground-water levels, the length of boundaries of the outflow reach should not change greatly, as the only appreciable influence should be the slight changes in slope of the potentiometric surface as it rises and falls. The static water level in the canal should differ from natural ground-water levels at about the same 1:2 ratio determined for the May 1968 water levels; that is, the maximum rise should be about one third the difference in the altitude, at the canal centerline, between the natural ground-water level in the potentiometric trough of the outflow area and the natural level on the potentiometric ridge at the southwest edge of the Silver Springs drainage area. Therefore, even if the stage of the Summit Pool were independent of the controlled Eureka Pool stage, minimum water levels at observation wells near the natural potentiometric trough, in the zone of outflow, could be several tenths of a foot below the canal's design minimum water level of 40 feet above mean sea level, but the minimum static water level in the canal would be no lower than 39.8 feet. Again disregarding the influence of Eureka Pool, the maximum ground-water levels predicted from long-term records indicate a maximum static water level of 52.1 feet above mean sea level for Summit Pool, or 2.9 feet below the design maximum of 55 feet. However, as is shown in Figure 33, with the stage of Eureka Pool controlled between 36 and 40 feet above mean sea level, the stage of the Summit Pool will be indirectly restricted through Silver Springs to a range of about 10.5 feet, or from about 41 to 51.5 feet above mean sea level.

Shown in Figure 35 are four estimated stage duration curves for the Summit Pool. They are based on results of the flow-net analysis, on a straight-line relationship among the 3-year records of the Summit Pool area observation wells and the 36-year record at the artesian well near Sharpes Ferry bridge on the Oklawaha River, and on the predicted relationship between the Eureka Pool and Summit Pool stages. The solid curve represents the estimated stage duration in the Summit Pool if the stage could fluctuate independently of the controlled Eureka Pool water level. If such a situation could exist, the stage of Summit Pool should be at or above 45.5 feet 50 percent of the time. Since the Eureka Pool stage will act as a partial control on the Summit Pool stage, it will not be possible for the Summit Pool to fluctuate naturally through the range indicated by the solid curve. Therefore, in addition to the solid curve, three dashed curves for Summit Pool stage duration have been estimated and plotted in Figure 35 based on each of three given controlled stages (36, 38, and 40 feet) in Eureka Pool. These curves indicate that 50 percent of the time, depending upon the particular stage held in Eureka Pool, the stage in Summit Pool will equal or exceed a level ranging from about 44 feet to about 48 feet above mean sea level. The above figures assume a stabilized water level in the pool at a time when

