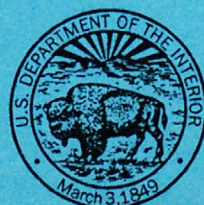# Computation of the Time-Varying Flow Rate from an Artesian Well in Central Dade County, Florida, by Analytical and Numerical Simulation Methods

U.S. Geological Survey

Open-File Report 94-457

Prepared in cooperation with the

Metro-Dade County Department of Environmental Resources Management

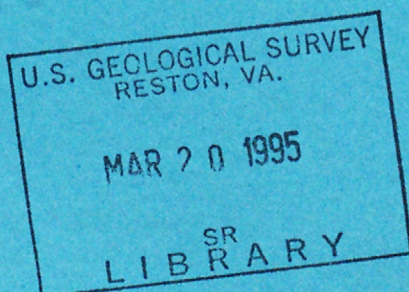# Computation of the Time-Varying Flow Rate from an Artesian Well in Central Dade County, Florida, by Analytical and Numerical Simulation Methods

By Michael L. Merritt

U.S. GEOLOGICAL SURVEY

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U.S. DEPARTMENT OF THE INTERIOR
BRUCE BABBITT, Secretary

U.S. GEOLOGICAL SURVEY
Gordon P. Eaton, Director

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CONVERSION FACTORS, VERTICAL DATUM, ACRONYMS, AND ABBREVIATIONS

Multiply	By	To obtain
inch (in.)	25.4	millimeter
inch per year (in/yr)	25.4	millimeter per year
foot (ft)	0.3048	meter
foot per day (ft/d)	0.3048	meter per day
foot squared per day (ft²/d)	0.09290	meter squared per day
mile (mi)	1.609	kilometer
acre	0.4047	hectare
gallon per minute (gal/min)	0.06309	liter per second
pound per square inch (lb/in²)	6.895	kilopascal
pound per cubic foot (lb/ft³)	0.01602	gram per cubic centimeter

Temperature in degrees Celsius (°C) can be converted to degrees Fahrenheit (°F) as follows:

$$°F = 1.8 \times °C + 32$$

Sea level: In this report, "sea level" refers to the National Geodetic Vertical Datum of 1929--a geodetic datum derived from a general adjustment of the first-order level nets of the United States and Canada, formerly called Sea Level Datum of 1929.

Acronyms

FGS = Florida Geological Survey
USGS = U.S. Geological Survey
DERM = Metro-Dade County Department of Environmental Resources Management
SFWMD = South Florida Water Management District

Abbreviated Water-Quality Units

mg/L = milligrams per liter
µS/cm = microsiemens per centimeter

Computation of the Time-Varying Flow Rate from an Artesian Well in Central Dade County, Florida, by Analytical and Numerical Simulation Methods

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Abstract

Simulation of the development of a plume of brackish water in a surficial aquifer caused by discharge from a flowing artesian well required an accurate estimate of the rate of flow during a 40-year period. The rate of flow from the well, constructed with 12-inch casing in 1944, was measured to be 2,350 gallons per minute 2 months after completion and 1,170 gallons per minute in 1964, 1965, and 1969. The rates measured 20 years apart appeared to be mutually inconsistent unless the difference could be explained by the: (1) drawdown of the aquifer over time, (2) raising of the altitude at which the water was discharged, (3) installation of 80 feet of 8-inch liner, or (4) deterioration of the condition of the well over time. The latter possibility implies losses of flows through holes in the casing and increased friction losses. Application of an analytical solution relating the time-varying rate of flow to a constant opposing head (a rewriting of the "constant drawdown" formula) failed to reconcile the measured rates, though estimates provided by use of the formula differed by only about 15 percent.

Numerical simulation techniques were also used to estimate the rate of artesian flow from the source aquifer, a permeable zone at about 1,200 feet below land surface, near the top of the Upper Floridan aquifer in southeastern Florida. The selected simulation code contained a well-riser model that was used to account for friction losses in the well. Construction of a highly generalized model of the Floridan aquifer system for the flow-rate analysis, and the determination of a set of predevelopment head values for use as an initial condition, permitted simulation of the effects of recharge through leaky confining layers or from aquifer boundaries at a finite distance from the well. The conceptual model supported by the calibration of the model of the Floridan aquifer system is that a hydraulically uniform Lower Floridan aquifer provides recharge to the Upper Floridan aquifer through a leaky middle confining unit, and head variations in the Upper Floridan aquifer are related to variations in the thickness of the zone of fresh and brackish water extending downward into the confining unit.

Results of the flow-rate analysis indicated that the flow rate should reach equilibrium after about 1 week because the Upper Floridan aquifer received recharge from the Boulder Zone through the intervening middle confining unit. The well modifications probably only decreased the rate of flow by 18 to 19 percent. A sensitivity analysis indicated that variation in the roughness coefficient of a degree that could represent severe deterioration of the well casing decreased the estimated flow rate by about 17.5 percent. Another sensitivity analysis indicated that installation of 80 feet of 8-inch liner had only a slight effect on the flow rate. The flow-rate inconsistency was not fully

resolved by the analysis, but could be explained as a combination of the result of well modification, deterioration of the condition of the well, and other factors not amenable to analysis (inaccuracy in one or more of the flow-rate measurements or greater than estimated losses through the casing). The transmissivity of the source aquifer was estimated to be 11,125 feet squared per day at the well site. Besides the construction of a generalized model of the Floridan aquifer system, the most significant result of the analysis was the demonstration of a simulation approach for accurately showing the relation between aquifer characteristics and the rate of flow from artesian wells. However, this approach requires considerably more analytical effort and data describing aquifer properties than does application of the constant drawdown formula.

INTRODUCTION

The rapidly increasing demand for drinking water and the potential for contamination of surficial aquifers have led the counties of southern Florida to develop new well fields farther inland and hydraulically upgradient of population centers and commercial activity that could be a source of aquifer contamination. Dade County has investigated the possibility of locating a new well field tapping the surficial Biscayne aquifer in the vicinity of Levee 31N south of Tamiami Trail (fig. 1), an area of relatively undeveloped wetlands. However, even in this remote location, a potential contamination hazard in the form of a body of brackish water in the surficial aquifer was found. The elongated plume of brackish water was delineated by a U.S. Geological Survey' (USGS) reconnaissance of water quality (Waller, 1982). The source of the brackish water was identified as a flowing artesian well (the "Grossman well") in Chekika State Recreation Area (fig. 1).

After Dade County arranged to end the spill of brackish water, the USGS entered into a cooperative agreement with the Metro-Dade County Department of Environmental Resources Management (DERM) to use digital models to assess the potential future movement and rate of dissipation of the body of brackish water. The objective of the study was to provide a tool to assess the possible interaction of the plume with the hydraulic influence of the proposed new well field. Various results of the study, including a regional flow model and a digital simulation of the plume, have been documented separately (Merritt, 1995a and 1995b, in review).

A simulation of the plume development required an estimate of the flow rate of the well sufficiently accurate that any uncertainty would not appreciably affect the conclusions of the analysis. However, an early flow measurement in 1944 appeared inconsistent with measurements made in 1964, 1965, and 1969, and there were no known reasons to consider any of the measurements to be less accurate than the others. Possible explanatory factors were: (1) an error in one or more measurements of flow, (2) a decrease in the rate of flow as the potentiometric surface of the Upper Floridan aquifer was drawn down, (3) an increase in the altitude at which the water was discharged after redesign of the well, (4) the installation of the narrow 80-ft liner, (5) increasing friction losses from corrosion of the well casing, or (6) an underestimate of losses through the corroded casing when the later measurements were made.

A commonly used method of estimating transmissivity from artesian flow rates and their variation with time is the constant drawdown analysis (Lohman, 1979, p. 23). A variation of this method was used in this study to estimate flow rates. However, the method does not account for the important influence on the flow rate of friction loss in the wellbore and casing, which varies with the flow rate, or for recharge through aquifer boundaries at a finite distance from the well. For this reason, a numerical model that contained a well-riser model with the capability to estimate friction losses was selected for the analysis. However, the well-riser model and wellbore model in the selected simulator required modification to make possible an iterative solution of the relation between heads in the aquifer and artesian flow from the well. In addition, previous attempts to simulate heads in the Floridan aquifer system on the Florida Peninsula have foundered because the simulators used lacked the capability to take into account density variations caused by coastal saltwater intrusion. A variable-density simulator was needed to construct a simple, generalized model of an areally extensive flow zone within the Upper Floridan aquifer in southern Florida so that the interaction of aquifer drawdown, recharge through aquifer boundaries, and wellbore friction loss as a function of time could be simulated.

Figure 1. Location of Chekika State Recreation Area, the study site, in Dade County, Florida.

The use of numerical simulation procedures to resolve the flow-rate inconsistency question represents a novel and more sophisticated, though computationally more intensive, approach to this class of problems than the more familiar use of analytical methods. The modeling results add to the tools available to water managers for assessing the rate of spillage from artesian well sources and for making estimates of aquifer transmissivity from observed flow rates.

Purpose and Scope

This report presents the results of the analytical and numerical simulation analyses of the rate of flow from the Grossman well and compares the results obtained by using the two methods. Best estimates of the total flow rate, its variation with time, and aquifer characteristics are derived from the numerical simulation analysis. The report describes the result of the attempts to resolve the difference in measured rates on the basis of changes in the design and condition of the well and the effects of flow on the aquifer system.

To provide a basis for the analysis, local hydrogeologic conditions in the Upper Floridan aquifer are identified using data from the site of the flowing well and from other sites in Dade County. A generalized regional model of the artesian zone yielding flow to the well used for the numerical simulation analyses is described, as are modifications made to a simulator to make the flow-rate analyses possible. The report also describes additional analyses performed with the calibrated model to assess the sensitivity of simulation results to values specified for parametric coefficients representing the geometry and condition of the well.

Acknowledgments

The author wishes to acknowledge the efforts of the following individuals: Ms. Heather Dewar and Ms. Wendy Ledis of the Miami Herald in the search for and loan of a black-and-white photograph of the flowing artesian well for reproduction in this report; Dr. Robert G. Font of the Oryx Energy Company of Dallas, Tex., whose search of the computerized Sun Oil Company data base quickly located a complete copy of the Grossman well lithologic log of 1944; and Mr. Thomas Sechler of the Florida Department of General Services, Mr. Donald Bizzell of the Florida Department of Natural Resources, Mr. Marty Braun of the South Florida Water Management District (SFWMD), and principals of Webb's Well Drilling of Jupiter, Fla., for details of the plugging of the artesian well and site redevelopment of the Chekika State Recreation Area. Additionally, the author thanks James Brenenstuhl, Florida Department of Natural Resources, Assistant Ranger at the Chekika State Recreation Area, for details of the history of the park and flowing well. The author also thanks Mr. Nevin D. Hoy, retired, USGS, for additional historical details. Appreciation is expressed to Charles A. Appel, USGS Office of Ground Water, Reston, Va., for his assistance in locating source material describing the basis of the computational method used for determining the rate of flow from an artesian well.

DRILLING OF THE ARTESIAN GROSSMAN WELL AND HYDROLOGIC CONSEQUENCES

The drilling of oil test wells in Dade County began in 1939 near Pinecrest, Fla., about 4.5 mi west of the county boundary (fig. 1), at Cory No. 1, owned by the Peninsular Oil and Refining Company. Two other early wells were drilled by W.G. Blanchard near Forty-Mile Bend during 1940-44. Owners of a sparsely developed area of land in central Dade County allowed the Miami Shipbuilding Company to prospect for oil on part of the land in 1944 under provisions of Trustees of the Internal Improvement Fund State Lease no. 1. A test well was drilled in this seasonally inundated part of the Rocky Glades (a term applied to the region where limestone of the highly permeable Biscayne aquifer extends to land surface). The well site was near a natural hardwood hammock (a small area, 0.5 to 1.5 ft above the average surrounding land surface and high enough to be suitable for the growth of hardwood vegetation). The land owners later gave their family name to the hammock, so that it became known as Grossman Hammock. The well was commonly referred to as the Grossman well. The well site was about 6 mi east of the center of the Everglades flow system, known as Shark River Slough.

The drillers requested information and technical advice from local offices of the Florida Geological Survey (FGS) and the USGS and reported the occurrence of artesian freshwater with an odor of sulfur at a depth of 1,200 ft. A visit to the site in November 1944 by N.D. Hoy of the USGS, at the request of the FGS, revealed that all drilling equipment had been removed, and water was flowing from a piece of 12-in. black iron casing protruding about 1 to 2 ft above land surface. (Land surface was about 8 ft above sea level at this location.) Drilling was said to have been discontinued 2 months earlier (N.D. Hoy, U.S. Geological Survey, written commun., 1944). Hoy measured the flow to be about 2,350 gal/min, using an indirect method derived empirically by Lawrence and Braunworth (1906). The relation was later derived from physical principles by Rouse (1949). The method requires measurements to be made of the height of the jet above the lip of the casing (h) and of the inner diameter of the casing (d). When these are expressed in inches, the flow rate of the well in gallons per minute (Q) is computed as:

$$Q = 5.68 \ d^2 h^{1/2} \tag{1}$$

The Water Measurement Manual (U.S. Bureau of Reclamation, 1984) cautions that "the principal difficulty with this method is in measuring the coordinates of the flowing stream accurately."

Hoy also collected a water sample. The water proved to be brackish rather than fresh, and water samples collected subsequently have had chloride and dissolved-solids concentrations of about 1,200 and 3,000 mg/L, respectively. The best available information on the original depth of the well was obtained by F.A. Kohout (U.S. Geological Survey, written commun.) from Mr. McCord, an oil company owner, sometime before 1966. McCord cited a depth of 1,248 ft. The well was 1,250 ft deep when logged in 1983. The depth interval from which the brackish water flowed was within the permeable upper part of the Floridan aquifer system, known informally as the Upper Floridan aquifer.

On subsequent visits, Hoy reported that the well was crudely capped by an inverted drill bit that only made partial contact with the side of the casing, and that pieces of wood were jammed into the remaining openings to stop water from escaping. Nevertheless, not all openings were effectively plugged, and jets of water rose 6 to 10 ft above the wellhead.

A second oil test well, USGS local number G-3234, was drilled nearby by the Coastal Petroleum Company in late 1949 (Trustees of the Internal Improvement Fund no. 1, Lease 340-A, permit no. 115). This well extended to 11,500 ft below land surface, and negligible data were collected from depth intervals corresponding to the source of brackish water flows in the earlier well. The deep well produced only a small amount of oil (legally a dry hole) and was abandoned several years later. The location of this well is no longer marked by visible signs at land surface but is probably less than 100 ft from the earlier well. The deep oil test well has been cited in numerous regional inventories of deep wells (Chen, 1965; Maher, 1971; Puri and Winston, 1974; Beaven and Meyer, 1978; and Smith and others, 1982).

A small (1.5 acre) and shallow (4 ft deep) lake was formed at the site, probably in the middle or late 1940's, by the construction of soil dikes to contain the well discharge (Adler, 1973). The flowing well was located on the spoil bank. Overflow from the lake was channeled into a second smaller borrow pit, 15- to 20-ft deep, excavated in 1947 to obtain fill for building soil embankments to the south and east of the two ponds. When the water table was high, water from the borrow pit spilled southward over the low Everglades land contained within the outer soil embankments that prevented further southward flow of the artesian water into adjacent Everglades grasslands.

In the early 1950's, the landowners landscaped the area around the flowing well for use as a public recreational area and spa. Water from the flowing well emerged from the top of an ornamental cairn of rocks surrounded by a small pool from which water cascaded over an artificial waterfall into the lake (fig. 2). The exit level of the brackish water was raised to about 8 to 10 ft above land surface by the soil dike and cairn of rocks.

Through holes in the clay liner on the bottom of the lake, through the sides of the 15-ft deep borrow pit, and from the surface of the rocky glades land, the brackish water from the well infiltrated the 45-ft thickness of the surficial Biscayne aquifer. Flow from the well was reported by the owners to have
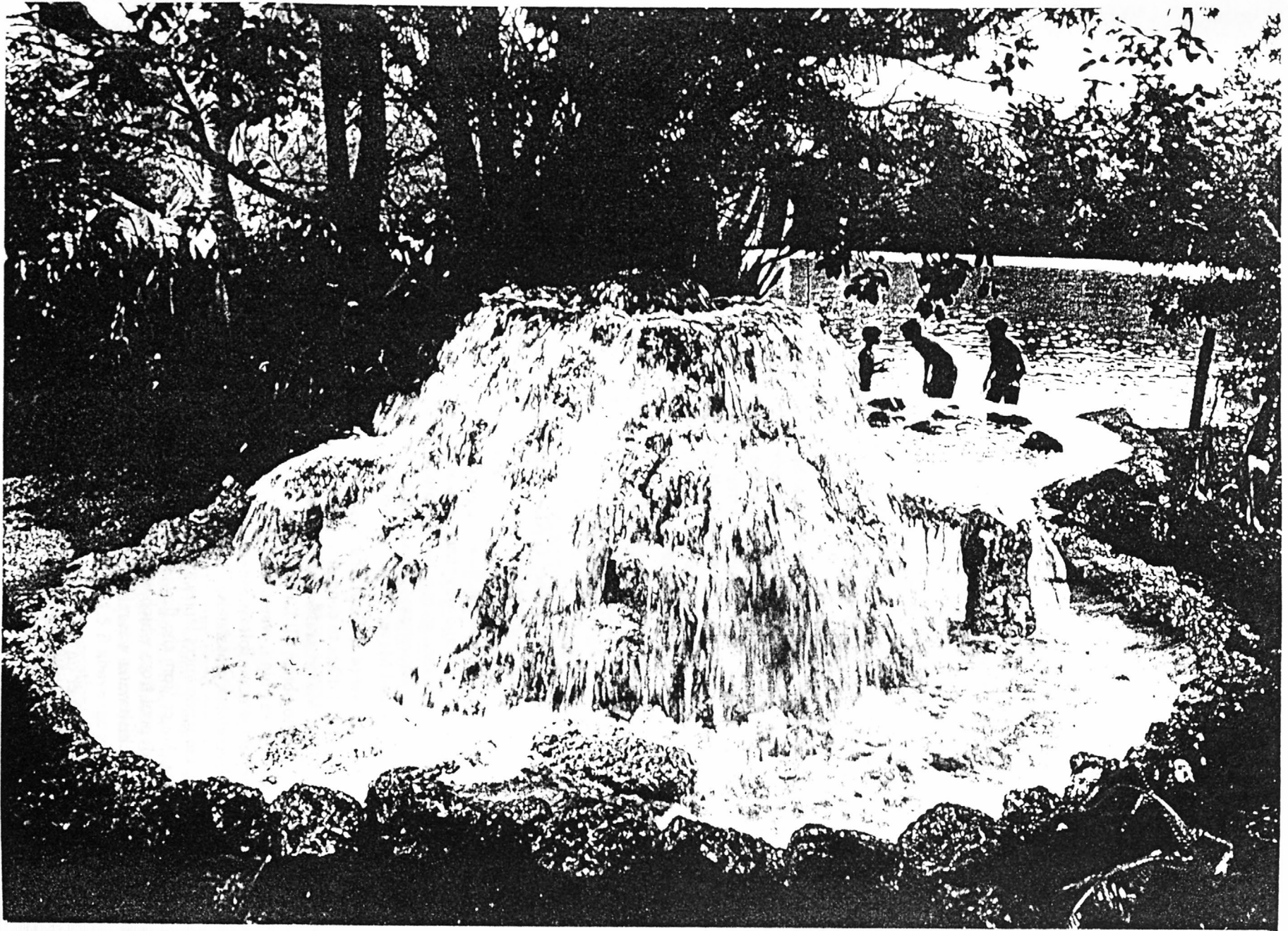
Figure 2. The flowing Grossman well at Chekika State Recreation Area and surrounding landscape, in June 1970. Lake Chekika is beyond and to the right of the well. Photograph courtesy of the Miami Herald.

diminished substantially in the early to mid-1950's, suggesting that brackish water had another means of entering the Biscayne aquifer, possibly through corrosion holes in the casing. An 8-in. polyvinyl chloride liner was hung in the well to 80 ft, reportedly in 1958. However, this probably did not entirely eliminate the flow of artesian water through holes in the casing, as no documentation presently available indicates that the liner was sealed at the bottom with cement.