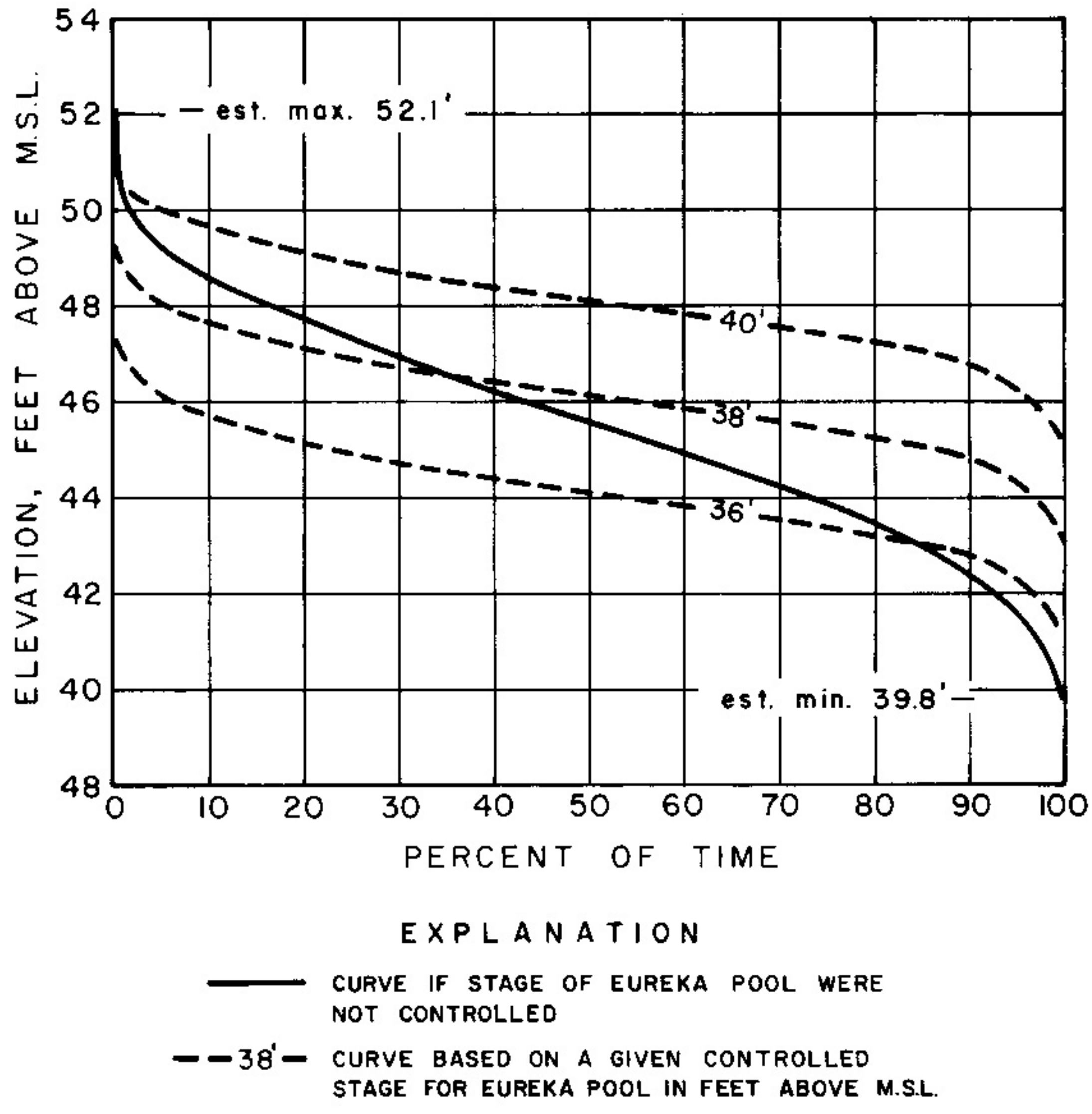


Figure 35. Estimated stage duration curves for Summit Pool.

ground-water inflow and outflow are equal and losses due to lockage, leakage at locks, and evaporation are replaced by pumpage from lower pools.

Obviously, if the stage in Eureka Pool acts as a partial control on the Summit Pool water level through the aquifer by way of Silver Springs, the Eureka Pool stage will also exert some control on the altitude and shape of the potentiometric surface in the Summit Pool-Silver Springs area. Therefore, in order for canal operations to alter the natural ground-water regimen in the area as little as possible, Eureka Pool should be operated throughout its 4-foot stage range (36 to 40 feet above mean sea level) in a manner properly coordinated with natural fluctuations in ground-water levels.

A predictable static water level in the Summit Pool has been essential to construction of the conceptual model in Figure 32. Knowledge of the canal stage permits good control for predicting the configuration of the potentiometric surface close to the pool. The effect of the canal on the potentiometric surface diminishes with distance from the canal, and eventually the effect becomes negligible, although it theoretically extends all the way to the aquifer boundaries.

As shown by the model, water levels will be drawn down on both sides of the canal where only ground-water inflow occurs and raised on both sides of the canal where outflow occurs, although outflow occurs from only the downgradient side and from the bottom of the canal.

Ground-water flow rate, Summit Pool to Silver Springs.--A knowledge of the time it will take for water to move through the aquifer from the zone of outflow in the Summit Pool to discharge at Silver Springs would be very important should the Summit Pool become contaminated. If contaminants infiltrate the aquifer from the Summit Pool, knowledge of time of travel would of course, be an important factor if it were decided to lower the stage of the pool to back-flush the aquifer. Also, knowledge of the velocity of flow in the aquifer would be necessary in order to forewarn downgradient users when to expect the arrival of contaminants. Reliable estimates of velocity are useful in determining the probably lateral distribution of various contaminants in the aquifer from a point of inflow. The slower the water moves the more limited the spread of contamination, in general, because time is allowed for purification factors such as settling, adsorption, and absorption to do their work. If movement is rapid, it is necessary to take special precautions to prevent contamination-precautions not always so important if movement is slow.

Various fluorescent dye and radioactive tracer methods for measuring rates of flow have been investigated. Such methods, if workable, are the most accurate, because they offer a means for direct measurement. However, high dispersion and adsorption rates of most dyes discourage their use in the area for tracing flow over distances much greater than a mile or so (Knochenmus, 1967).

The hazards of introducing radioactive tracers discourages their use. An effort has been made to utilize variations in the concentration of natural tritium as a tracer. Although the tritium studies have yielded some interesting information pertaining to variations in recharge in the study area, it has been found that these variations tend to complicate and make impracticable the use of natural tritium for time-of-travel measurements.

In the absence of a practical means for direct measurement, an approximation of time of travel can be inferred from the slope of the potentiometric surface and from the transmissivity.

Assuming that the flow is laminar one can calculate the velocity from the equation:

$$V = \frac{PI}{7.48a}$$

where:

V = velocity, in feet per day (fpd)

P = permeability, in gpd/square ft.

I = hydraulic gradient, in ft/ft

a = aquifer porosity, percent

The factor 7.48 converts gallons per day per square foot to feet per day.

The zone of outflow indicated in Figure 32 is in the area represented by flow cells 1 and 2 of Figure 34. The average transmissivity across each of the two cells is 23,000,000 and 44,000,000 gpd/ft (3,080,000 and 5,900,000 ft^2/day, table 2), which average to 34,000,000 gpd/ft (4,560,000 ft^2/day) along the zone of outflow. According to Figure 32 the tube of outflow narrows toward Silver Springs to about one-eighth its width at the canal, and the potentiometric gradient near the spring end of the tube is several times less than the gradient at the canal end. Also, more water passes through the lower part of the tube because of recharge between the canal and the springs. These factors indicate increasingly higher transmissivity along the route of flow from the canal to the spring. Therefore, if the effective thickness of the aquifer is assumed to be 100 feet and the average effective porosity to be 30 percent all the way from the canal to Silver Springs, time of travel of water in the aquifer will vary with changes in transmissivity and the gradient of the potentiometric surface. Since transmissivity increases in the direction of the spring, time of travel will, in accordance with the above equation, tend to decrease in the same direction. Conversely, the reduced gradient of the potentiometric surface in the direction of the spring will tend to slow the rate of flow. A rough approximation of the time of travel from the Summit Pool to Silver Springs, through the central part of the flow zone, has been made by totaling five increments of travel time, each calculated for one of five segments of a flow line reaching from the canal to the spring. In Figure 32 the flow line is shown extending through the flow zone from the 42.00- to the 39.25-foot equipotential line, and divided into five segments by the 41.00-, 40.00-, 39.75-, and 39.50-foot equipotential lines. From the flow net analyses, permeability values of 0.55, 0.78, 5.7, 30, and 28 mgd/ft^2 were derived respectively for segments 1 through 5 of the flow line, and the following estimated increments of time were calculated individually by substitution in the foregoing equation:

Segment - 1: about 40 days
Segment - 2: about 40 days
Segment - 3: about 30 days
Segment - 4: about 15 days
Segment - 5: about 15 days

The total travel time through all five segments from Summit Pool to Silver Springs is, therefore, about 140 days.