After the Alaskan Earthquake of 1964, the flow rate at the wellhead was measured by Howard Klein of the USGS to resolve questions about the effect of the earthquake upon the artesian flow. The measured rate was about 1,170 gal/min. The measurement technique was the same as used by Hoy in 1944. Subsequent measurements by F.W. Meyer of the USGS in 1965 and 1969 (remeasurements of the height of the jet above the lip of the casing) indicated that the flow rate was about the same as in 1964. In 1970, a section of land including the park was purchased by the State of Florida and opened to the public as a State facility that was later named Chekika State Recreation Area (fig. 1).

A regional ground-water quality reconnaissance by the USGS during 1978-79 revealed the presence of a plume of brackish water in the Biscayne aquifer that extended about 8 mi southeastward of the well (Waller, 1982). DERM obtained an agreement with the Florida Department of Natural Resources that the well would be plugged to prevent additional contamination of the Biscayne aquifer. After construction of two supply wells in the Biscayne aquifer to provide water to a new fountain, the well was plugged in March 1985. For the next 2 years, DERM continued water-quality monitoring in the area affected by the plume (Labowski, 1988). The USGS, in cooperation with DERM, began a study of the plume using numerical simulation techniques in 1985.

HYDROGEOLOGIC CONDITIONS IN STUDY AREA

Hydrogeologic conditions near the Grossman well in central Dade County are inferred from a few sets of lithologic descriptions, geophysical logs of the well, and descriptions of the quality of water from several wells, including the Grossman well. These data are then interpreted to define the stratigraphy and to describe lithologic properties, depth to producing zones, the condition of the well, and the quality of water produced by the artesian zone. Because no aquifer tests of the artesian zone were performed at the site, data from other sites are cited to provide estimates of hydraulic properties (hydraulic conductivity and porosity). The direction and rate of ground-water flow in the artesian zone are estimated on the basis of previous studies.

Lithology and Stratigraphy

A description of the lithology of the surficial aquifer system near Chekika State Recreation Area (fig. 1) was prepared by Causaras (1987) based on rock samples acquired during drilling of a 250-ft test hole (USGS local number G-3310) located about 1,500 ft north of the Grossman well. Causaras (1987) found the bottom of the surficial aquifer system, the boundary between the Tamiami and Hawthorn Formations, at 215 ft. The latter formation is principally a sandy, silty, or clayey marl. The base of the Biscayne aquifer, locally identified as rocks of the Miami Limestone (Hoffmeister and others, 1967) and the Fort Thompson Formation, is at 45 ft below average local land surface.

Lithologic descriptions of local subsurface deposits that extend to the depth of the Grossman well (1,250 ft) are available only from the drilling of the well itself. The drilling log from the nearby 11,500-ft oil well lacks detail, merely referring to lime, with sand and shells found in the lime above 762 ft. The descriptions of samples in the Grossman well by Louise Jordan (Sun Oil Company, written commun., 1944) begin at 290 ft. The emphasis in these descriptions (see appendix) is on the identification of fossil species. Clayey sand is present to 475 ft. Miocene and Oligocene fossils were found to 1,150 ft, which is identified, with some reservations as to the exact depth, as the top of the Claiborne Group (equivalent in stage to the Avon Park Formation in southern Florida [Applin and Applin, 1967]). The Miocene and Oligocene deposits correspond to the Hawthorn Formation and the Tampa and Suwannee Limestones.

The Avon Park Formation of Eocene age, herein defined to include the former Lake City Formation as recommended by Miller (1986), is underlain by the Oldsmar Formation. All are composed of layers of limestone and dolomite. The Oldsmar Formation includes a dolomitic layer of cavernous porosity having exceptionally high permeability. The top of this layer, known to drillers as the Boulder Zone and to hydrologists as the Lower Floridan aquifer, generally occurs from 2,700 to 3,200 ft below land surface in southern Florida. The rocks between the Upper and Lower Floridan aquifers are known as the middle confining unit of the Floridan aquifer system and are generally of low permeability, though layers of permeable dolomite are present in some wells. Rocks below the Boulder Zone are massive dolomites of low permeability.

Interpretation of Geophysical Logs

Selected geophysical logs acquired from the Grossman well (fig. 3) helped in further analysis of subsurface conditions and the condition of the well. Table 1 presents a complete list of geophysical logs known to have been acquired from the well. All were run during natural artesian flow at a rate of about 1,170 gal/min. The caliper log by the SFWMD on November 3, 1983, clearly shows the bottom of the 8-in. liner at 80 ft and the bottom of the original 12-in. iron casing at 486 ft. The earlier (June 11, 1969) single-point resistance log run by the FGS provides confirmation of the casing depths. Variations in the single-point trace within the open borehole are generally considered to correspond to variations in lithology, and the signal should show increases next to clayey intervals.

Both FGS caliper logs (June 11 and August 15, 1969) showed an apparent constriction that could be a squeezing inward of clayey material between 580 and 605 ft. Two of the 1969 sets of spinner flowmeter readings (before diameter compensation) show an anomalously high value in this depth interval that could be associated with a localized constriction in hole size. However, the SFWMD log of November 1983 (fig. 3), while showing abrupt borehole diameter changes in this depth interval, does not show a significant narrowing of the hole. The remaining sections of the FGS caliper logs of the borehole were similar to the SFWMD log. The USGS caliper log of 1974 was reported to have been prevented from reaching below 500 ft by an obstruction in the hole. The ledge or narrowing of the hole centered at 1,140 ft is confirmed by both of the FGS caliper logs, indicating a minimum diameter of 7 or 8 in., and by the SFWMD log, indicating a minimum diameter of 9 in. All caliper logs show a highly rugose borehole between 1,170 and 1,210 ft, suggesting the possibility that major solution features might occur in this interval.

The SFWMD natural gamma log of November 3, 1983, shows an interval of high counts between 1,125 and 1,175 ft that correlates with phosphatic beds present near the base of the Oligocene throughout southern Florida (Meyer, 1989a, p. 13). Thin, discrete flow zones are often present near the Oligocene-Eocene contact below this marker bed (Merritt, 1995c, in press), and this is confirmed at the Grossman well site by the three flowmeter logs run by the FGS. Each log consisted of stationary point readings; that is, spinner flowmeter counts were recorded while the probe was held motionless at selected depths within the well. The counts were converted to flow rates based on the hole or casing diameter at the appropriate depth, and the rates were expressed as a percentage of casing flow. The 1983 SFWMD caliper log was used for the diameter compensation.

The results of the flow-rate conversion are shown in figure 3. Downward spikes at 590 and 1,140 ft resulted from use of the measured diameter of the borehole where the caliper logs indicated it to be sharply reduced. Apparently, the small hole diameter was larger than measured by the caliper log. Considering the high calculated flows between 1,150 and 1,180 ft, there is little evidence for appreciable flows from the formation above 1,180 ft. The primary flow zone seems to occur between 1,180 and 1,205 ft. Below this interval, flow is virtually zero.

The two flowmeter logs with data above 100 ft seem to show a slightly lower rate of flow within the 80-ft liner than just below, suggesting that some water may have entered the annular space between the liner and the corroded 12-in. casing, entering the Biscayne aquifer through holes in the latter. Because the flow rate in the liner seems to be about 20 percent less than the flow rate measured in the casing between 250 and 400 ft (where the two flowmeter logs agree), and the flow rate measured at the top of the well in 1964, 1965, and 1969 was 1,170 gal/min, the total flow rate from the artesian zone is estimated to be 1,400 gal/min.
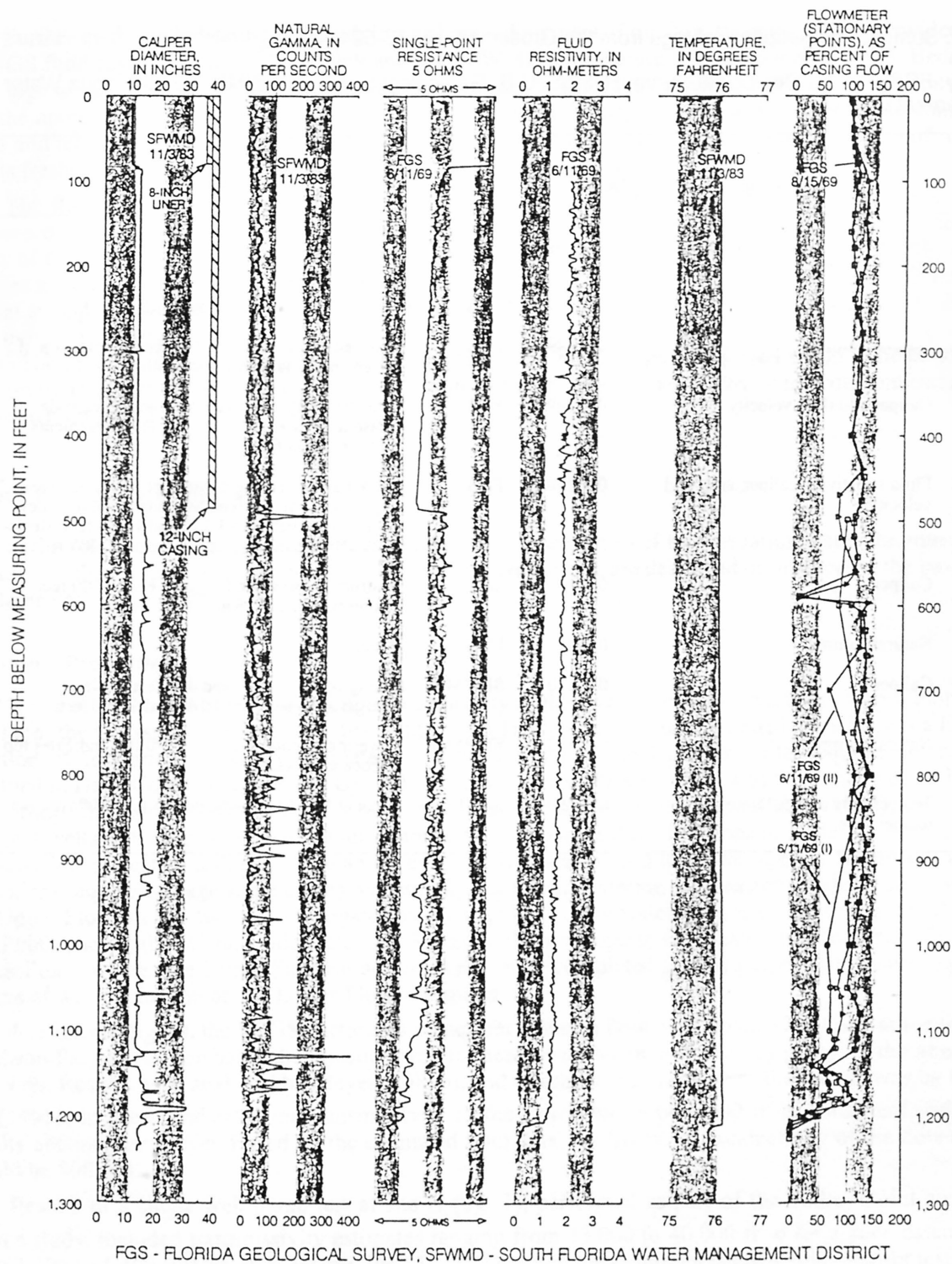
Figure 3. Geophysical logs of the Grossman well, Chekika State Recreation Area.

Table 1. Summary of geophysical logs from the Grossman well (S-524)

[Agency: FGS, Florida Geological Survey; USGS, U.S. Geological Survey; SFWMD, South Florida Water Management District]

Date acquired	Type of log	Depth (feet)	Agency	Remarks
06-11-69	Spontaneous potential and single-point resistance	0-1,249	FGS	Shows bottom of casing at 485 feet.
	Fluid resistivity	0-1,149	FGS	Rapid salinity increase below 1,205 feet.
	Natural gamma	0-1,249	FGS	Reduced counts below 1,165 feet indicate top of Eocene rocks is below this depth.
	Caliper and fluid velocity	0-1,249	FGS	Two stationary point velocity runs. Counts decrease to zero at 1,200 feet. Hole enlargements 1,140 to 1,200 feet.
08-15-69	Fluid resistivity, caliper, and fluid velocity	0-1,240	FGS	Caliper shows casing bottoms at 84 and 482 feet; clay squeeze at 580 feet. Stationary point velocities decrease between 1,120 and 1,216 feet. Resistivity shows salinity increase below 1,202 feet.
06-25-74	Caliper	0-500	USGS	Obstruction prevented logging below 500 feet. Casing bottoms at 76 and 481 feet.
	Natural gamma	0-500	USGS	None.
11-03-83	Caliper	0-1,240	SFWMD	Casing bottoms at 80 and 486 feet. Hole enlargements between 1,160 and 1,212 feet.
	Natural gamma	0-1,250	SFWMD	Lowering of counts below 1,170 feet indicates top of Eocene rocks.
	Temperature and differential temperature	0-1,250	SFWMD	Break at 1,208 feet indicates bottom of flow zone.

Further evidence indicating the depth interval providing most of the flow from the well is provided by the FGS fluid-resistivity log of June 1969 and the SFWMD temperature log of November 1983. Because both logs were run while the well flowed, the sharp breaks in the traces between 1,205 and 1,210 ft establish the approximate depth of the bottom of the flow zone as the depth above which the specific conductance and temperature of the water column are changed by an influx from a permeable flow zone of water that is fresher and warmer than the stagnant water in the bottom of the hole.

The fluid resistivity and temperature logs do not indicate the top of the flow zone. However, the absence of any evident breaks in the traces above about 1,180 ft that might indicate further influxes of water of different quality indicates that a single zone, probably the zone between 1,180 and 1,205 ft, contributes most of the flow from the well. Generally, the salinity of the formation water increases with depth, so that an appreciable amount of flow from a higher zone could have caused a noticeable break in the fluid-resistivity trace. Flow from a higher zone might also have produced a noticeable break in the temperature trace. The gradual increase of fluid resistivity upward from the bottom of the well could indicate slight contributions of less saline water from the formation, but more likely shows instrument drift, continuing as it does within the 12-in. casing.

Upper Floridan Aquifer

In the subsequent sections, the hydraulic and chemical properties of the formation that is the source of the artesian flow to the Grossman well, the Upper Floridan aquifer, are described or inferred on the basis of available data.

Hydraulic Properties

No direct estimates of the transmissivity or porosity of the flow zone found within the Upper Floridan aquifer at the Grossman well site have been obtained. (The hydraulic property called storativity is a linear function of formation porosity.) Estimates of transmissivity in the Upper Floridan aquifer are rarely obtained in Dade County because few Upper Floridan aquifer wells are drilled, and the drilling of multiple well groups that can be used for aquifer tests is even less common. At the present time (1994), groups of adjacent wells in the Upper Floridan aquifer are known to exist at three locations in Dade County: (1) the Hialeah Water Treatment Plant, where an observation well was drilled to detect injected freshwater in an operational aquifer storage and recovery system; (2) two sites in extreme southeastern Dade County, where the Upper Floridan aquifer was investigated as a source of cooling water for a nuclear powerplant at Turkey Point; and (3) the Miami-Dade Water and Sewer Authority Department Wastewater Treatment Plant at Black Point, where five Upper Floridan aquifer wells were completed for the purpose of monitoring the effects of waste injection on the Lower Floridan aquifer.

At Hialeah (fig. 1), the USGS performed an aquifer test of a flow zone about 12 ft thick at the top of the Avon Park Formation using the injection well to measure drawdowns caused by pumping the observation well. Results were analyzed by Meyer (1989b), and a computer simulation of the drawdowns by Merritt (1995c, in press) indicated the transmissivity of the flow zone to be 9,600 ft^2/d, consistent with the results obtained by Meyer. Based on the estimated thickness, the hydraulic conductivity of the flow zone would be 800 ft/d.

Results of a single-well pump test at site A (fig. 1), performed as part of the Turkey Point cooling source study, included transmissivity estimates ranging from 13,000 to 40,000 ft^2/d for a zone extending from 1,126 to 1,400 ft (Dames & Moore, Inc., 1972), although the requirements of a valid aquifer test may not have been fully satisfied (C.A. Appel, U.S. Geological Survey, written commun., 1974). A series of three subsequent aquifer tests at site B (fig. 1), about 5 mi west of site A, where measurements were made in four observation wells resulted in an estimate of about 67,000 ft^2/d (Dames & Moore, Inc., 1975).

Aquifer testing within the Upper Floridan aquifer at Black Point (fig. 1) was performed in April 1991 by a private firm under contract to the Dade County Water and Sewer Authority Department. The Upper Floridan aquifer test considered to have provided the most reliable result was a constant drawdown test

performed after modifications to the wellhead permitted sufficient discharge to impose a significant hydraulic stress on the aquifer. A transmissivity value of 2,535 ft^2/d was reported (Hydrologic Associates U.S.A., Inc., 1991), which is low compared to the values measured at the other sites. The tested interval extended from 980 to 1,020 ft. Based on an examination of gamma logs, this is approximately coincidental with the top of Eocene-age rocks where a flow zone is typically found in southeastern Florida (Merritt, 1995c, in press), but further testing would be required to verify that the flow zone is actually within the tested open interval.

Aquifer porosity is sometimes described in terms of total porosity, the volume of water contained within a volume of rock, or, in terms of effective porosity, the volume of water contained within the flow channels in the rock. The latter is water that moves most readily in response to an imposed hydraulic gradient. Total porosity is effective porosity combined with the volume of water trapped within rock pores. If the rock contains no solution features, the three types of porosity are trivially identical, as the only movement of water in response to a hydraulic gradient is interstitial (through rock pores). Many ground-water flow models in which "porosity" is a specification actually require the value of effective porosity. Neutron porosity logs measure total porosity. Specific yield in surficial aquifers is sometimes used as an estimate of effective porosity.

Few estimates of the total porosity within the Upper Floridan aquifer have been obtained because neutron porosity logs are rarely performed (drillers wish to avoid the undesirable consequences of losing a nuclear source in a borehole). Results of a nuclear porosity log run by Schlumberger, Inc., for the USGS on January 8, 1975, in the Hialeah observation well are discussed by Merritt (1995c, in press). The log was diameter compensated, and the measured values showed large variations, ranging from 20 to 65 percent. The average value of total porosity seemed to be about 35 percent in both the permeable flow zone and in the overlying and underlying confining layers.

Estimates of specific yield in the highly permeable surficial Biscayne aquifer range from 20 to 25 percent. Because this is a measure of the volume of porous solution features drained when heads declined, this range of values might actually be more representative of the effective porosity in zones with solution porosity as are found in the Upper Floridan aquifer.