With a travel time of about 140 days for water to move the nearly 5 miles through the central part of the ground-water flow zone from the Summit Pool to Silver Springs, the estimated average rate of flow is close to 200 feet per day. The nearest point of outflow to the springs is about 4 miles, and the farthest is about 6½ miles. Thus, at the average velocity, flow to the spring from the closest point would require about 100 days, while about 170 days would be required for the trip from the farthest point of outflow.

Considerable variation in actual velocity may occur along the flow route because of variations in effective porosity and effective thickness of aquifer, among other factors. Velocity may increase and decrease frequently. In fact, though transmissivity must increase toward the springs, is it possible that effective porosity may increase sufficiently to cause velocities to be slower near the spring area than near the canal. Also with other factors remaining constant the flow velocity will vary at a rate inversely proportional to the thickness of the flow zone. For instance, if most of the flow were limited to a solution channel system 20 feet thick instead of a flow zone 100 feet thick, the permeability would be about five times greater than needed the flow rate estimated above, and the average velocity from the canal to Silver Springs would be about 1,000 feet instead of 200 feet per day.

Studies made by Knochenmus (1967) in an area not far south of the Summit Pool indicate that ground water carrying fluorescent dye traveled from Ocala Caverns to Wolf Sink, nearly 7,000 feet northward, at a possible average velocity of 1,150 feet per day. At this average sustained velocity water would move from the Summit Pool to Silver Springs in about 20 days.

The calculated velocity of 200 feet per day is an approximation, and is subject to large variation caused by many factors, one of the most important being how well connected is the solution channel system. Once flow reaches solution channels that are open directly to the spring orifices, ground-water flow velocities may approach surface stream velocities. However, 200 feet per day is an unusually high velocity for ground-water flow over more than short distances, and for purposes of this investigation it is considered reasonable. Ordinarily, ground-water flow rates of more than a few feet per day are considered exceptional.

Effect of lockage on Summit Pool.--If the Summit Pool fully penetrated the effective thickness (assumed 100± feet) of the aquifer, it would intercept northward flow from nearly one third of the Silver Springs drainage area. The intercepted flow, under average conditions, would be about 180,000,000 gpd. If both locks remained closed and the stage of the pool were static, inflow from and outflow to the aquifer would balance so that flow to the springs would only be deflected, and not prevented from moving through the aquifer to the springs. Since the canal will on the average penetrate less than one fifth of the effective thickness of the aquifer, it is reasonable, considering some vertical inflow through the canal bottom, to expect the canal to actually intercept only about one fourth of the total flow moving through the canal's line of section (fig. 28). If such is the case, under average conditions only about 45,000,000 gpd of ground water will pass through the canal channel on its way to discharge at Silver Springs. This is equivalent to about eight percent of the average discharge of the springs.

Design operations of the canal indicate a maximum of 27 lockages per day are possible at each lock. Thus, 54 rockfalls of water could be drained from the Summit Pool in 24 hours. The volume of water drained from the pool through the locks during maximum operation exceeds the volume of ground-water inflow to the pool under conditions of equilibrium, but lockage losses will be replaced by pumping water into the Summit Pool from lower pools. The volume of water moved through the locks, and therefore the amount of pumpage required, will depend not only on the number of lockages, but also on the stages of the Summit Pool and the two adjoining pools.

Each lock has vertical sides and is 600 feet long by 84 feet wide. The canal channel of the Summit Pool reach is 28 miles long, has sides which slope 1 foot per 3 feet of width, a flat bottom 150 feet wide, and a design water depth variation of 12 to 27 feet, depending on natural ground-water levels.

Under design maximum water-level conditions, 5,600,000 gallons will be required to fill Dosh Lock and 10,000,000 gallons to fill Dunnellon Lock, or a total of 15,600,000 gallons for one complete passage through the Summit Pool. This is equivalent to about 1.9 percent of the maximum daily flow of record at Silver Springs (834,000,000 gallons).

Under design minimum water-level conditions 750,000 gallons would be required to fill Dosh Lock, and 6,000,000 gallons would be needed for Dunnellon Lock, or 6,750,000 gallons for one complete passage. This equals about 1.9 percent of the minimum daily flow of record for Silver Springs (348,000,000 gallons).

The maximum 27 complete passages per day under maximum water-level conditions would drain 421,000,000 gallons from the Summit Pool, a volume equivalent to about 50 percent of the maximum daily flow of record from Silver Springs. The maximum 27 passages under minimum water-level conditions would require 182,000,000 gallons, or slightly more than 50 percent of the minimum daily discharge of record at Silver Springs.

If no ground-water inflow or outflow and no evaporation is assumed, lockage loss from 27 complete passages during maximum water level conditions would lower the stage of the Summit Pool 1.2 feet. Under minimum water-level conditions, the 27 passages would lower it about 0.7 foot. The above lockage water volume data is listed in Table 3 for ease of reference.

Table 3.--Lockage-water volume data, Summit Pool.

	Maximum Water Level	Minimum Water Level
Dosh Lock volume, gallons	5,600,000	750,000
Dunnellon Lock volume, gallons	10,000,000	6,000,000
Total volume both locks, gallons	15,600,000	6,750,000
Volume for 27 complete passages, gallons	421,000,000	182,000,000
Water level decline in pool from 27 passages, feet	1.2 ft	0.7 ft

In summary, the maximum 27 passages will, under any natural water-level condition, require a volume of water equivalent to about 50 percent of the daily flow of Silver Springs. Water discharged through the locks is originally derived from inflow to the Summit Pool from the aquifer. This, of course, does not mean that the lockage would reduce the flow at the springs by 50 percent. Only a part of the water needed for lockages would be diverted directly from ground water moving toward the springs to be discharged. Inflow to the pool from the Silver Springs drainage area is estimated to be equivalent to only about eight percent of the daily flow of Silver Springs. The inflow from outside the Silver Springs drainage area, that is, between the west edge of the drainage area and Dunnellon Lock, is expected to be much less.

The inflow from the aquifer is not sufficient to maintain the necessary supply for maximum canal operations. With no replenishment of water from the lower pools, stage of the Summit Pool could be drawn down below normal operating levels, and thus alter the natural ground-water flow system significantly for excessive distances from the pool. Drawdown would eventually extend to Silver Springs and cause complications in flow rate of the springs and in the stage relationship between the springs and Eureka Pool. Therefore, in order to keep alterations to the natural ground-water regime at a minimum, all lockage losses from the Summit Pool must be replaced by pumping water from lower pools. By disregarding any evaporation losses from the canal and possible leakage around locks, the net effect of pumping water from the lower pools then would be that ground-water inflow to the Summit Pool would equal outflow toward natural ground-water discharge areas, and the discharge at Silver Springs would be unaffected.

At the rate of about 51 inches per year evaporation loss from open water surfaces in the area, average daily evaporation from the Summit Pool under maximum water-level conditions should equal about 4,000,000 gallons. However, the 53-inch average annual direct rainfall on the canal surface will offset most of the evaporation loss, because all but about 15 inches of the annual rainfall would have been lost to evapotranspiration anyway, were the canal not in existence.

Pollution control in the Summit Pool.--If the water level in the Summit Pool becomes nearly stabilized in accordance with natural ground-water-level conditions, there will be a state of equilibrium between ground-water inflow to and ground-water outflow from the pool. During a state of equilibrium, contaminants if mixed with or dissolved in the canal water will move toward a zone of outflow and enter the aquifer with the water. Therefore, it is of prime importance that every effort be made to avoid contaminating the waters of the Summit Pool.