Water Quality

Samples of the Grossman well discharge were obtained on numerous occasions over a period of 40 years, and results of analyses by the USGS for temperature, specific conductance, chloride, sulfate, and dissolved solids are listed in table 2. The 1944 sulfate concentration is given as listed in the laboratory report but is evidently in error, as are the chloride and sulfate concentrations of August 1974. The 1944 laboratory chloride and dissolved-solids concentrations seem too low and are probably in error, unless some process causing change in the quality of the flowing water, such as upconing of deeper, more saline water, occurred in the 18.5 years before the next sample was obtained. Another possibility is that these samples collected 2 months ago after drilling was discontinued could have been contaminated by non-native fluids introduced into the formation during drilling. However, the specific conductance reported in 1944 is consistent with values reported in subsequent years. With the questionable concentrations excluded, the mean specific conductance was 4,760 µS/cm and the mean concentrations of chloride, sulfate, dissolved solids (residue upon evaporation to dryness at 180°C), and dissolved solids (sum of dissolved constituents) were 1,230, 470, 3,070, and 2,900 mg/L, respectively. These values are about 6.5 to 8.5 percent of those characteristic of seawater.

In drilling waste disposal wells to the Boulder Zone along the southeastern coast, a transition from brackish water quality to water with a salinity characteristic of seawater often occurs within a depth interval of about 100 ft centered between 1,200 and 2,000 ft (Reese, 1995). The depth to the transition zone increases inland and north-northwest toward an area of recharge for the Upper Floridan aquifer in central Florida.

Table 2. Analyses for selected constituents in the artesian discharge from the Grossman well

[U.S. Geological Survey Laboratory analyses unless otherwise noted. All concentrations in milligrams per liter except where noted. °C indicates degrees Celsius. Dashes indicate no data]

Date	Temperature (°C)	Specific conductance (microsiemens per centimeter)	Chloride	Sulfate	Dissolved solids Residue at 180 °C	Sum of constituents
12-23-44	--	--	1,150[1]	--	--	--
12-23-44	24.5	4,490	970	48[2]	1,810	--
06-14-63	--	4,300	1,300	490	3,360	3,110
04-29-64	24.5	4,780	1,200	480	3,000	2,850
07-06-65	24.5	--	1,225[3]	--	--	--
10-26-68	24.5	4,800	1,210[3]	--	--	--
10-18-73	--	4,450	1,200	430	--	2,860
11-15-73	25.5	5,070	1,300	470	--	2,950
01-29-74	--	5,170	1,300	530	--	--
04-20-74	--	5,000	1,300	490	--	--
08-22-74	25.0	4,500	13[2]	8.9[2]	--	--
01-27-75	--	5,100	1,200	460	--	--
09-17-75	24.0	4,770	1,200	420	2,950	2,790
10-27-78	--	4,700	1,200	480	2,980	2,860
Mean		4,760	1,230	470	3,070	2,900

[1]Rough field titration (G.G. Parker, U.S. Geological Survey, written commun., 1944).
[2]As listed in the laboratory report, but probably in error. Not included in mean.
[3]Field titration by U.S. Geological Survey Miami office personnel.

Regional Flow System and Ambient Pressures

The Upper Floridan aquifer is recharged in an area centered around Polk City (fig. 4) in the central part of peninsular Florida (Meyer, 1989a). The peninsula is part of a broader carbonate platform known as the Florida Plateau, the surface of which is submerged at shallow depths for some distance to the east, west, and south of the land area shown in figure 4. At the edges of the plateau, ocean depths greatly increase. On the Atlantic side, the edge of the platform is also the edge of the Atlantic Shelf.

Hydraulic head in the Upper Floridan aquifer decreases from the area of recharge toward the edges of the peninsula to the east, south, and west. From figure 4, which is a generalization based on data from scattered locations, the direction of flow in the aquifer is approximately southeast at the site of the Grossman well, and the local head in 1980 would have been about 47 ft. Some documented head measurements in shut-in wells open to the Upper Floridan aquifer of the type that were used as data for figure 4 are subject to error because they are from short-cased wells with long open intervals and represent the effect of multiple zones of varying fluid density. However, only data believed to be representative of the Upper Floridan aquifer were used by Meyer (1989a) in preparing this figure. Meyer (1989a) found no evidence of appreciable vertical head gradient within the Floridan aquifer system when increases of salinity with depth were taken into consideration.

Predevelopment heads in the Upper Floridan aquifer before major withdrawals for public supply in central and western Florida and minor withdrawals for irrigation and recreational use in southern Florida are not available from southern Florida. Heads measured at a few scattered locations in southern Florida since 1961 by the USGS show inconsistent variations that do not seem to indicate a regional trend.

There are no records indicating that the Grossman well was ever shut in for a pressure test. F.W. Meyer (U.S. Geological Survey, retired, oral commun., 1990) stated that he started to shut in the well for a pressure measurement but abandoned the effort when the pressure did not increase as much as anticipated, a development which he attributed to leakage through the corroded casing behind the 8-in. liner.

FLOW RATE OF THE GROSSMAN WELL

To establish the total flow rate of water leaving the flow zone occurring at 1,200 ft within the Upper Floridan aquifer in the Grossman well and ultimately entering the Biscayne aquifer and to determine the variation of the flow rate with time, an analysis was required to reconcile the dissimilar reported flow rates at land surface and to account for changes in the condition of the well and its partial reconditioning for functional and ornamental purposes. A method was needed to quantitatively estimate the rate of artesian flow from a well of specified design and from an aquifer with specified hydraulic properties. Two methods were used. The first method was based on the constant drawdown analysis presented by Lohman (1979). The second was a more powerful, but also more cumbersome, numerical simulation method. The application of the two methods is described in the remaining sections of this report.

Analytical Approach — Constant Drawdown Computations

The constant drawdown method, developed by Jacob and Lohman (1952) and presented by Lohman (1979, p. 23), estimates aquifer transmissivity from the variation of the discharge rate following the uncapping of an artesian well:

$$T = \frac{2.30 \ Q(t)}{4\pi \ s_w} \ \log_{10} \ \frac{2.25 \ Tt}{r_w^2 \ S} \tag{2}$$
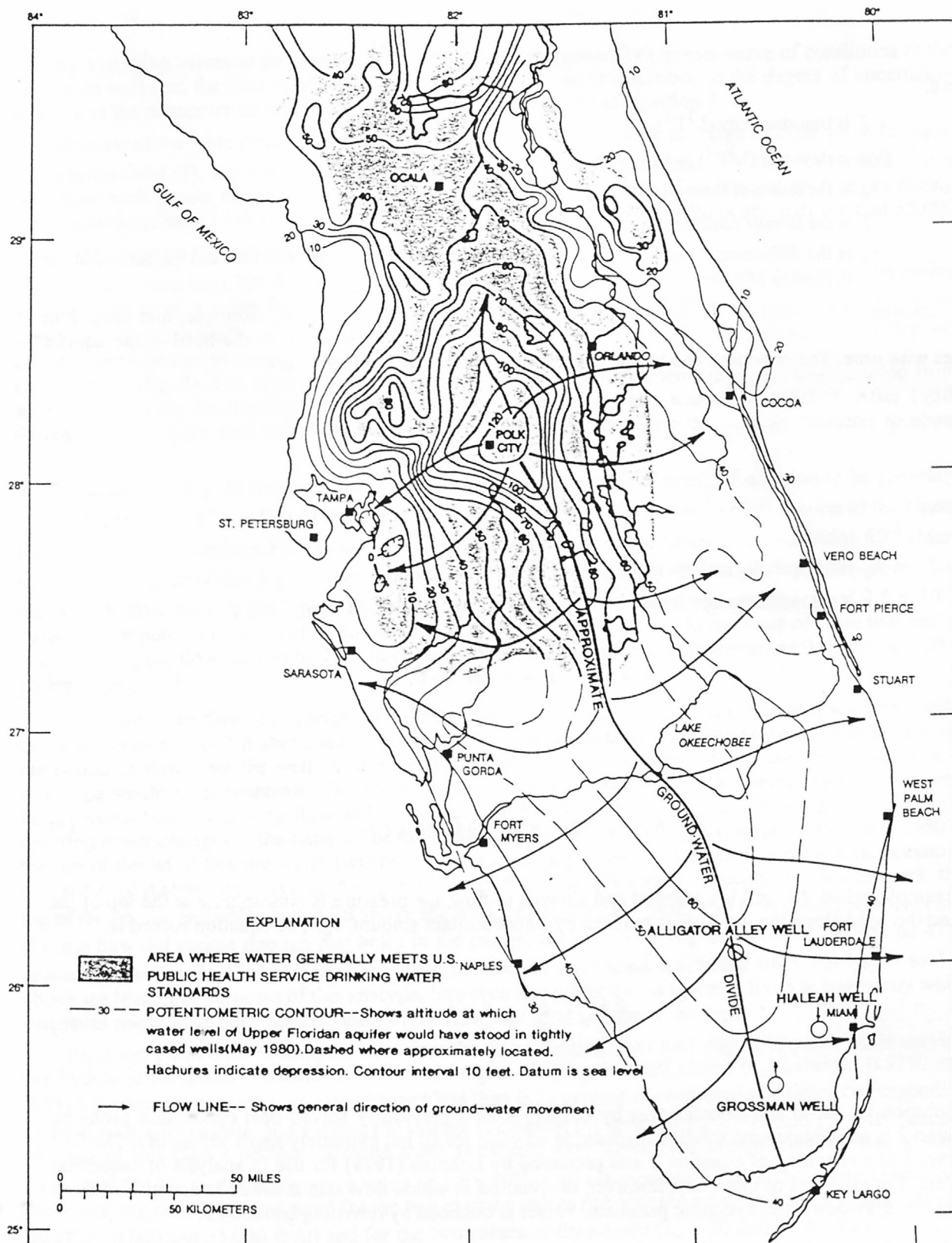
Figure 4. The potentiometric surface of the Upper Floridan aquifer in peninsular Florida in May 1980 and the area where aquifer water is potable (from Meyer, 1989a). Also shown are selected wells providing data used in this study.

where,

T is transmissivity (L^2T^{-1}),

$Q(t)$ is flow rate (L^3T^{-1}) as a function of time (t),

r_w is the radius of the well (L),

S is the storage coefficient (dimensionless), and

s_w is the difference in head (L) between the point of discharge on the wellhead and the aquifer at the beginning of flow.

Equation 2 is based on the assumptions that the aquifer is homogeneous, isotropic, and infinite in extent. The term "constant drawdown" applied to this equation is misleading, as the head in the aquifer varies with time. The boundary conditions cited by Jacob and Lohman (1952) are:

$$h = h_o \ \ for \ \ t = 0 \ \ and \ \ 0 < r < \infty \tag{3}$$

where,

h is head,

h_o is the spatially uniform initial value of head in the aquifer,

r is the radial distance from the well, and

$$h \rightarrow h_o \ \ as \ \ r \rightarrow \infty \ \ for \ \ t > 0, \tag{4}$$

and

$$h = h_o - s_w \ \ for \ \ r = r_w \ \ and \ \ for \ \ t > 0. \tag{5}$$

For example, when the well is uncapped and allowed to flow, the pressure is atmospheric at the top of the jet, and the head along the wellbore is reduced by a concomitant amount, s_w. The equation solved is:

$$\frac{\partial^2 h}{\partial r^2} + \frac{1}{r} \frac{\partial h}{\partial r} = \frac{S}{T} \frac{\partial h}{\partial t} \tag{6}$$

and solution by Jacob and Lohman was by analogy with a previously solved heat conduction problem. Equation 2 is an approximate solution assumed to be valid for all but extremely small values of t. The first difference of a rewriting of equation 2 was presented by Lohman (1979) for use in analysis of sequential rate data. For purposes of this study, however, the relation in which flow rate at time t is determined from the aquifer drawdown and hydraulic parameter values is obtained by rewriting equation 2:

$$Q(t) = \left(log_{10} \frac{2.25 \ T \ t}{r_w^2 \ S} \right)^{-1} \frac{4\pi \ s_w \ T}{2.30}. \tag{7}$$

By assigning values to parameters in equation 7 that are generally representative of conditions at the Grossman well site, the time variation of flow from the well can be described to the degree of uncertainty inherent in the parameter estimates and subject to the assumptions of equation 7.

Because of the wide variation in measured values, it was not possible to assign a single value for aquifer transmissivity (T), and a range of likely values was postulated. A transmissivity value of 10,000 ft^2/d is consistent with results of the aquifer test at Hialeah, but the simulation of water-quality changes during the recovery cycles at Hialeah and transmissivity estimates from the Turkey Point site indicate that 35,000 ft^2/d might be more representative.

Because more than 700 ft of open borehole of varying diameter lie between the bottom of the casing and the flow zone, a value for the radius (r_w) should be selected that is generally representative of the diameter of both the wellbore and casing. Before 1958, 485 to 500 ft of 12-in. casing were grouted in place above a borehole that averaged 14 in. in diameter for about 460 ft, 12 in. for another 200 ft, and 15 in. for another 60 ft (fig. 3). The bottom section had numerous major enlargements that probably resulted from secondary porosity development. The average casing-borehole diameter was about 1.075 ft. After 1958, the upper 80 ft of the well was 8-in. polyvinyl chloride tubing, changing the average diameter to about 1.050 ft.

Lohman (1979, p. 8) estimates the aquifer storage coefficient (S) of confined aquifers to be generally of the order of 1 x 10^{-6} per foot of thickness. Based on a computer simulation, a 60-ft section of the Upper Floridan aquifer in northern Florida was estimated to have a storage coefficient of about 2.75 x 10^{-4} (Merritt, 1984), suggesting that 5 x 10^{-6} per foot or greater might be more representative. In southeastern Florida, a 12-ft flow zone in the Upper Floridan aquifer had an estimated storage coefficient of 7.8 x 10^{-5}, based on computer simulation of an aquifer test (Merritt, 1995c, in press). On the basis of these two analyses, assuming the flow zone to be 25 ft thick, the storage coefficient at the Grossman well might be 1.20 x 10^{-4} to 1.60 x 10^{-4}.

The head in the flow zone (height of a fluid column in a standpipe of sufficient height) was previously estimated to be about 47 ft above sea level (fig. 4) and is assumed to be unaffected by other stresses during the period of flow from the well. At this height, the pressure at the surface of a static fluid column in a standpipe would be atmospheric. The constant drawdown is the difference in the initial height of a static fluid column before releasing flow and the subsequent height of the jet of water flowing from the well, ignoring minor changes in the latter as the flow rate changes. Before the improvements of the early 1950's, the top of the jet of flowing water (where the water column pressure was equal to atmospheric pressure) was about 10 ft above sea level, so that s_w = 47 ft – 10 ft = 37 ft. After improvements to create the park, the top of the jet was probably about 17 ft above sea level, and s_w = 47 ft – 17 ft = 30 ft. It should be noted that, if some flow did escape through rust holes in the casing, the corresponding drawdown s_w would be 47 ft minus the head in the Biscayne aquifer, which historically has varied seasonally from 1 ft below to 7 ft above sea level. For purposes of this analysis, however, this possibility is ignored. Even in later years when rust holes might have been present, most of the flow was from the top of the well.

Equation 7 was used to compute flow rates for times ranging from 0.01 day to 40 years while varying the hydraulic parameter estimates in the cited ranges. Varying the well radius (r_w) between 0.5250 and 0.5375 ft caused computed flow-rate changes less than 0.25 percent at computational times corresponding to the flow measurements, and varying the storage coefficient (S) between 1.2 x 10^{-4} and 1.6 x 10^{-4} caused changes of less than 2 percent. Because these variations were considered insignificant, r_w was given a value of 0.5250 ft, and S was given a value of 1.5 x 10^{-4} in all subsequent computations using equation 7. Time-varying rates were computed for the two extreme values defining the estimated range of transmissivity (T = 10,000 and 35,000 ft^2/d) and for the two values of drawdown (s_w = 30 and 37 ft). Results (fig. 5) indicate that regardless of the total value of aquifer transmissivity or constant drawdown, the major part of the computed decrease in flow occurs in the first month. In fact, an analysis of the numerical values shows that about 60 percent of the computed flow-rate decrease that occurs in 40 years takes place during the first week. However, the computed flow rate continues to decrease slightly even after 40 years.
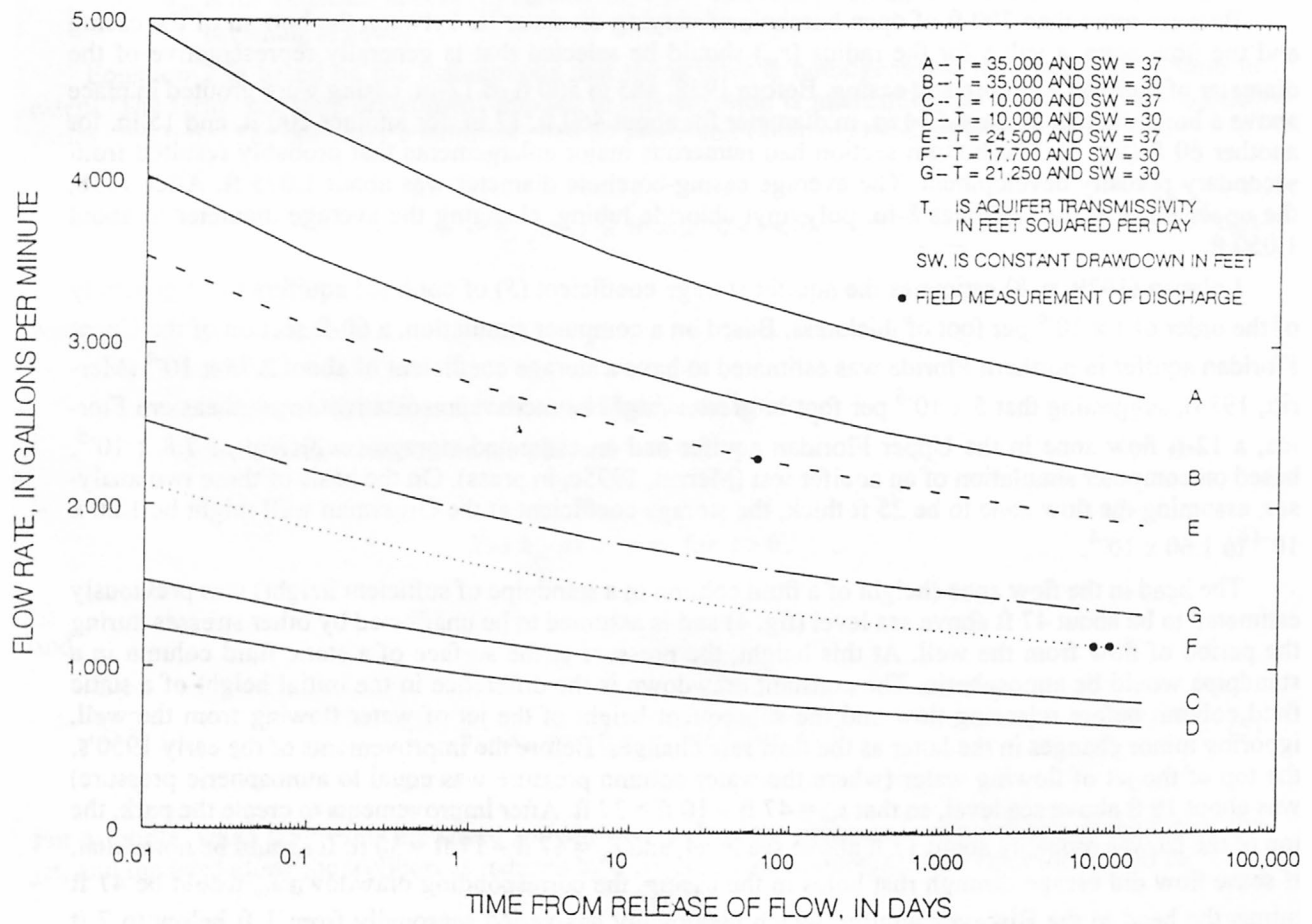
Figure 5. Rates of flow from the Grossman well in central Dade County as a function of time by constant-drawdown formula for various assumed aquifer transmissivities.

When Hoy visited the site of the flowing well in 1944 about 2 months after drilling had been discontinued and measured a flow rate of 2,350 gal/min, the length of the period of artesian flow (t in eq. 7) was probably about 2 months, and the controlling head difference was the 37-ft constant drawdown (s_w). Further calculations with equation 7 reveal that the rate cited by Hoy in 1944 can be approximated by assuming an aquifer transmissivity of about 24,500 ft^2/d (fig. 5, curve E).

The constant drawdown would have changed to 30 ft before 1954, when relandscaping raised the top of the well. When the USGS measured the flow to be 1,170 gal/min in 1964, 1965, and 1969, the well had flowed under a constant drawdown of 37 ft for as long as 10 years and under a constant drawdown of about 30 ft for at least 10, 11, and 15 years prior to each measurement, respectively. Equation 7 does not compute the flow rate for a drawdown (s_w) that is not constant in time. However, figure 5 shows flow rates computed with equation 7 to be nearly constant from 10 to 20 years after the beginning of flow. Therefore, the rates of 1,170 gal/min cited by Howard Klein, F.A. Kohout, and F.W. Meyer in the 1960's can be approximated by assuming the constant drawdown to be 30 ft since 1944, and the rates are represented with equation 7 by assuming an aquifer transmissivity of about 17,700 ft^2/d (fig. 5, curve F). Assuming those measured rates to be too low by 20 percent because of leakage around the 8-in. liner, the estimated 1960's rate of 1,400 gal/min can be computed using a transmissivity of 21,250 ft^2/d (fig. 5, curve G).

When a transmissivity of 24,500 ft^2/d is used in equation 7, however, the computed 1960's flow rate would be about 1,600 gal/min, 200 gal/min more than the actual (assumed) total 1960's flow rate and 430 gal/min greater than the actual measured flow rate. When a transmissivity of 21,250 ft^2/d is used, the 1944 flow rate is computed to be 2,050 gal/min, 300 gal/min less than the measured rate. When a transmissivity of 17,700 ft^2/d is assumed, the flow rate is calculated to have been about 1,725 gal/min when Hoy visited the site in 1944.

Based on the application of equation 7, the rate reported in 1944 could be consistent with the rates measured in the 1960's only if leakage around the 8-in. liner amounted to 35 percent of the flow in the 1960's or one or more of the cited measured rates were partly influenced by measurement error. The analysis is probably not affected by the partial capping of the well sometime in the late 1940's or early 1950's, or the possibility that the well flow may have been stopped for installation of the 8-in. liner in 1958 following losses from leakage through the casing that might have exceeded 20 percent of the total flow. The fact that computed flow rates approximately stabilize after 1 month indicates that the effect of temporary stoppages of flow or temporary changes in the drawdown would be relatively insignificant over time.

The cited sources of error might fully explain the discrepancy between the measured and calculated flow rates. However, additional reservations are that the analyses using equation 7 are based on the assumption that the aquifer is of infinite extent, which is in contradiction to the fact that the Upper Floridan aquifer probably crops out under the sea at the edge of the Atlantic Shelf about 32 mi east of the well, and may also be recharged by leakage through vertically adjacent confining layers. Furthermore, friction losses in the wellbore and casing are not considered.

These difficulties indicate that there can be appreciable uncertainty in estimates made with the constant drawdown formula. If boundaries of the finite aquifer can be characterized as representing continuous recharge or discharge governed by head conditions independent of those in the aquifer, water levels interior to the region may show substantially different trends after drawdowns reach the boundaries than if the aquifer were infinite in extent. Friction losses would be greater at early higher flow rates and would, therefore, tend to reduce the rate of variation with time. The degree to which such factors affect the calculation of flow rate as a function of time will be assessed in the subsequent sections describing the numerical simulation analysis.

Numerical Simulation Approach

A realistic representation of the variable drawdown and its relation to the rate of flow required consideration of several boundaries at varying distances from the well. Available analytical solutions that accounted for boundaries were considered not to have sufficient generality for such a representation.

Therefore, a more accurate calculation of flow from the Grossman well required use of a computer code for numerical simulation of the ground-water flow system. Implementation of this approach required the development of a generalized model of the flow system in the Upper Floridan aquifer in a large region that included the study area.

Description of the Simulator

The selected simulator, the SWIP code, was developed by INTERCOMP Resource Development and Engineering, Inc. (1976), under sponsorship of the USGS. The code was later revised for the USGS by INTERA Environmental Consultants, Inc. (1979). Despite its intended use for waste-injection problems, it received wider use within the USGS as a general-purpose, three-dimensional simulator of solute and thermal-energy transport in ground water. Outside the agency, the SWIP code has been adapted for special purposes by various public and private organizations. Within the agency, the code was used as a resource in the development of the three-dimensional integrated finite-difference code HST3D (Kipp, 1987), which has a similar structure but somewhat wider range of applicability to field problems of flow and solute and thermal transport.

In the SWIP code, absolute pressure is the solution variable of the flow equation, and the model accounts for fluid density and viscosity dependence on temporal and spatial variations of pressure, temperature, and solute concentration. The solution variables are expressed in residual notation (for example, $p^{n+1} = p^n + \delta p^{n+1}$, where p^n is the value at the end of the previous timestep). This has the advantage of reducing the effect of roundoff error. Solution of equations is by standard finite-difference techniques. The aquifer simulated can be fully confined or have a free surface, and the equations can be solved in Cartesian or cylindrical coordinates. In the work documented in this report, only the flow equation was solved.

The mathematical equations solved by the model are based on the assumption of porous media flow. In application to secondary-porosity media, such as the dissolved carbonates of the study area, the porous media assumption is assumed to apply in a spatially averaged sense; that is, the scale of the model is sufficiently large that local heterogeneities in the rock matrix do not need explicit representation and rock properties can be adequately represented by spatially averaged parameters.

Numerous modifications have been made to the SWIP code by the author to facilitate its application to various problems. Those modifications that are revisions or extensions of the mathematical procedures of the 1979 version of SWIP have been coded as options so that the original solution methodology remains available.

Aspects of the model that are of special interest to this study are the procedures used in the SWIP code to represent flow between the fluid-filled borehole and the heterogeneously layered sequence of aquifers and confining strata (the wellbore simulator) and to describe loss of momentum and thermal energy within the wellbore and casing caused by friction and thermal conduction (the well-riser simulator). These procedures are described in detail in the next section.

Wellbore and Well-Riser Representations

In the SWIP code, the wellbore representational logic is encoded within the production simulator (subroutine PROD) and the well-riser representational logic within the well-riser simulator (subroutine WELLB). The documentation of the code (INTERCOMP Resource Development and Engineering, Inc., 1976) includes a brief description of this logic. The methodology was modified for use by Kipp (1987) in the HST3D code, and the HST3D model documentation includes a more-detailed description of the wellbore and well-riser representation methods than provided in the earlier documentation of SWIP.

The overall rate of flow between the wellbore and aquifer is determined by a form of the Theim equation for steady-state flow to a well (Lohman, 1979, p 11-12), in which a "well index" (INTERCOMP Resource Development and Engineering, Inc., 1976; Kipp, 1987, p. 33) is used to relate the pressure gradient between the wellbore and aquifer to a flow rate between the wellbore and aquifer. The well index (WI) is computed as:

$$WI = \frac{2\pi T}{ln\left(\dfrac{r}{r_w}\right)} \tag{8}$$

where,

$T = \displaystyle\sum_{1}^{n} K_i \Delta z_i$ is the transmissivity (L^2/T),

K_i is the hydraulic conductivity (L/T) of layer i,

Δz_i is the thickness (L) of layer i,

n is the number of layers contributing flow,

r is the radius (L) to the midpoint of the grid cell containing the well, and

r_w is the radius of the well (L).

The production simulator of the SWIP code provides various options for assigning relative rates of flow between the wellbore and the various layers of the model grid that may represent a series of aquifers and confining layers. The various procedures differ in complexity and sophistication.

The simplest procedure (type 1 well) is one in which relative flow rates to or from the various layers of cells are entirely prespecified by allocation factors that are usually based on the thicknesses and relative hydraulic conductivities of the layers. Skin effects, such as clogging of the formation during injection, which vary in degree between layers, can also be accounted for by specifying appropriate values for the allocation factors. The type 1 well approach is appropriate when the head relation between the well and aquifer does not change appreciably with depth. This condition might not apply when injection of water with density appreciably different from that of the receiving aquifer occurs over a substantial depth interval of the aquifer. It also might not apply during injection or production (withdrawal) when the thickness of the formation is sufficient for an appreciable native-water density variation to occur by virtue of vertical changes in aquifer water temperature or mineralization.

When the type 2 well option is used, the production simulator performs a layer-by-layer check of the pressure relation between the open-hole part of the well and the background pressure in the aquifer. If a reversal of the pressure gradient needed for injection (or withdrawal) occurs, the allocation factor for the layer is set equal to zero. Whether or not any layers are made noncontributory, the allocation factors for the various layers are recomputed based on the relative pressure relations.

Type 3 well calculations proceed as type 2 calculations in the production case, except that when well-riser calculations are not performed, the borehole pressure, computed from the specified well rate and the well index, is compared with a user-specified bottom hole pressure and the larger of the two used to compute allocation factors. If well-riser calculations are performed, the resulting pressure at the top of the well is compared with a user-specified pressure value. If the pressures do not match within a tolerance, an iterative procedure is implemented in which the wellbore pressure is varied until computed and specified top hole pressures agree. The well rate is computed from the wellbore pressure thus determined.

When type 4 is selected, the user-specified allocation factors are not modified, and layer flow rates are based on user-specified bottom hole pressures. Checks on the results of the wellbore calculations and possible further iterations are as for well types 2 and 3. In well types 2, 3, and 4 calculations, the well rate can be expressed explicitly or implicitly in residual form with the correction factor a function of the computed pressure change.

The well-riser model accounts for friction losses within the wellbore and casing in developing a profile of pressure as a function of depth. In production problems, a surface pressure is determined. A pressure increment for computation and the temperature difference between the top and bottom of the wellbore and casing are prespecified. In the production case, as computations proceed upward along the wellbore and casing, the head loss due to friction is computed and pressure is calculated as a function of vertical position, friction loss, and density. The latter is corrected for changes of pressure and temperature, which vary

linearly with length. The volumetric rate of flow is corrected for changes in density, and the heat loss to the formation is computed. In the final step, a linear extrapolation is used to compute the pressure at land surface.

Artesian (Type 5) Well Option

The type 3 well option at first appeared to provide the features needed for computation of artesian flows as a function of the time-varying head in the aquifer. However, the use of the iteration scheme as coded was not possible because the iteration procedure could not be made to work in the application problem of this study. Because none of the other cited well representations appeared to be capable of simulating the relation between the head in the aquifer and the flow rate of the well, a special iterative method was encoded in the production simulator as a type 5 well option to work in conjunction with the well-riser model. The method generally replicated the logic of the type 3 well option but used an iterative scheme that worked well for the application problem of this study.

The iterative procedure, performed within each model timestep, is based on the comparison of a computed surface pressure with a user-specified surface pressure. Depending on whether the surface pressure computed by the well-riser model exceeds or is less than the user-specified pressure, the borehole pressure is adjusted by an amount that is a function of the change in wellbore friction loss and surface pressure difference between iterations, modified by a user-specified damping factor. The adjustment is expressed mathematically as follows:

$$P_b^n = P_b^{n-1} \pm \frac{|F^{n-1} - F^{n-2}| \cdot C}{D_a} \tag{9}$$

where,

C is the lesser of 1.0 and C', where $C' = \dfrac{|P_T - P_{TC}^{n-1}|}{|P_T - P_{TC}^{n-2}|}$

P_b is the borehole pressure in the artesian zone (pound per square inch),

P_T is the user-specified pressure at land surface (pound per square inch),

P_{TC} is the computed casing pressure at land surface (pound per square inch),

F is the friction pressure loss in the well (pound per square inch) computed by the well-riser model,

D_a is the damping factor, and

$n, n-1, n-2$ represent the present and two previous iterations.

Whether the plus or minus sign is assigned in equation 7 depends on whether the previous computed land-surface casing pressure is greater or less than the user-specified value. In the first iteration, P_b^{n-1} is set equal to the average of the values computed in the last two iterations of the previous timestep. Using the average of the last two values avoids errors arising from oscillations that can occur in sequential iterations when computed friction losses are high and vary greatly with small changes of the estimated borehole pressure.

Convergence was considered to be achieved when the flow rate, estimated from the aquifer to wellbore pressure gradient using the specified well index, was different from that of the previous iteration by less than 1 gal/min. Computation of aquifer pressures for the timestep then proceeded normally. The iterative well-rate calculation procedure is explicit in that the aquifer pressure value used to compute a surface pressure is the value from the previous timestep.

The selection of damping factor values in each application is important, as the change in friction loss might grow or become so small that apparent convergence of the iterations is achieved without achieving a match of computed and user-specified surface pressures. In practice, damping factors had to be determined

for each new set of aquifer hydraulic parameters and needed to be varied with increasing simulation time in each computation. Experience and intuition rather than an automatic procedure were used in selecting a set of damping factors for a simulation run. If a set of values proved to be unsuccessful, the simulation run was repeated with a new set of values. Damping factors as low as 0.02 and as high as 200 were used in the various computations.

Modifications were also made to the well-riser model to permit the specification of wellbore diameter changes at specified depths. This allowed a more realistic representation of the spatially varying borehole friction loss.

Design of the Regional Model of Flow in the Upper Floridan Aquifer

To account realistically for the influence of recharge from aquifer boundaries, the type 5 well option calculations needed to be performed within the framework of a model of the aquifer within those boundaries that represented heads to a reasonable degree of accuracy. Results of a previous study (Merritt, 1995c, in press) indicate that the flow zone at the site of the Grossman well (S-524) near the erosional surface of Eocene rocks correlates with flow zones in rocks of the same age at other locations scattered along the east coast of peninsular Florida from St. Lucie County to the middle Florida Keys. Meyer (1989a, p. 14) described a similar flow zone near the surface of Eocene rocks in the Alligator Alley test well (G-2296) in the central-southern part of the Florida Peninsula (fig. 6).

A highly detailed consideration of data describing the Upper Floridan aquifer in central and southwestern Florida was beyond the scope of the present study. The flow zone near the surface of Eocene rocks at Grossman Hammock was considered (for modeling purposes) as a zone of uniform thickness (25 ft) and uniform hydraulic characteristics that extends west, east, and south to the edges of the Florida Plateau and northward into central Florida where it is known to have a surficial outcrop in the recharge area (fig. 4). This flow zone will henceforth be identified as the Upper Floridan aquifer as a generalization adopted for modeling purposes, and the possible local influence of lower flow zones of unknown geographical extent will be considered to be included in the representation of the Upper Floridan aquifer as a single layer. This procedure is consistent with the lack of known vertical head gradients of appreciable magnitude within the Floridan aquifer system (Meyer, 1989a).

The eastern edge of the Florida Plateau is approximately coincidental with the land coast of the peninsula (fig. 6), but the shallow surface of the plateau extends nearly 200 mi west of the western coastline and extends southward beyond the Florida Keys. Figure 6 shows contours of the 650 and 1,300-ft depths (below sea level) of the undersea surface of the plateau (ocean bottom) east, south, and west of the southern Florida Peninsula (Uchupi, 1966). Depths on the east coast and south of the Florida Keys are within 100 ft a short distance upslope of the 650-ft contour. The upslope on the western side of the peninsula is less abrupt. The 1,300-ft contour is used in this report to indicate the edge of the Continental Shelf (the eastern, western, and southern edges of the Florida Plateau). Generally, ocean depths increase rapidly to 10,000 ft beyond the 1,300-ft contour to the west, but the sea bottom begins to flatten out beyond the 1,300-ft contour in the Straits of Florida south and east of the peninsula. The flow zone at Grossman Hammock is at about 1,165 to 1,190 ft below sea level. The apparently correlative flow zones at widely scattered locations on the east coast vary in depth from 900 to 1,200 ft below sea level. The top of the Eocene on the west coast is shown to be at about 1,400 ft below sea level by Meyer (1989a, fig. 3), and it might be still deeper farther to the west. Nevertheless, a generalization accepted for the modeling analysis is that the depth of the flow zone is uniform throughout the southern part of the Florida Plateau at its depth at Grossman Hammock and that the flow zone has a subsea outcrop on the eastern, western, and southern edges of the plateau. The actual variation of flow-zone depth was partly accounted for in the assignment of hydrostatic boundary conditions.

A single-layer Cartesian grid with dimension 19 x 18 was superimposed on part of the Upper Floridan aquifer as shown in figure 6. Grid cell dimensions ranged from 100 x 100 ft in grid cell (11,10) containing the flowing Grossman well to 416,000 by 330,000 ft in grid cell (1,1). The north-south axis of the grid was oriented at an angle of 22 degrees from the true north-south direction, so that it would approximately parallel the flow divide along the southern part of the Florida Peninsula indicated by Meyer (fig. 4). This procedure greatly facilitated some applications of the model. Eastern, western, and southern grid boundaries
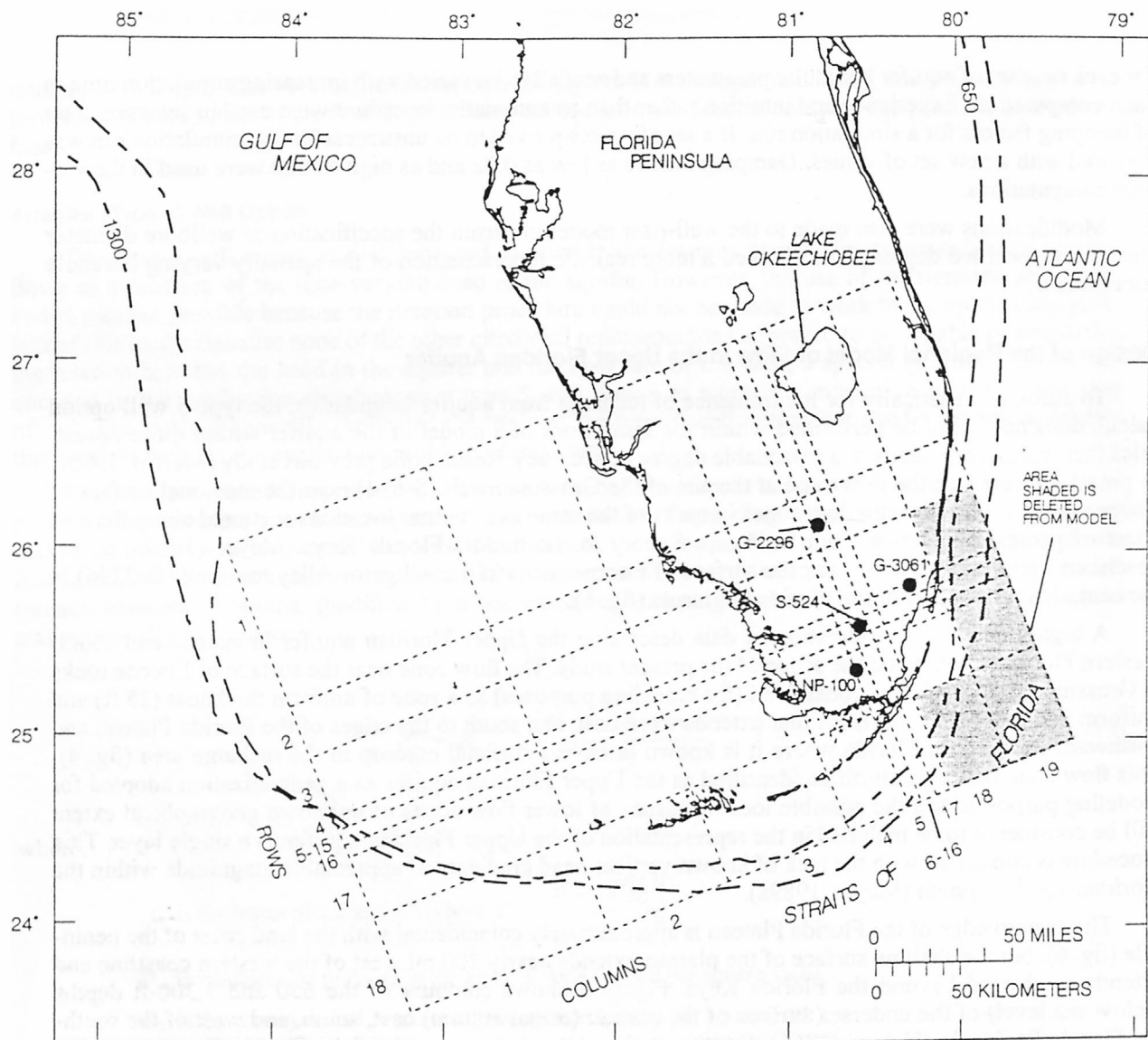
Figure 6. Grid design of the Upper Floridan aquifer model superimposed on sketch showing the extent of the Florida Plateau.

approximately corresponded to the 650- to 1,300-ft depth contours on the Florida Plateau where the flow zone would be expected to have a subsea outcrop based on the assumption of approximately uniform depth. Where the grid extended beyond the assumed undersea outcrop in the southeastern corner, grid cells 3 to 18 in column 19 and grid cell (18,18) were made inactive (deleted from the computations), and boundary conditions were specified at the edges of the adjacent active cells.