Any normal use of the canal by barges and boats will tend to cause at least minor dirtying of the water, but the natural filtering capacity of the aquifer immediately adjacent to the canal can minimize movement of particulate contaminants into the aquifer, provided large caverns do not open into the canal channel at the zone of outflow. Dissolved contaminants will remain in the water as it enters and moves through the aquifer, so enforcement of sanitation and pollution control regulations will be important in order that chances for contamination of the pool waters be minimized. However, the risk of accidental spills remains. Such spills should be anticipated and plans made for handling such emergencies before polluted water can enter the aquifer.

The zones of outflow might be temporarily dammed off until floating contaminants such as oil or other insoluble materials could be skimmed off or pumped out of the canal with modern barge- or boat-mounted cleaning apparatus before the contaminants could reach the zone of outflow. Studies indicate that the chance of contaminated water entering the aquifer directly from the lower pools, especially Eureka and Rodman Pools, is much less than in the Summit Pool. Therefore, if contaminants, especially highly water soluble ones, could not be removed by the means just mentioned, the Summit Pool might be drained rapidly through the locks. Thus the contaminants could be removed through the lower pools and out to sea with less risk to ground-water supplies. The stage of the Summit Pool would be lowered temporarily, and flow would be reversed in the zones of outflow so that water could not enter the aquifer from the pool. In fact, if some contaminants had already entered the aquifer, they might be flushed back out. Also, studies indicate that water can be moved out of the Summit Pool through the locks much faster than the pool can be replenished with water from the aquifer. Of course, other considerations such as risks to aquatic life in the lower pools and estuaries would have to be weighed against the risk to the quality of the water in the aquifer.

Only the more obvious considerations concerning pollution have been mentioned, but the possible pollution of the aquifer through zones of outflow is of primary importance, and should be thoroughly studied and documented before the canal is placed in operation. However, the collection of information on actual rates of flow in the canal from zones of inflow to zones of outflow, the effects of outside influence such as winds on rate of surface-water circulation in the pool, the exact locations for concentrated outflow, size of solution channels opened by the excavation in the zone of outflow, and on other factors as well, will have to wait for actual canal excavation.

It may be necessary, after all the facts concerning water movement in and out of the Summit Pool are known, to implement monitoring procedures to track the movement of contaminants to a zone of outflow.

If during the excavation of the Summit Pool in the probable zone of principal outflow (fig. 32), solution channels large enough to cause a large concentration of outflow are found, consideration may be given to partial grouting of the cavities in order to spread the outflow over a wider area. This would improve the natural filtration of the canal water as it passes into the aquifer.

Presumably the Summit Pool will be excavated in a series of sections separated by plugs which will virtually preclude circulation between the sections during the major part of the digging. This will tend to prevent movement through the outflow zones of most of the turbid water bound to result temporarily from the excavation work. Consideration should be given to de-watering excavation sites in the zones of outflow to further reduce movement of turbid water into the aquifer, although the large volumes of water to be handled and disposal problems may make this impractical. Sectional excavation would provide opportunities to solve unforeseen problems while still practicable.

Salt-water migration into the upper pools of the waterway, through the locking process, is sometimes a problem with lock-equipped waterways that rise from coastal areas. The seriousness of the problem may vary considerably from one waterway to another, depending to a large extent on the relative amounts of step or rise at the locks. Salt water that has moved into the lock along the lock bottom from the lower level is intermingled with fresh water from the upper level during the lock filling operation. However, if mixing by turbulence and other means is not complete, some of the salt water tends to sink back to the bottom of the lock after the lock is filled. As the upstream gates are opened to allow traffic to move into the upper pool, the resultant mixture of fresh and salt water high in the lock tends to flow into the upper pool along the bottom as the higher density water sinks in the fresher water of the upper pool. The greater the water-level differential at the lock, and the lesser the turbulence induced by the locking procedure, the smaller the amount of salty water that is lifted high enough during filling to flow into the upper pool.