Where boundaries correspond to the assumed undersea outcrop, specified pressure values were assigned to the vertical centers of outer edges of grid cells representing a 25-ft thick aquifer, the top of which was at 1,172 ft below sea level (1,180 ft below land surface). A time-invariant pressure of 541.14 lb/in^2 was specified, representing the pressure of 1,184.5 ft of seawater (density of 64.0 lb/ft^3). The equivalent freshwater head at sea level is about 32 ft. Assigned pressures along the northern boundary are 541.14 lb/in^2 in the first three columns. Higher values assigned over the peninsula in columns 4 to 19 are based on the heads estimated by Meyer (1989a). The highest estimated pressure along the northern boundary is 557.54 lb/in^2 in columns 8 to 14, which represents an estimated head of 70 ft.

Measured dissolved-solids concentrations near the top of the Upper Floridan aquifer ranged from 1,800 to 6,000 mg/L throughout the modeled area. In the initial model runs, the assignment of an areally uniform density value was considered to be an acceptable approximation. In later runs, the density of water in the aquifer was represented as varying spatially. Computed pressures were converted to freshwater heads for comparison with those measured in the field.

The field data were generally measured at land surface with pressure transducers having readouts in feet of freshwater head, and adjusted to sea level datum. However, the water columns in which the measurements were made generally contained brackish water from the artesian flow zones measured. Measurements at land surface of head in a hypothetical flow zone occurring at 1,150 ft below land surface containing water of 3,000 mg/L dissolved solids (similar to the flow zone of the Grossman well) might be higher by 2 or 3 ft if the well were actually filled with freshwater, as assumed by use of the standard measuring instruments. Generally, this degree of error is comparable to the scatter usually observed in time series of Upper Floridan aquifer head measurements that can be attributed to a variety of causes, not the least of which is instrument calibration error.

Calibration of the Regional Model of Flow in the Upper Floridan Aquifer

The conventional technique for obtaining an approximate calibration of this highly generalized model is to compare computed and measured pressures at locations where the latter are available. One such location is at G-2296, the Alligator Alley test well (Meyer, 1989a), near the center of grid cell (11,2) as shown in figure 6. The head measured in packer tests of relatively small intervals that approximately coincided with the top of Eocene rocks ranged from 56.6 to 58.8 ft.

Because a pressure measurement is not available from the Grossman well (S-524), the local head in the Upper Floridan aquifer is assumed to be 47 ft, as estimated by Meyer (1989a), for calibration purposes. Heads of about 42 ft were measured at the injection well (fig. 6, G-3061) and monitor well used for testing injection, storage, and recovery of freshwater in the Hialeah well field, a location that would approximately correspond to the center of grid cell (18,4) in figure 6. Another location where artesian heads have been measured on several occasions is at a well drilled for water supply in Everglades National Park. The well, NP-100, is on the western side of grid cell (5,17) as shown in figure 6. Measured heads generally ranged from 41.5 to 42.6 ft. Because of the short casing, these heads could be lower than those of the Upper Floridan aquifer if the well were also open to overlying flow zones of lower heads, but whether such higher flow zones exist at this location is not known.

The Upper Floridan aquifer model was run without stresses for a simulation time of 1×10^7 days, by which time hydraulic equilibrium had become established. In the initial run to equilibrium conditions, hydraulic conductivity was considered to be uniform and isotropic at 800 ft/d (the hydraulic conductivity estimated for the flow zone at the Hialeah well site [G-3061]). The resulting head distribution (fig. 7, case 1), contoured using ARC/INFO software (the ARCTIN contouring package), was substantially different from the distribution estimated by Meyer (1989a) (fig. 4) that indicated a region of higher heads extending

Figure 7. Results of several approaches to simulating predevelopment heads in the Upper Floridan aquifer.
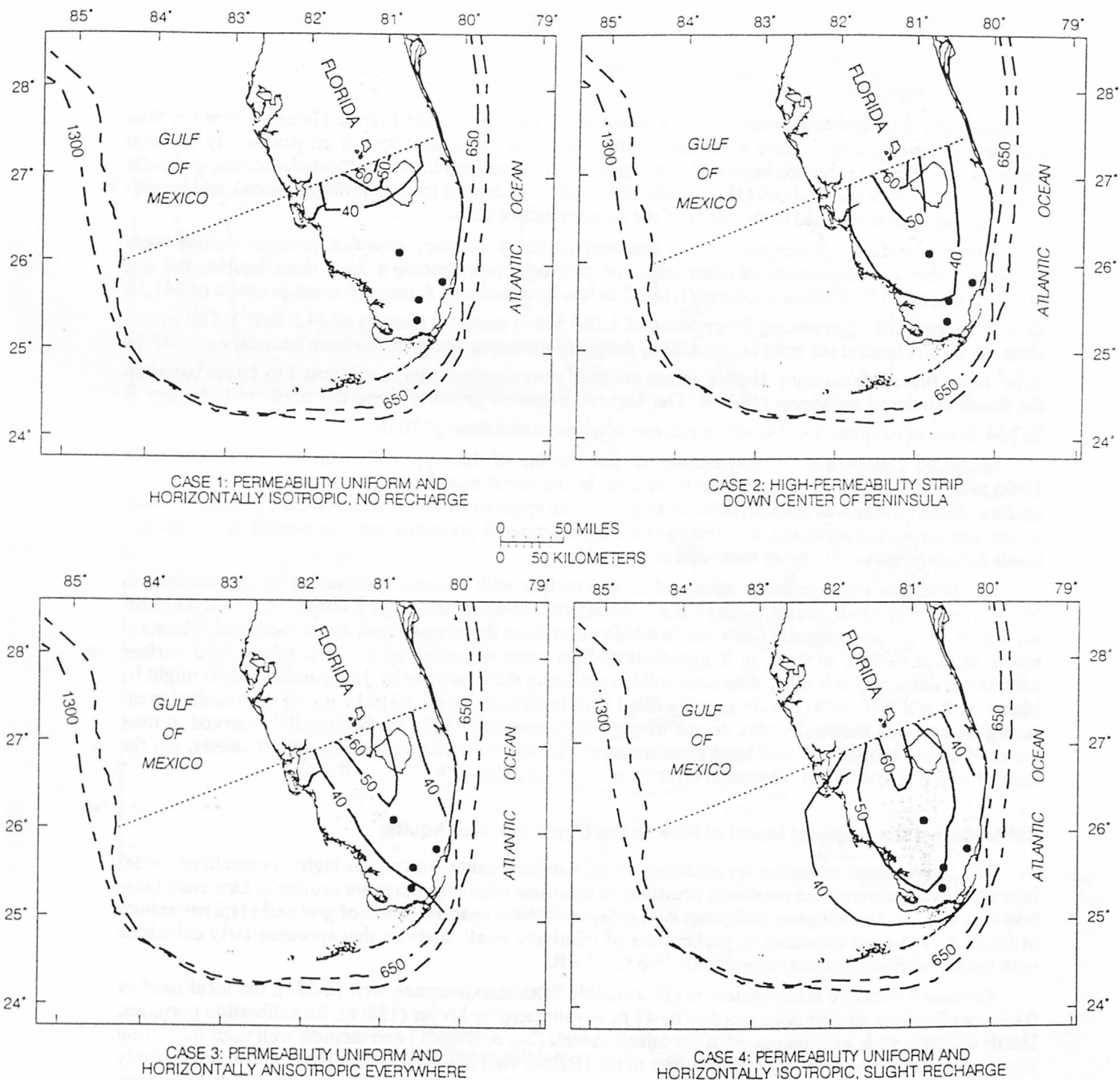
EXPLANATION

- – – – 650 – – – – SUBMARINE TOPOGRAPHIC CONTOUR - Shows depth below sea level in feet (Uchupi, 1966)

————— 50 ————— SIMULATED HEAD CONTOUR - Contour interval is 10 feet. Datum is sea level

.... NORTHERN BOUNDARY OF MODEL

• LOCATION OF UPPER FLORIDAN AQUIFER WELL - Where heads have been measured or estimated (shown in figure 6)

CASE 1: PERMEABILITY UNIFORM AND HORIZONTALLY ISOTROPIC, NO RECHARGE

CASE 2: HIGH-PERMEABILITY STRIP DOWN CENTER OF PENINSULA

CASE 3: PERMEABILITY UNIFORM AND HORIZONTALLY ANISOTROPIC EVERYWHERE

CASE 4: PERMEABILITY UNIFORM AND HORIZONTALLY ISOTROPIC, SLIGHT RECHARGE

downward toward the tip of the peninsula. In contrast, the highest heads in the simulation occurred in a small semicircular area just south of the peninsular part of the northern model boundary, and head gradients were nearly flat farther south. The simulated head at the Alligator Alley test well (G-2296) was only 35.4 ft. Heads at the Grossman well (S-524) and the Hialeah aquifer storage (G-3061) and recovery wells were even lower, and the head at node (5,17) near well NP-100 was only 32.7 ft.

The hydraulic conductivity value was decreased to 10 ft/d and increased to 5,000 ft/d in additional runs without appreciably changing the head distribution. When the lower value was used, the head gradients were somewhat steeper. The results of the simulation did not explain the higher heads measured at various locations in the Florida Peninsula.

Alternative hydraulic conductivity distributions were postulated and tested in an attempt to simulate head distributions that were more similar to the measured ones. One approach was to assume that a strip of high hydraulic conductivity extended along the axis dividing easterly and westerly flows down the center of the peninsula (fig. 4). Hydraulic conductivity was assumed to have a uniform and isotropic value of 1,500 ft/d in columns 5 to 17 and to have a uniform and isotropic value of 10 ft/d outside this strip. This procedure (fig. 7, case 2) somewhat extends the region of higher heads and steeper gradients down the center of the peninsula. However, heads at control points are still unacceptably low, 45.7 ft at the Alligator Alley test well and below 40 ft at other control points.

Another analysis tested the assumption that hydraulic conductivity was uniform and anisotropic throughout the simulation region. Hydraulic conductivity in the direction of the Y-axis of the grid, parallel to the axis dividing easterly and westerly flows, was assigned a value of 1,500 ft/d. Hydraulic conductivity in the direction of the X-axis toward the eastern and western coasts was assigned a value of 10 ft/d. The result (fig. 7, case 3) is that the region of high heads and steep gradients now extends farther down the peninsula than in either of the preceding cases. However, heads are still unacceptably low compared to measured ones. The computed head at the Alligator Alley well was only 48.3 ft.

It was evident that another approach would be required to simulate the measured head data. Because the head distribution based on measured data (fig. 4) appeared to resemble the effect of a surficial recharge process, it was postulated that some degree of recharge to the Eocene rocks occurred that was approximately restricted to the part of the Florida Plateau emergent above contemporary sea level (present-day Florida). One possible source of recharge would be a part of land surface rainfall that might percolate through intervening confining layers to the Upper Floridan aquifer. The model was run to steady state with hydraulic conductivity assumed uniform and isotropic at a value of 800 ft/d. An areally uniform recharge rate was specified and the value was varied until calibration was achieved using a value of 0.0613 in/yr. The resulting computed head distribution in the southern peninsula (fig. 7, case 4) matched measured data more closely than previous results. The head computed at the location of the Alligator Alley test well was 57.7 ft, within the range of the field measurements; and the head computed at the location of the Hialeah well was 43.6 ft, less than 2 ft higher than measured heads. The computed head at the Grossman well location was 50.2 ft, about 3 ft higher than estimated by Meyer (1989a) when drawing head contours. The head computed at node (5,17) was 47.1, about 4.5 ft higher than the head measured at well NP-100.

Although measured heads were approximately matched by this approach, it is difficult to justify the hypothesis that an artesian aquifer with heads 30 to 70 ft above land surface, overlain by more than 1,000 ft of clayey marls and layers of permeable and relatively impermeable limestone, could receive even a slight amount of rainfall recharge. The local land surface receives an average 53 in/yr of rainfall, of which the amount required to calibrate the generalized Upper Floridan aquifer model is a small part (0.12 percent). One possibility is that land-surface recharge does induce a small amount of downward seepage through the confining layer. Another possible recharge hypothesis is that the highly permeable Boulder Zone beneath the carbonates of low permeability underlying the Upper Floridan aquifer is a source of upward recharge to the Upper Floridan aquifer. To test these concepts and to determine whether the hypothesis of recharge from land surface is tenable given the hydraulic relations between the various aquifers in the vertical sequence, a simulation of the larger system of aquifers and confining layers was needed.

To test the hypothesis that Upper Floridan aquifer heads were influenced by rainfall on the peninsula, the model was extended vertically to land surface by the addition of seven layers. The uppermost three represented, in a highly generalized fashion, the surficial aquifer system in southern Florida and were assigned

boundary head values typical of average annual heads or average tidal stage at the appropriate locations. Layer 1, in which the water table occurred, was represented as having a free surface. Layers 1 and 2, having a combined thickness of 140 ft and extending 100 ft below land surface, were assigned a hydraulic conductivity of 30,000 ft/d where the Biscayne aquifer was present and 1,500 ft/d elsewhere. Layer 1 grid cells were deleted in areas of the grid covered by ocean (fig. 6). Layer 3, 200-ft thick, represented the Tamiami Formation, a semiconfining unit having low to moderate permeability, and was assigned a hydraulic conductivity of 100 ft/d. Layers 4 to 7 represented the intermediate confining unit (the Hawthorn Formation, Tampa Limestone, and the upper part of the Suwannee Limestone) as a bed 875 ft in thickness. The low value assigned to horizontal and vertical hydraulic conductivity was varied in a series of simulation runs in an attempt to simulate the elongated mound of observed heads in the Upper Floridan aquifer in peninsular Florida (now model layer 8). The maximum evapotranspiration rate was also varied to represent different rates of recharge to the uppermost layer. All layers had water of freshwater density.

The attempt to simulate the observed heads in the Upper Floridan aquifer with the eight-layer model was unsuccessful. When the value of vertical hydraulic conductivity assigned to layers 4 to 7 was low (0.01-0.00001 ft/d), a substantial head difference between the Upper Floridan aquifer and the Biscayne aquifer was simulated, but Upper Floridan aquifer heads were not influenced by any rate of recharge to the surficial aquifer and generally resembled those shown as case 1 in figure 7. When a vertical hydraulic conductivity value of 0.01 ft/d was specified, the simulated vertical pore velocity of 3.5×10^{-4} ft/d was upward and uniform vertically through all the model layers in the vicinity of the flowing well. When higher values of vertical hydraulic conductivity (10-100 ft/d) were assigned to layers 4 to 7, no appreciable steady-state head gradient could be maintained across the confining layer. Computed Upper Floridan aquifer heads were lower than in the tightly confined case and were similar to heads in the Biscayne aquifer layers.

Another attempt was made to simulate measured heads in the Upper Floridan aquifer by adding the underlying middle confining unit of the Floridan aquifer system and the Lower Floridan aquifer (the Boulder Zone). According to Miller (1986), the latter occurs throughout the region of the generalized Upper Floridan aquifer model (fig. 6). Eight additional layers were added to the model grid. Layers 9 to 15 represented the middle confining unit, assumed to be 1,800-ft thick, and a uniformly low value was assigned to horizontal and vertical hydraulic conductivity (though discrete, highly permeable layers do occur within the confining unit in parts of the region). The value assigned to hydraulic conductivity was varied in an attempt to replicate Upper Floridan aquifer heads. The Boulder Zone (layer 16) was considered to be 50-ft thick and to have a hydraulic conductivity of 1,000,000 ft/d. Because of its high hydraulic conductivity, heads in this layer were simulated as being approximately uniform and nearly equal to a uniform pressure value specified at model boundaries representing subsea outcrops. The assigned pressure value represented the weight of 3,025 ft of seawater. Layers 9 to 16 were considered to contain water of seawater density (64.0 lb/ft^3) in the assignment of hydrostatic boundary and initial conditions.

Because lower layers were represented as containing water of seawater density, density was spatially variable in the 16-layer model. However, the concentration equation was not solved (a standard SWIP code option). As a result, specified densities remained unchanged throughout the computations although their hydraulic influence on the pressure field was of significance in computing heads in the Upper Floridan aquifer (layer 8). However, solute transport by advection, dispersion, or diffusion was not simulated. This is a reasonable approximation, because the density distribution in the Floridan aquifer system is approximately at steady state except for highly localized manmade disturbances that do not affect the regional system.

Again, the attempt to replicate measured Upper Floridan aquifer heads was unsuccessful. When the hydraulic conductivity value assigned to middle confining unit layers was low, the computed Upper Floridan aquifer heads were similar to those of the previous eight-layer model and were not influenced by Boulder Zone heads. When the middle confining unit was assigned higher hydraulic conductivity values, heads in the Upper Floridan aquifer layer were increased (when the hydraulic conductivity of the intermediate confining unit was low) but tended to be areally uniform, as in the Boulder Zone. The amount of the head increase was insufficient to match measured heads anywhere on the peninsula.

Finally, a successful approach to simulating Upper Floridan aquifer heads was discovered. The 16-layer model was modified by specifying the areal variation of the vertical density distribution, again

represented as a steady-state condition. Instead of assuming that layers 1 to 8 contained freshwater and that layers 9 to 16 contained water of seawater density, the depth of the freshwater zone was varied throughout the area of the Florida Peninsula.

In reality, the freshwater and seawater zones of the Floridan aquifer system are separated by a transition zone about 50- to 100-ft thick, and the flow zone yielding water to the Grossman well lies within the upper part of the transition zone. However, ignoring the transition zone and assuming a two-density system with a sharp interface at a depth that varied areally was considered an adequate approximation for the generalized model of this study.

The depth to the seawater zone was assumed to be 2,350 ft in the center of the peninsula along the northern model boundary (columns 5-17) and to decrease eastward, southward, and westward. Corroborating data are generally unavailable at locations corresponding to the northern model boundary, but data from drilling of a 6,193-ft deep test hole near Orlando (fig. 4) showed that the freshwater and brackish-water zone extended to 2,375 ft. Data obtained from the Alligator Alley test well (Meyer, 1989a) were interpreted to suggest that the freshwater and brackish-water zone might be as deep as 2,250 ft below land surface. In grid cells near the peninsular coasts, the depth of the zone was represented as varying from 1,200 to 1,650 ft. These depths are representative of ones actually measured during the drilling of deep injection wells along the lower east coast of the Florida Peninsula (Reese, 1995).

When the vertical hydraulic conductivity value assigned to the middle confining unit layers (9-15) was moderate to large (1 ft/d or greater), the four cited measured or estimated heads in the Upper Floridan aquifer were matched within 1 ft by this procedure (fig. 8). However, when the vertical hydraulic conductivity value was reduced to 0.1 ft/d, the heads computed for the Upper Floridan aquifer were too low by 6 to 8 ft at these four locations. Another requirement for matching Upper Floridan aquifer heads was that the hydraulic conductivity value of the intermediate confining unit (assigned to model layers 4-7) be small (0.01 ft/d or less). This had the effect of hydraulically isolating the Upper Floridan aquifer from the surficial aquifer system. When the middle confining unit was assigned a vertical hydraulic conductivity of 1 ft/d and the intermediate confining unit was assigned a vertical hydraulic conductivity of 0.01 ft/d, the model simulated upward circulation from the Boulder Zone to the Tamiami Formation (layer 3) at a vertically uniform rate of 2.2×10^{-3} ft/d (pore velocity). It should be noted that the values used for vertical hydraulic conductivity work on a regional scale and could represent the effect of faults, fractures, or major solution features rather than the properties of the rock matrix.

The conceptual model supported by this analysis (fig. 9) is that a highly permeable and hydraulically nearly uniform Boulder Zone is separated from the Upper Floridan aquifer by a leaky middle confining unit that does not hydraulically isolate the two aquifers. Within the lower part of the leaky confining unit where native formation water is of seawater density, the head is uniform and nearly equal to that of the Boulder Zone. However, the intermediate confining unit does hydraulically isolate the Upper Floridan aquifer from the surficial aquifer system. Because the thickness of the freshwater zone that overlies water of seawater density and the corresponding depth to seawater varies across the Florida Peninsula, the height of the freshwater column that is supported by the nearly uniform Boulder Zone head also varies. (Because pressure in a body of seawater with uniform head increases with depth more rapidly than it does in a body of freshwater, the equivalent freshwater head also increases with depth in a body of saltwater of uniform head.) This is manifest in the mounding of heads in the central and northern part of the southern Florida Peninsula illustrated in figure 4 and schematically illustrated in figure 9. The confinement provided by the intermediate confining unit prevents these head gradients from being dissipated by upward circulation into the relatively permeable surficial aquifer system.

This conceptual model is supported by the findings of other investigators who have measured pressure at various depths in wells within the study area. Meyer (1989a) shows heads to be approximately in hydrostatic equilibrium (after correcting for varying density) in the Floridan aquifer system. The simulation of the conceptual model presented herein is highly generalized and does not account for variations in the depth and thickness of units and hydraulic conductivity, nor for the occurrence of local manmade stresses on the Upper Floridan aquifer and other aquifers. However, because the four measured or estimated heads are nearly replicated by the steady-state simulation, this simulation was used as an initial state for simulations of artesian flow at Grossman Hammock using the wellbore and well-riser models.

Figure 8. Simulated heads in the Upper Floridan aquifer assuming upward leakage from the Boulder Zone.
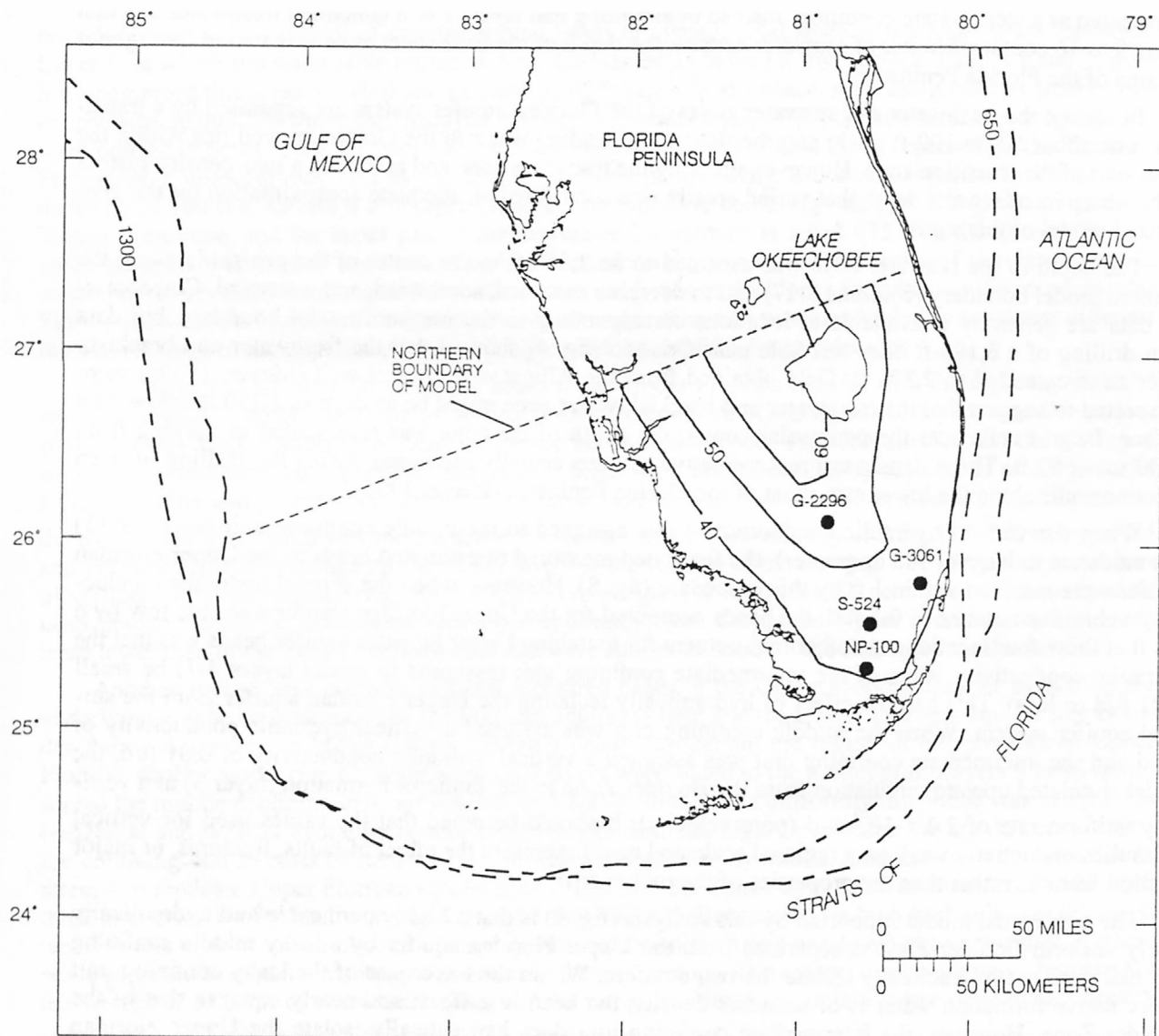
EXPLANATION

‐ ‐ ‐ ‐ ‐ SUBMARINE TOPOGRAPHIC CONTOUR - Shows depth below sea level in feet (Ucnupi, 1966)

———————— SIMULATED HEAD CONTOUR - Contour interval is 10 feet. Datum is sea level

● WELL AND IDENTIFIER
S-524

MODEL LAYERS	HYDROLOGIC UNIT	HYDRAULIC CONDUCTIVITY (FEET PER DAY)
1-2	BISCAYNE AQUIFER	30,000 OR 1,500
3	SEMICONFINING UNIT	100
4-7	INTERMEDIATE CONFINING UNIT	LESS THAN OR EQUAL TO 0.01
8	UPPER FLORIDAN AQUIFER	700
9-15	MIDDLE CONFINING UNIT	GREATER THAN OR EQUAL TO 1
16	LOWER FLORIDAN AQUIFER UNDERLYING CONFINING LAYER	1,000,000 NOT REPRESENTED IN MODEL

POTENTIOMETRIC SURFACE OF UPPER FLORIDAN AQUIFER

DEPTH, IN FEET BELOW SEA LEVEL

100

300

1,175
1,200

3,000
3,050

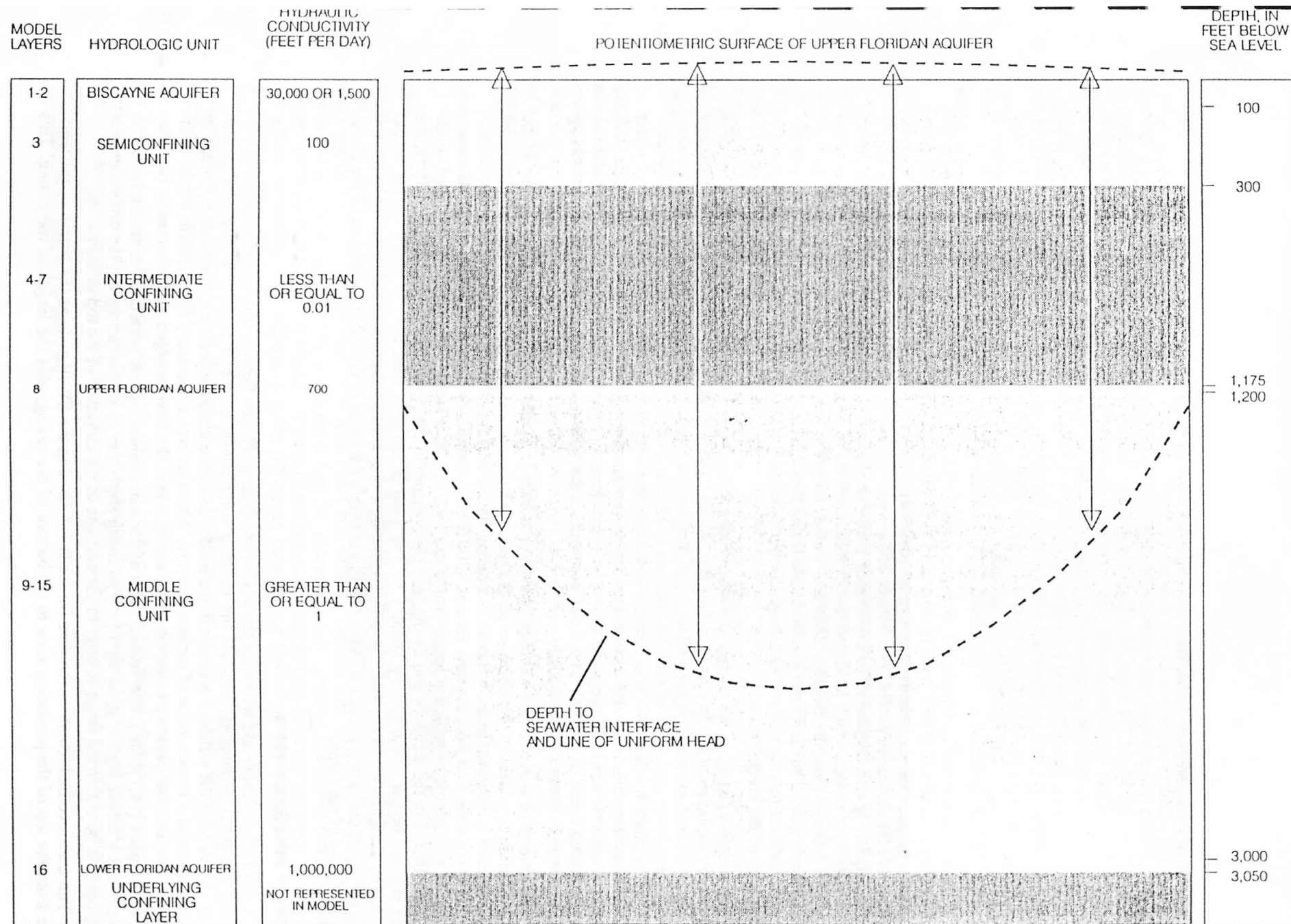DEPTH TO SEAWATER INTERFACE AND LINE OF UNIFORM HEAD

Figure 9. A conceptual model of hydraulic relations in the aquifer system extending from the water table to the base of the Lower Floridan aquifer (Boulder Zone) that is consistent with head measured in the Upper Floridan aquifer.

Simulation of Artesian Flow Rates with the Regional Flow Model

Adding the type 5 artesian well option to the simulation code and constructing the generalized Upper Floridan aquifer model set the stage for simulating artesian flow rates as a function of time and various assumed values of aquifer transmissivity. The sensitivity of the model to specified parameter values that might be incorrectly estimated, such as roughness coefficient and the diameter of the borehole, will be tested. Such analyses can indicate a possible range of error for the computed flow rates. The subsequent sections are concerned with the transmissivity estimate, the determination of the most probable flow rate, and sensitivity testing. The difference between 1944 and 1964 flow-rate measurements is examined on the basis of: (1) increasing aquifer drawdown through time, (2) the change in the height of the casing above land surface, (3) the insertion of the 8-in. liner, or (4) the deteriorating condition of the well.

Filling in of the well is rejected as an explanation of the diminution of flow with time because the cited well depth of about 1,250 ft was the same in 1983 as earlier reported by Mr. McCord (an oil company owner) and because the continuous discharge of water originating from near the bottom of the well would tend to retard the settling of debris in the bottom of the well. A change in the background head in the Upper Floridan aquifer is also rejected as an explanation partly on the basis of results of the simulation of heads with the 16-layer model, which indicated that hydraulic effects of manmade stresses would be dampened by upward leakage from the Boulder Zone and would, therefore, be highly localized. In addition, head measurements since 1961 have not indicated a declining trend. Other possible explanations, measurement error and underestimating well losses through corroded casing, are not amenable to study with the analytical methods of the investigation.

The possible change in the salinity of water flowing from the well was considered as another possible contributory explanation for the change in flow rate. As previously noted, the results of the 1944 laboratory analysis of a water sample from the well indicated that the measured chloride and dissolved-solids concentrations were lower than those of samples obtained in 1963 and afterward. However, the chloride concentration determined by the 1944 field titration was consistent with results of post-1944 analyses of water samples from the well, as was the 1944 laboratory measurement of specific conductance. Therefore, it is considered unlikely that a change in the quality of water flowing from the well did occur. The simulations to be described in the following sections will assume that the density of water in the wellbore and casing did not change between 1944 and 1963.

If a change in water quality did occur, however, the effect would be relatively minor because of the small differences in density that characterize the water samples. It was previously noted that if freshwater were somehow substituted in the wellbore and casing for the brackish water (3,000 mg/L of dissolved solids) that entered the well from the flow zone at 1,200 ft below land surface after 1963, the height of a water column in a standpipe of sufficient length would rise by 2 to 3 ft. If water of 1,800 mg/L dissolved solids were substituted in the wellbore and casing, and the pressure in the flow zone were same as in 1963, the water column height would rise by about 1 ft. The head in the well near the top of the casing would increase by a slightly lesser amount. The effect on the flow rate would be approximately equivalent to lowering the top of the casing about 1 ft. As noted, this is within the range of error of the estimate of the elevation of the top of the casing.

Parameter Value Specifications

Use of the wellbore and well-riser models for computing artesian flows required specification of values for additional parameters. Because the iterative computation of the artesian flow rate was accomplished by varying the wellbore pressure in the production zone until the pressure computed at the top of the well matched a known value of surface pressure, this value was a necessary specification for the computation. The absolute pressure at the top of the casing where the flow emerged was assumed to be atmospheric pressure (14.7 lb/in^2). The vertical point of the water column where pressure was atmospheric was actually a few inches higher, at the top of the jet rising above the lip of the casing, but the difference was considered to be within the range of error of the estimate of the elevation of the top of the casing.

The user-specified length of the pipe (or well) is treated in the SWIP code as the length of the wellbore from the top of the producing zone to the bottom of the casing, plus the length of the casing. This

length changed by a small amount when the cairn of rocks around the well was constructed and the casing was extended vertically. Without loss of realism, however, the well was considered to extend 1,180 ft from the top of the producing zone to land surface in all computations. The additional friction loss in the few feet of additional casing after well modification was assumed to be negligible. This procedure permitted an alternative representation of the pressure at the top of the well as the pressure within the actual casing at the elevation of land surface, equal to the weight of the water column from land surface to the top of the casing plus atmospheric pressure. When the casing extended about 2 ft above average local land surface (10 ft above sea level), as it did when the well was first constructed, the absolute pressure at the top of the nominal 1,180-ft well was specified to be 15.56 lb/in^2. When the casing extended about 9 ft above average local land surface (17 ft above sea level), as it did after well modifications in the 1950's, the top-hole pressure was specified to be 18.59 lb/in^2.

As treated by the well-riser model, the relation between the wellbore pressure at the production zone and the pressure at the top of the well depended on parameters describing the inner diameter and roughness of the wellbore and casing. The diameter of the well casing or borehole was specified in the SWIP code as a variable quantity so that actual variations along successive sections of the borehole and casing could be represented. Based on the caliper log (fig. 3), the borehole was assumed to be 15 in. in diameter below 1,140 ft, 13 in. in diameter from 1,140 to 950 ft, and 14 in. in diameter from 950 ft to the bottom of the 12-in. casing. In simulations of conditions after well modifications in the 1950's, the diameter was reduced to 8 in. above 80 ft below land surface to account for insertion of the polyvinyl chloride liner. The two casings might have had slightly smaller inner diameters, but the cited values are used as an adequate approximation for computational purposes. The specified roughness coefficient was the SWIP code default value (0.0000125 ft), which is said to represent smooth pipe (INTERA Environmental Consultants, Inc., 1979). The well diameter and roughness coefficient were later varied in sensitivity analyses. The effect of the insertion of the 8-in. liner was specifically considered in one sensitivity analysis.

The damping factor (D_a) for the iterative computational process (eq. 7) also needed to be specified. In practice, this factor needed to be selected by trial and error for each set of computational parameters and needed to be reduced in stages as the computations progressed and the computed flow rate decreased. Generally, larger damping factors were required for larger friction losses. Specification of a damping factor that was too low led to solution divergence and the premature end of the computational sequence. Specification of a damping factor that was too high led to premature convergence to a surface pressure that did not match the prespecified top-hole pressure.

Simulation of 1944 and 1964 Flow Measurements

In the first set of simulations, the model of 1964 conditions, the condition of the well after the insertion of the 8-in. liner extending to 17 ft above sea level in the 1950's was represented. For simplicity, this condition was assumed to have prevailed from the drilling of the well in 1944. Hydraulic conductivity values of 300, 500, and 800 ft/d were specified for the Upper Floridan aquifer in successive simulations, and corresponding changes were made to the well index and damping factors. The results (fig. 10) show that the decrease of flow rate with time is less rapid than that computed by the constant drawdown formula. The change in the flow rate is negligible after 1 week (therefore, assuming the liner to be present since 1944 did not affect the result). This indicates that the difference between the 1944 flow rate measurement, made about 2 months after flow began, and the later measurements cannot be explained as a consequence of increasing aquifer drawdown over time. The smaller decrease with time is mainly the result of induced recharge from the Boulder Zone through the leaky confining layer, which, with the drawdown, reaches equilibrium after about 1 week.