The step in water level at the ends of the Barge Canal is considerable, that is, about 20 feet at St. Johns Lock and about 25 feet at Inglis Lock. The St. Johns Lock is far enough up the St. Johns River valley from the Ocean to virtually eliminate any risk of salt-water movement into the east end of the canal. However, on the west end of the canal, salt water from the Gulf is present in the canal to Inglis Lock. The 25-foot step-up from salt water to fresh water in the Inglis Pool and the flushing action of continuous flow from Inglis Pool to the lower reaches of the Withlacoochee River will control upstream salt water movement to a large extent, but the proximity of the salt water to the Inglis Pool warrants a close examination of the chances for an appreciable migration of salt water into the Inglis Pool, and even up into the Summit Pool and ultimately into the aquifer. Though it seems unlikely that salt-water migration would be appreciable in a short period of time, long-term migration could present a problem if proper precautions are not taken. Preventive procedures, if made a part of the routine locking operation should maintain salt-water migration at an acceptable minimum. Such procedures might include lock filling practices that would hold intermingling to a minimum and if practicable would allow some time for sinking of salt water back into a low position in the lock before the upstream gates are opened. Also, the Inglis Lock by-pass channel from the Inglis Pool to the lower Withlacoochee River will drain off some or possibly all of the salty locked water.

As a part of the canal quality-of-water monitoring program, continuous observations are to be made at the upstream end of each lock in the canal, in order to quickly detect any upstream migration of salt water.

Another water-quality question needing consideration is the possibility of introducing contaminated water into the aquifer at the outflow zone of the Summit Pool by pumpage from Eureka Pool to replace lockage losses. The pumping station at the downstream end of Dosh Lock will be near the area of inflow to the Eureka Pool from both the Silver River and the upper Oklawaha River. Pumpage to the Summit Pool is likely to consists of a mixture of these two waters plus the lockage water from the Summit Pool itself. If the pumpage were to consist almost exclusively of a mixture of unaltered flow from Silver Springs through Silver River and lockage loss from the Summit Pool, little reason for concern about contaminating the aquifer would exist, as these waters are unlikely to be contaminated. On the other hand, if a significant quantity of upper Oklawaha River water or comparatively stagnant reservoir water, possibly with high nutrient content, turbidity, and color, were pumped to the Summit Pool, the quality of the water in the Summit Pool would deteriorate and a chance for contamination of water in the aquifer would result. Also, consideration should be given to possible water quality problems resulting from movement of significant quantities of water from Eureka Pool or Oklawaha River basin through the Summit Pool into Inglis Pool, and thereby into the Withlacoochee River basin.

Again, proper canal operation practices and pump intake locations may go far toward eliminating such possibilities for contamination. Monitoring the pump intake area for pollution should generally preclude inadvertent pumping of contaminated waters into the Summit Pool. Also, operation of a pumping station below Dunnellon Lock to return westward lockage losses to the Summit

Pool directly from Inglis Pool, rather than to replace all Summit Pool losses with pumpage from Eureka Pool, may be a solution to the possible problem of excessive movement of Eureka Pool water into Inglis Pool.

One more water-quality consideration relative to pumping water into the Summit Pool from lower pools is that water from predominantly surface sources may be more undersaturated with respect to calcium bicarbonate than are the ground-waters normally circulating through the aquifer in the Summit Pool area. This, coupled with the fact that the water may have a comparatively high carbon dioxide content gained from aeration during pumping, could increase the solution of limestone in the zone or zones of outflow from the Summit Pool into the aquifer. This would tend to increase aquifer permeability in such zones.

Special Survey of City of Ocala Drainage Well System

During the investigation several water-level measurements were made at three of the four supply wells in the city of Ocala well field. Abnormally high water levels were noted in two of the wells and use of these measurements as control for mapping the potentiometric surface of the Floridan aquifer produced a configuration which was 4 or 5 feet higher than expected.