The simulation of heads in the generalized Upper Floridan aquifer was repeated after each change of the hydraulic conductivity with negligible effect on heads computed at nodes corresponding to the four observation locations (fig. 6). This result is to be expected because heads at the base of the brackish water are in hydrostatic equilibrium with the hydraulically uniform Boulder Zone. The principal control on the spatial variation of heads in the Upper Floridan aquifer was the depth of the freshwater layer, which remained unchanged in the analyses.
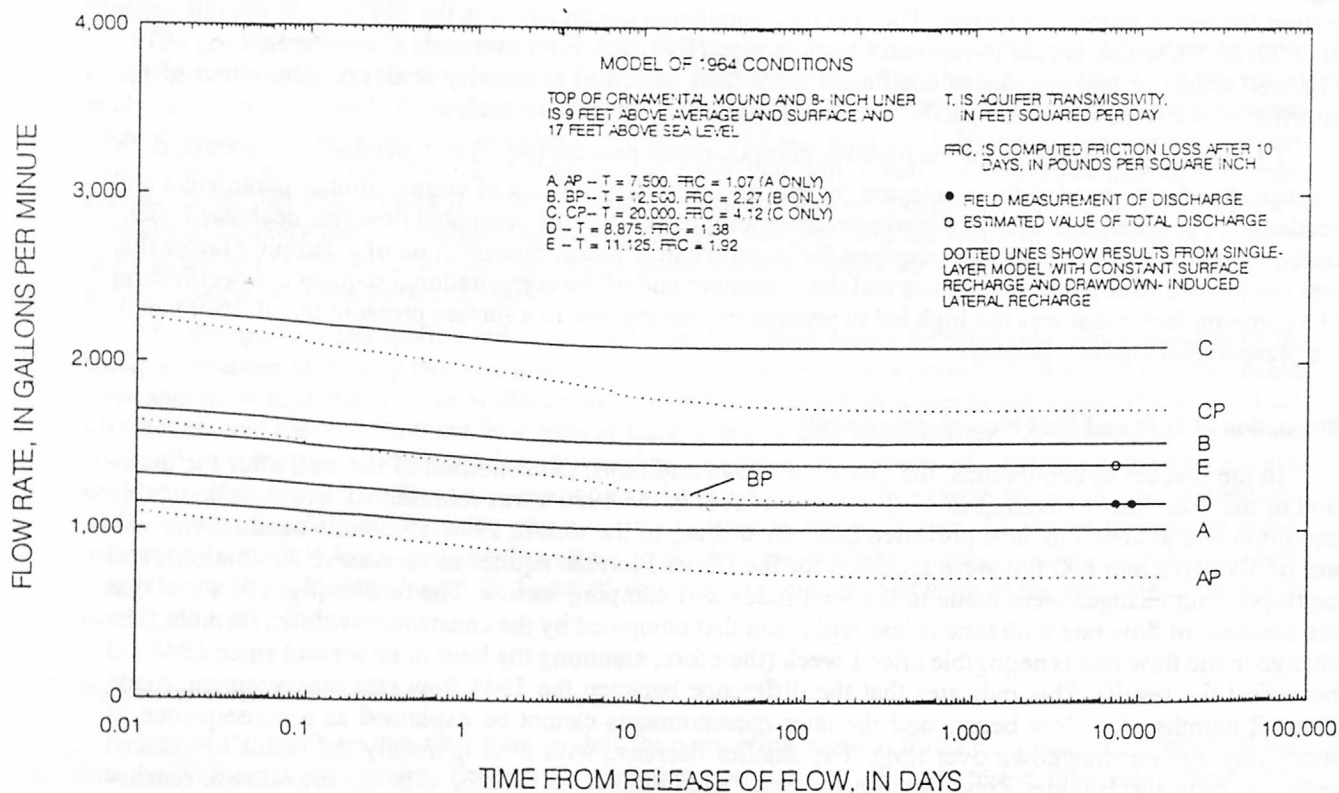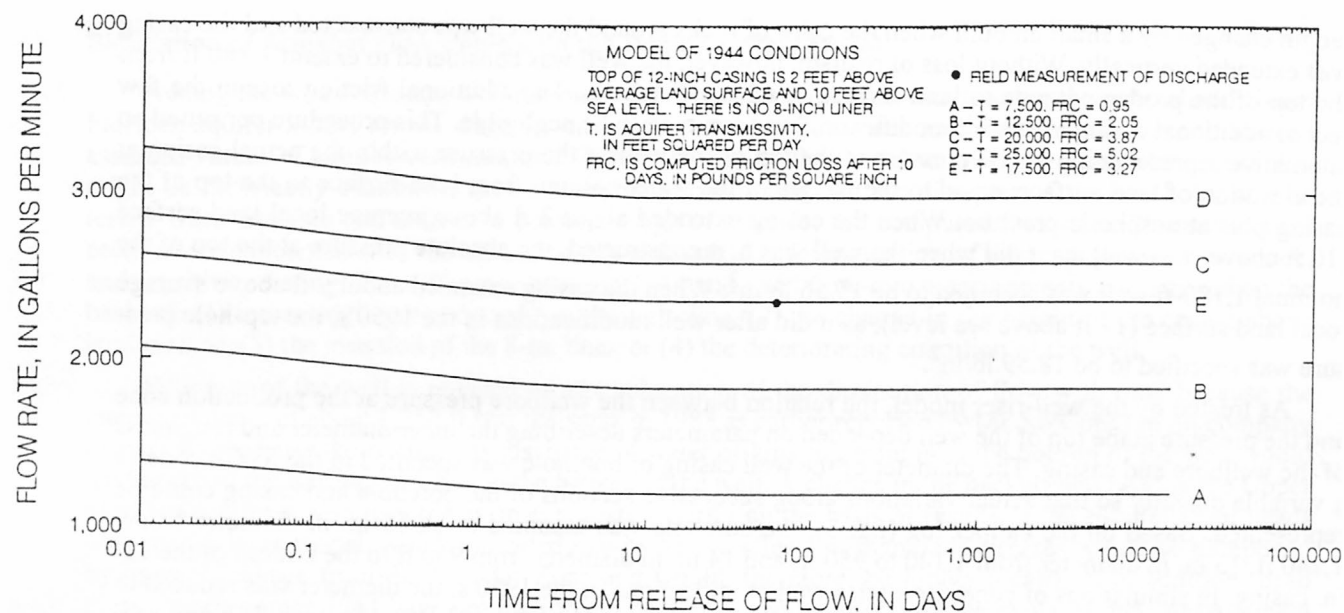
Figure 10. Simulation of 1944 and 1964 flow rates from the Grossman well in central Dade County.

To some degree, recharge from horizontal specified pressure boundaries also contributes to the smaller flow rate with time that is simulated with the 16-layer model. To evaluate this relation, flow from the Grossman well was simulated in the previously described simulation attempt in which a constant rate of surface recharge was specified in a single layer model with constant pressures specified at the horizontal recharge boundaries (fig. 7, case 4). (No Boulder Zone recharge was represented in case 4.) Model runs were made with hydraulic conductivity estimates of 300, 500, and 800 ft/d (fig. 10, dotted lines AP, BP, and CP). The decrease of the flow rate with time is substantially greater than when recharge from the Boulder Zone is simulated with the 16-layer model. This indicates that when recharge is entirely from horizontal boundaries in the surface recharge case, substantially greater drawdowns in the aquifer and a substantially longer time period are required for recharge to be in equilibrium with the depletion of aquifer storage than in the case where Boulder Zone recharge occurs. The flow rate does not approximately stabilize until after 1 year in the one-layer surface recharge case. On the other hand, the simulated decrease of the flow rate with time in the one-layer surface recharge case is appreciably less than that computed with the constant drawdown formula (fig. 5), which assumed that there are no drawdown-dependent recharge sources other than aquifer storage.

In the 16-layer model incorporating Boulder Zone recharge, the reduction of friction loss in the well as the flow rate decreases during the first week also contributes slightly to the reduction in the rate of the flow-rate decrease. When the hydraulic conductivity was 500 ft/d, the computed friction loss was 2.79 lb/in^2 after 0.01 day but decreased to 2.27 lb/in^2 after 1 week. The corresponding flow rate decrease was from 1,709 to 1,529 gal/min.

The measured 1964 flow rate of 1,170 gal/min and the estimated total 1964 flow rate of 1,400 gal/min were replicated by using hydraulic conductivity values of 355 and 445 ft/d, respectively, equivalent to transmissivities of 8,875 and 11,125 ft^2/d for the 25-ft thick aquifer (fig. 10, curves D and E). The higher value is accepted as a best estimate of the transmissivity of the Upper Floridan aquifer at this geographical location.

Another analysis was performed to determine if the flow rate measured about 2 months after the well was drilled in 1944 could be replicated with hydraulic conductivity values similar to those used to replicate the 1964 flow rate. The model of 1944 conditions was designed by specifying a land-surface pressure value of 15.56 lb/in^2, and the specification of an 8-in. casing diameter from land surface to 80 ft below land surface was deleted. Hydraulic conductivity values of 300, 500, and 800 ft/d and corresponding values of the well index were used. The simulated flow rates (fig. 10) were greater than in the 1964 case, as might be expected given that the opposing head at land surface was lower. As before, the simulated flow rates after 2 months and in 1964 were the same. When a hydraulic conductivity value of 500 ft/d (transmissivity of 12,500 ft^2/d) was used, the simulated flow rate was 1,835 gal/min. When the hydraulic conductivity was increased to 800 ft/d, the simulated flow rate was 2,580 gal/min, exceeding the rate measured in November 1944. When the hydraulic conductivity was 700 ft/d, the simulated flow rate was 2,355 gal/min, about the same as the measured value. Based on the 1944 flow measurement, the transmissivity of the flow zone would be 17,500 ft^2/d.

When the hydraulic conductivity value of 700 ft/d that was used to simulate the 1944 flow measurement was specified in the model of 1964 conditions, the simulated flow rate of 1,930 gal/min was 18 percent lower than the measured rate (2,355 gal/min), but still significantly higher than either the flow rate measured in 1964 (1,170 gal/min) or the estimated total flow rate (1,400 gal/min) in 1964. When hydraulic conductivity values of 355 and 445 ft/d, which were used to simulate the measured and estimated total 1964 rates, respectively, were specified in the model of 1944 conditions, the respective simulated flow rates (1,395 and 1,670 gal/min) were 19 percent higher than those simulated with the model of 1964 conditions. However, these rates are still significantly less than the measured 1944 rate. Thus, well modifications that had the effect of increasing the opposing surface pressure by raising the top of the casing and of increasing the friction loss by inserting an 8-in. liner in the well can only partly explain the difference between flow rates measured in 1944 and 1964.

The degree to which installation of an 80-ft section of 8 in. liner reduced the rate of flow from the well was assessed by deleting the liner specifications from the model of 1964 conditions. The specified surface

pressure was unchanged, which implied that the height of the well remained unchanged. Other parameter specifications were also unchanged. The assumed value of hydraulic conductivity was 445 ft/d, which previously had led to the simulation of a flow rate of 1,400 gal/min (fig. 10). Deleting the liner specification increased the simulated flow rate an unsubstantial 3.6 percent, to 1,450 gal/min. The friction loss in the well, however, decreased 30 percent from 1.92 to 1.34 lb/in^2. Another result of this analysis was to demonstrate that the difference in flow rates (1,670 and 1,400 gal/min) simulated by the models of 1944 and 1964 conditions, respectively, when a hydraulic conductivity of 445 ft/d was specified in each, was primarily the result of raising the top of the casing.

Analysis with the modified wellbore and well-riser algorithms of the SWIP code indicate that the discrepancy between the cited 1944 and 1964 flow measurements is substantially greater than indicated by the constant drawdown analysis. Three of the six suggested explanations for the difference have been the subject of analysis with the model, of which only one, raising the top of the well, can partly explain the difference in measured flow rates. This leaves the deteriorating condition of the well and the consequent increase of friction loss as the remaining possible explanation that can be posed for model analysis. The significance of the roughness of the interior of the casing and its relation to friction loss and flow rate will be examined in the following section.

Well Characteristics Affecting the Degree of Friction Loss

The discussion of the flow-rate analysis using the wellbore and well-riser representations has highlighted the significance of friction losses in the wellbore and casing in determining the flow rate. Therefore, an investigation to test the sensitivity of the model to parameters (well diameter and roughness coefficient) that represent characteristics of the well that determine the amount of friction loss is appropriate. The casing diameter was known, but the average borehole diameter had to be estimated for various depth intervals on the basis of caliper logs. The condition of the casing wall was never directly investigated, but corrosion could have increased its roughness and rust holes were believed to have led to a reduction of flow at land surface. The roughness of the borehole is not known, and the only relevant data are the caliper logs that show short intervals of high rugosity separated by longer intervals of relatively constant dimension. Generally, there is some uncertainty associated with the diameter and roughness specifications in the models of 1944 and 1964 conditions.

The simulation used as a control for sensitivity analyses specified a uniform 12-in. borehole and/or casing extending from the flow zone at 1,200 ft below land surface to 17 ft above land surface, and the hydraulic conductivity of the Upper Floridan aquifer was 445 ft/d. The roughness coefficient of the control was the SWIP code default value that represented smooth pipe (0.0000125 ft). The artesian flow rate after 1 week was computed to be 1,386 gal/min, and the corresponding friction loss was 1.71 lb/in^2.

In one set of sensitivity analyses, the well diameter was considered uniform between land surface and the flow zone and was varied from 1 to 48 in., with results shown in figure 11. When the diameter was varied in these computations, the well index (eq. 8) was also varied by a concomitant amount. All other parameters remained the same as in the control simulation. The smooth pipe roughness coefficient was used.

The 12- to 15-in. diameter of the Grossman well seems to be nearly optimum from an economic standpoint in allowing the artesian zone to discharge water. (The cost of well casing increases substantially with increasing diameter.) When the diameter of the well in the simulation was doubled to 24 in., the simulated flow rate increased only 18 percent to 1,630 gal/min, and the friction loss decreased from 1.71 to 0.08 lb/in^2. When the diameter was doubled again to 48 in., the flow rate increased only slightly to 1,755 gal/min, and the friction loss was negligible. Had the Grossman well been constructed with a casing having a diameter larger than 12 in., the cost would have greatly increased without substantially increasing the flow rate.

Specifying well diameters of less than 12 in. in the simulation substantially diminishes the simulated flow rate. When the diameter was halved to 6 in., the flow rate was more than halved, to 600 gal/min. The friction loss increased sixfold to 10.72 lb/in^2. The simulated flow rate was 120 gal/min for a specified diameter of 3 in., 43 gal/min for a 2-in. diameter, and 7 gal/min for a 1-in. diameter. In the latter case, the
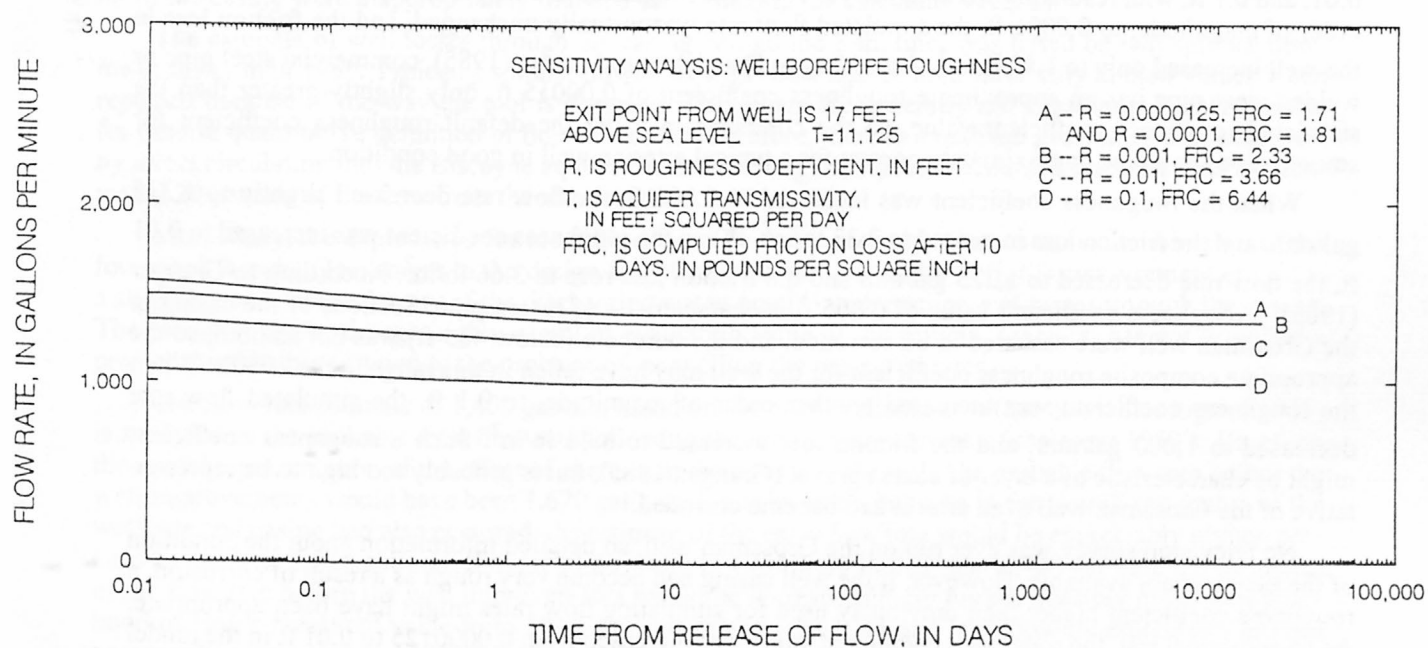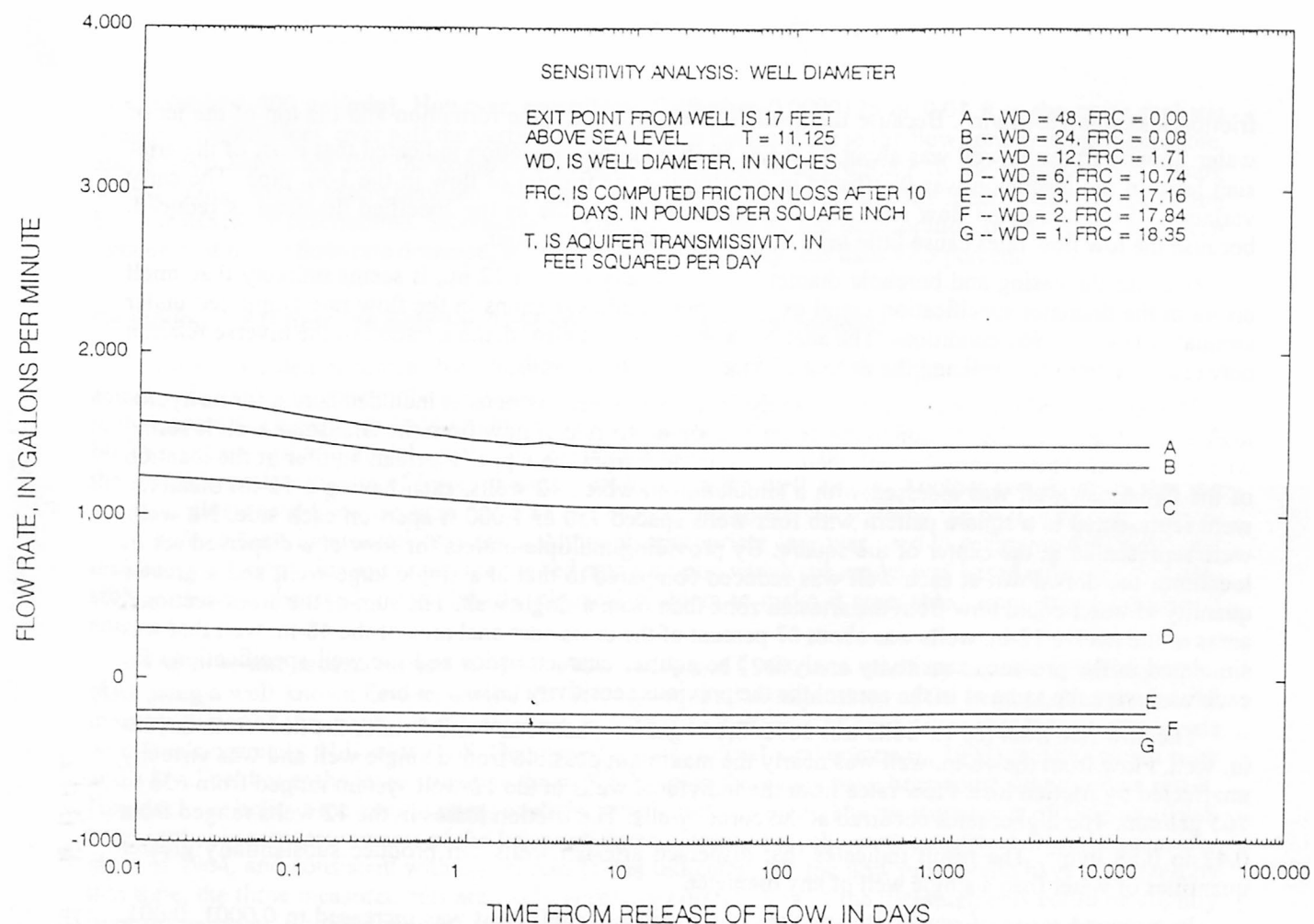
Figure 11. Sensitivity of simulated flow rates from the Grossman well in central Dade County to well diameter and wellbore/pipe roughness.

friction loss was 18.4 lb/in^2. Because the head difference between the formation and the top of the jet of water flowing from the well was about 47 ft (20.34 lb/in^2), the simulation indicated that most of the artesian head in the formation was dissipated in overcoming the friction of flow in the 1-in. pipe. The early variation of the simulated flow rate with time becomes negligible as the specified diameter is reduced, because the low flow rates cause little head drawdown within the aquifer.