Several possibilities were considered as the cause for the high water levels, one being that they may represent a mound in the water table resulting from concentrated recharge to the aquifer by about 40 storm-drainage wells drilled to augment and improve the natural internal drainage system of sink holes, many of which tended to overflow following times of heavy runoff. Another possibility considered was that since the supply wells were deeper than most other wells in the observation network, a higher hydraulic head may exist at depth. However, one supply well which was as deep as the two wells in which water levels were anomalous had a normal water level.

To check the possibility of mounding, a special survey was made in which water levels were measured in 26 of the drainage wells. The survey showed the abnormal potentiometric high did not exist and that only a northward plunging ridge of low relief exists in the potentiometric surface in the area. The ridge appears to be related to the topographic high of the Ocala Hill physiographic subdivision (fig. 5), although a slight mound near the north end of the ridge may be due to almost continuous drainage through wells of two large runoff-retention ponds.

The survey also indicated that only the two public supply wells and one other deep industrial supply well on the north edge of town were erroneous water-level measurements. The three wells were found to have a 30- to 35-foot column of pump lubricating oil floating on the water, and when water-level readings were adjusted for the oil column, they were near normal.

Even though a high potentiometric mound from the storm-drainage system does not exist, drainage entering the aquifer through the wells moves down the slopes of the subtle potentiometric ridge underlying Ocala, posing a risk to the quality of the water in the aquifer and ultimating moving through the aquifer toward Silver Springs. By permitting direct flow from the surface into the aquifer without benefit of the natural filter provided by the surficial sand and clayey sand present in proximity to a drainage well, is subject to pollution from storm drainage. However, if the aquifer surrounding the well bore is not especially cavernous, or if cavities are sand-filled, the rock of the aquifer and (or) the sand fill will act as a filter and will tend to purify the water as it moves away in the aquifer. In a highly cavernous area not sand filled the contaminated water may mover much farther before it is cleaned up. State regulations now limit the drilling of drainage well in the state because of the risk of contaminating ground-water supplies.

Additional Studies Needed or Considered

Geology

Additional knowledge of the geology is needed to confirm and refine some concepts of the geohydrology of the Barge Canal area as set forth in this report. The basic need is for more precise knowledge of the stratigraphy of the area, without which detailed delineation of the geologic structure is difficult. A thorough knowledge of both stratigraphy and structure is essential to the best definition of the hydrogeologic framework. It is intended that as construction of the canal progresses, every effort will be made to map, sample and study all new stratigraphic exposures resulting from the excavation work. Additional geological information should be gathered by examination of good sets of lithologic samples available from test holes and wells. Various remote sensing methods, such as aerial infrared and color photography, should be investigated and utilized where possible to aid in the delineation of certain stratigraphic and structural characteristics of the area, the knowledge of which may help to further define routes of preferential ground-water flow, and thereby aid in preventing pollution of water in the aquifer.

Hydrology

Consideration should be given to the possible benefits resulting from making a series of large-scale pumping tests along the centerline of the Summit Pool prior to excavation. The purpose of such tests would be to confirm or refine knowledge of the quantitative characteristics of the Floridan Aquifer gained through flow-net analysis and other investigative methods.

The tests would require large diameter wells cased into the top of the aquifer, and drilled to the design bottom depth for the canal at 28 feet above mean sea level. A set of smaller diameter observation wells would be required around each well pumped. The objective of the tests would be to gather the data necessary to accurately simulate the circulation between the aquifer and the canal. If adequate data were obtained, an analog model could be produced by which various water-level and water-quality control operations in the Summit Pool could be tested.

A possible secondary use for wells drilled for these tests could be tracer dye tests for direct measurement of ground-water flow velocities within short distance of the Summit Pool.

Also, any dewatering operations likely to be made as a part of canal construction should be utilized as pumping tests. Therefore, provisions should be made in future dewatering contracts for proper monitoring and documentation of water levels and pumping rates. Such use of dewateing operations at the Dunnellon and Dosh Lock excavations would be particularly valuable.

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