Because the casing and borehole diameter are generally at least 12 in., it seems unlikely that small errors in the diameter specification could explain appreciable variations in the flow rate computed under simulated 1944 or 1964 conditions. The analysis also provides a clear demonstration of the inverse relation between flow from the well and the degree of friction loss in the well.

The result of an additional sensitivity analysis related to well diameter is included herein for its hydrologic interest, although its relation to the determination of the rate of flow from the Grossman well is somewhat tangential. The potential magnitude of artesian flow from the Upper Floridan aquifer at the location of the Grossman well was assessed with a simulation in which 12 wells, each having a 12-in. diameter, were represented in a square pattern with four wells spaced 750 or 1,000 ft apart on each side. No wells were represented in the center of the square. By providing multiple outlets for flow at a dispersed set of locations, the drawdown at each well was reduced compared to that at a single large well, and a greater quantity of water could flow from the artesian zone than from a single well. The sum of the cross-sectional areas of the twelve 12-in. wells was about 87 percent of the cross-sectional area of the 48-in. well that was simulated in the previous sensitivity analysis. The aquifer characteristics and the well specifications for each well were the same as in the control for the previous sensitivity tests.

The flow rate from the 12 wells was about 8,140 gal/min, between 4 and 5 times the flow from the 48-in. well. Flow from the 48-in. well was nearly the maximum possible from a single well and was virtually unaffected by friction loss. Flow rates from the individual wells in the 12-well system ranged from 635 to 765 gal/min. The higher rates occurred at the corner wells. The friction losses in the 12 wells ranged from 0.42 to 0.58 lb/in^2. The result indicates that dispersed artesian wells can produce substantially greater quantities of water than a single well of any diameter.

In a second series of sensitivity analyses, the roughness coefficient was increased to 0.0001, 0.001, 0.01, and 0.1 ft, with results shown in figure 11. When the roughness coefficient was increased by nearly an order of magnitude, to 0.0001 ft, the simulated flow rate was virtually unchanged, and the friction loss in the well increased only to 1.81 lb/in^2. According to Daugherty and others (1985), commercial steel pipe or welded steel pipe has an approximate roughness coefficient of 0.00015 ft, only slightly greater than the simulated roughness coefficient value, so the computation using the default roughness coefficient for smooth pipe should be approximately correct for a typical artesian well in good condition.

When the roughness coefficient was increased to 0.001 ft, the flow rate decreased slightly to 1,335 gal/min, and the friction loss increased to 2.33 lb/in^2. When the roughness coefficient was increased to 0.01 ft, the flow rate decreased to 1,225 gal/min and the friction loss rose to 3.66 lb/in^2. According to Granger (1985), a roughness coefficient value of 0.005 ft is characteristic of rusted pipe. Sections of the casing in the Grossman well were assumed to be corroded, and the borehole (below 485 ft) was not smooth, so the appropriate composite roughness coefficient for the well may have fallen in this range in the 1960's. When the roughness coefficient was increased another order of magnitude, to 0.1 ft, the simulated flow rate decreased to 1,000 gal/min, and the friction loss increased to 6.44 lb/in^2. Such a roughness coefficient might be characteristic of a layer of coarse gravel (Granger, 1985), but is probably too high to be representative of the Grossman well even after it had become corroded.

No television survey was ever run on the Grossman well, so detailed information about the condition of the casing is not available. However, if the well casing had become very rough as a result of corrosion, a roughness coefficient higher than previously used for simulating flow rates might have been appropriate. As an additional test, the roughness coefficient value was increased from 0.0000125 to 0.01 ft in the model of 1944 conditions. The flow rate was reduced by 17.5 percent, to 1,940 gal/min. When, in addition, the 8-in. liner was represented and land-surface casing pressure was increased to 18.59 lb/in^2 (1964 conditions), the flow rate was further reduced to 1,510 gal/min (only 8 percent greater than the estimated total 1964

flow rate of 1,400 gal/min). However, regardless of whether 0.0000125 or 0.01 ft is the more realistic roughness coefficient, over half the vertical distance from land surface to the flow zone was open borehole, and it seems unlikely that the composite roughness coefficient appropriate to the borehole and casing would change to this extent between 1944 and 1964 even if the condition of the inner surface of the casing had substantially deteriorated. Increasing friction loss caused by the deterioration of the well casing can explain part of the flow-rate decrease, but the amount is probably less than 17.5 percent.

Resolving the Difference Between 1944 and 1964 Flow Measurements

It has been demonstrated that changes in the design and condition of the well probably can partly account for the difference in the flow rates measured 20 years apart in 1944 and 1964. Another possibility is that water discharging in 1944 might have been slightly lighter. Other possible explanations not amenable to analysis with the model are the inaccuracy inherent in the method of measuring well discharge and the possible underestimate of losses through rust holes in the well casing. Another possibility is that the model analysis could be inaccurate because one or more parameters describing the well or aquifer characteristics have been inaccurately specified. Although appropriate care was used in designing the model and making parameter value estimates, the limited data set upon which the model was based enhances this possibility, as does the need to simplify the physical problem to make it amenable to numerical simulation analysis.

The flow-rate measurements of 1944, 1964, 1965, and 1969 were made by capable USGS hydrologists using a well-known field technique that merely required measurement of the height of the jet and the inner diameter of the casing. It is reasonable to have a high degree of confidence that the measurements were made correctly. N.D. Hoy (U.S. Geological Survey, retired, oral commun., 1993) recalls using a ruler and a level held over the jet to eliminate the possibility that there may have been some surging. As a result, Hoy does not believe the measurement could be substantially in error. The measurements of 1965 and 1969 were probably remeasurements of the height of the jet. Because results were consistent with the measurement of 1964, and consistent with model results that indicated that the flow rate should have stabilized by that time, the three measurements are likely generally correct. Any of the measurements could be slightly in error, however, if the height of the jet were not accurately measured, or if the value for the inner diameter of the casing were inappropriately rounded off in making the calculation (eq. 1).

The estimate of well losses through the casing behind the 8-in. liner was based on two spinner flowmeter logs run in 1969. Typically, spinner flowmeter logs show appreciable variability in their values when repeated because of the way the tool is designed. These logs are generally not considered highly reliable for precise quantitative definition of flow amounts. Therefore, there is a realistic possibility that water lost by direct circulation into the Biscayne aquifer before reaching the surface accounted for more than the estimated 20 percent of the flow from the artesian zone.

Most likely, the explanation for the difference in flow-rate measurements is some combination of the four cited factors: (1) changes in the design and physical condition of the well, (2) measurement error, (3) a slight increase in the density of the discharging water, and (4) underestimate of losses through the casing. The model analysis, though highly useful in evaluating some of the principal factors, has not provided a precise quantitative solution to the problem of reconciling the measured rates.

The flow-rate estimate of 1,400 gal/min based on the only available data (the flowmeter logs of 1969) is accepted as the probable total flow rate following the well improvements of the early 1950's. Based on the model representation of the effect of changes in the design of the well, the probable flow rate before the well improvements would have been 1,670 gal/min. If a substantial increase in the overall roughness of the wellbore and casing had also occurred, the estimate of the early flow rate would be appreciably higher, perhaps as much as 2,025 gal/min. If the total flow rate in 1964 were higher than 1,400 gal/min, the estimates of the flow rate before the well improvements would be correspondingly higher, possibly in the neighborhood of Hoy's measured value from 1944. For this reason, the 1944 and later flow measurements are not necessarily contradictory, and if the apparent difference were greater than it should be as a result of a slight measurement error, could be considered mutually consistent.

Effect of Artesian Flow on Regional Flow System

When the estimated total flow rate in 1964 (1,400 gal/min) was simulated in the 16-layer model by specifying an aquifer transmissivity value of 11,125 ft^2/d, significant simulated head decreases in the model layer representing the Upper Floridan aquifer were restricted to a region surrounding the discharging well that is too small to illustrate at the scale of figure 8. The region with drawdowns greater than 1 ft extended about 1 mi from the well. The cell-averaged computed drawdown in the 100 x 100 ft grid cell containing the flowing well was 24.4 ft.

In contrast with these results, the computed drawdowns extended over a much larger area when Upper Floridan aquifer heads were simulated in the single-layer surface-recharge model (fig. 7, case 4). The region with drawdown greater than 1 ft extended 17 to 18 mi from the flowing well. The contrast between the 16-layer and single-layer surface recharge simulations illustrates an aspect of hydraulic interaction within the Floridan aquifer system that is of considerable significance for water managers who might consider the use of the Upper Floridan aquifer as a source of supply water for reverse osmosis, for blending with fresher water, or as a receptacle for the temporary storage of freshwater. Assuming that the 16-layer simulation of the surficial aquifer system, intermediate confining unit, and the Floridan aquifer system is based on a realistic conceptual model, results indicate that leakage to or from the Boulder Zone through the middle confining unit of the Floridan aquifer system substantially mitigates the hydraulic effect of stresses applied to the Upper Floridan aquifer. The upward leakage has little effect on the Boulder Zone because of its high hydraulic conductivity and ability to receive recharge from undersea outcrops. Upper Floridan aquifer drawdowns or head buildups from manmade stresses that might otherwise be regionally significant in extent instead remain highly localized.

If upward leakage through the middle confining unit occurs, it would likely be at a low rate and be distributed over a large area. In the 16-layer model of 1964 conditions, the flowing well generated a simulated upward pore velocity of 0.5 ft/d in the uppermost layer of the middle confining unit, 130 ft directly under the well. At a point 350 ft below the well, an upward pore velocity of 0.1 ft/d was simulated. Assessing the possibility for upconing of saline water in the Upper Floridan aquifer of southern Florida is beyond the scope of this study, but results suggest that a significant problem might exist where boreholes were only a short distance above water of substantially higher salinity.

Significance of Numerical Simulation Approach

The more realistic flow-rate estimates made with the numerical simulation approach differ substantially from estimates made with the constant drawdown formula because the simulation approach accounts for friction losses in the well and casing and the effect of recharge across leaky confining layers or from aquifer boundaries located at a finite distance. The tradeoff is that simulation analysis requires the application of a sophisticated model code and the construction of a reasonably accurate model of the source aquifer. The constant drawdown analysis provided a transmissivity estimate (21,250 ft^2/d) that was 91 percent greater than that (11,125 ft^2/d) determined from the numerical simulation analysis. (The latter estimate is assumed to be better because the simulation provides a realistic depiction of the aquifer flow system.) The easily applied constant drawdown formula did provide an estimate of transmissivity that agreed with the simulation estimate within a 100 percent error range. Analytical formulas based on greatly simplified assumptions are useful for quick estimates of this type and are significant and useful tools of analysis even when more accurate answers can be derived through the use of more elaborate methods.

The explication of the numerical simulation approach for analysis of artesian flow-rate data using a three-dimensional flow simulator demonstrates a potentially useful tool for estimating the transmissivity of the Upper Floridan aquifer in southern Florida. The use of a variable-density simulator to support a conceptual model of the Floridan aquifer system that approximately replicates measured heads is a useful innovation that can provide guidance for future analyses of flow and transport in the Floridan aquifer system. In addition, the construction of a generalized model of flows in the Floridan aquifer system, intermediate confining unit, and surficial aquifer system has relevance for water managers assessing the past hydraulic effects of flowing artesian wells and agricultural withdrawals and the probable future effects of withdrawals for reverse osmosis supply, blending with freshwater, and subsurface storage of freshwater. The demonstration that the effect of manmade stresses on the aquifer is likely to be dampened by leakage from underlying aquifers is a result of considerable significance for local water managers.

SUMMARY

In 1944, the Grossman well was drilled 1,250 ft into the Upper Floridan aquifer in a remote rocky glades region of Dade County, Florida, apparently as part of an oil-prospecting venture, and was left flowing under artesian pressure. The 12-in. black iron casing extended 1 to 2 ft above average local land surface, and the discharge was measured to be 2,350 gal/min in November 1944.

The site was landscaped as a private park in the early 1950's. Emplacement of spoil material around the well and the construction of an ornamental cairn of rocks raised the exit level of the water to about 8 to 10 ft above land surface. The discharge filled a small manmade lake that overflowed into a borrow pit excavated into the Biscayne aquifer. Because a lessening of the flow rate seemed to indicate that water was escaping into the aquifer through corrosion holes in the casing, 80 ft of 8-in. plastic liner was inserted into the upper part of the well. Additional flow measurements in 1964, 1965, and 1969 indicated 1,170 gal/min to be the flow rate of the well. The chloride and dissolved-solids concentrations averaged about 1,200 and 3,000 mg/L in water samples collected by the USGS between 1963 and 1978.

The site was acquired by the State of Florida in 1970 for development as a public recreational area. During 1978-79, a ground-water quality reconnaissance showed a large region of the Biscayne aquifer to have been invaded by the brackish water from the artesian well. The well was plugged in 1985. In the same year, the USGS began a study, in cooperation with DERM, to examine the contamination of the Biscayne aquifer using numerical simulation methods. The site has recently become part of Everglades National Park.

The application of numerical methods to simulate the plume of brackish water required that an apparent discrepancy between measured 1944 and later flow rates be resolved and that the flow-rate history be determined. Factors considered as possible explanations for the difference in flow rates were: (1) a diminution of flow as heads in the Upper Floridan aquifer were drawn down, (2) the raising of the altitude at which the water was discharged, (3) the installation of the narrow 80-ft liner, (4) increasing friction losses caused by corrosion of the casing, (5) error in one or more measurements of flow, (6) a change in the density of the discharging water, and (7) an underestimate of losses through the corroded casing.

The constant drawdown formula was applied to estimate transmissivity based on the flow measurements made in 1944, 2 months after suspension of well drilling, when the drawdown was 37 ft, and in 1964, when the drawdown was 30 ft. The measured 1944 flow rate of 2,350 gal/min was replicated when a transmissivity of 24,500 ft^2/d was specified. The measured 1964 flow rate of 1,170 gal/min and the estimated total flow rate of 1,400 gal/min (based on the estimated 20 percent reduction of flow at the bottom of the liner) were replicated with specified transmissivities of 17,700 and 21,250 ft^2/d, respectively. For all transmissivity specifications, the major part of the computed decrease in flow occurred in the first month. However, the computed flow rate continued to decrease slightly even after 40 years.

The constant drawdown analysis indicated that the two flow-rate measurements could be consistent if one or the other were in error by as little as 15 percent. However, the constant drawdown analysis did not account for friction losses in the well that varied in magnitude with the flow rate or for recharge through confining layers and from aquifer boundaries located at finite distances.

A three-dimensional simulator of flow and solute transport in ground water was used for a numerical simulation of the rate of flow from the well. The simulator accounted for the friction loss occurring in the wellbore and casing and the interaction of this friction loss with other factors. Because none of four previously coded methods of simulating the wellbore/formation hydraulic relations were found to provide an adequate solution for this application problem, an additional procedure was encoded. The "type 5" well option is an iterative procedure in which the pressure in the borehole adjacent to the receiving zone was varied until the pressure at the top of the well, after accounting for friction losses, matched a prespecified value. Correction factors were a function of the change of friction loss computed in previous iterations, the top-hole pressure error, and a user-specified damping factor that had to be determined on a trial-and-error basis for each new set of parameters. The wellbore simulator was revised to permit the representation of changes in the wellbore/casing diameter at specified depths.

In order to simulate predevelopment steady-state heads in the Upper Floridan aquifer, a 16-layer Cartesian grid with a 19 by 18 horizontal discretization was used to represent the sequence of major aquifers and confining units from land surface to the Boulder Zone (part of the Lower Floridan aquifer) at 3,000 ft. This highly generalized model replicated measured Upper Floridan aquifer heads when layers representing the middle confining unit separating the Upper Floridan aquifer from the Boulder Zone were assigned vertical hydraulic conductivities equal to or greater than 1 ft/d. This represented the assumption that the middle confining unit was leaky and did not maintain a hydraulic separation between the two aquifers. The model analysis did not determine whether the leaky nature of the unit represented the effect of widely distributed faults, fractures, or major solution or more uniform properties of the carbonate rock matrix.

Because of its high hydraulic conductivity, head was virtually uniform and constant in the model layer representing the Boulder Zone. The variable head in the Upper Floridan aquifer on the opposite side of the leaky middle confining unit was a consequence of the varying thickness of the zone of fresh and brackish water extending from land surface into the middle confining unit. Calibration of the model also required that low values of vertical hydraulic conductivity (less than 0.01 ft/d) be assigned to the intermediate confining unit separating the surficial aquifer system from the Upper Floridan aquifer. These two aquifers were, therefore, hydraulically isolated.

When the numerical simulation approach was used in an attempt to resolve the apparent discrepancy between the 1944 and 1964 measured flow rates, the simulated decrease of flow rate with time was found to be substantially less rapid than that computed by the constant discharge formula, becoming negligible after 1 week. This result reflects the influence of induced recharge from the Boulder Zone through the leaky middle confining unit. The result also indicates that time-related drawdown increases do not provide a resolution of the discrepancy in measured flow rates.

Computations of flow rate using various Upper Floridan aquifer transmissivity specifications were made under 1944 conditions (casing 2 ft above land surface, without 8-in. liner) and 1964 conditions (casing 9 ft above land surface, 80 ft of 8-in. liner in the well). Under the latter conditions, the measured 1964 flow rate of 1,170 gal/min and the estimated total flow rate of 1,400 gal/min were replicated with transmissivity estimates of 8,875 and 11,125 ft^2/d, respectively. The latter value is accepted as the best estimate of aquifer transmissivity at this geographical location.

Under 1944 conditions, the measured flow rate of 2,350 gal/min was replicated with a transmissivity estimate of 17,500 ft^2/d. When this transmissivity estimate was used with 1964 conditions, and when the transmissivity estimates of 8,875 and 11,125 ft^2/d were used with 1944 conditions, the simulated flow rates only changed about 18 to 19 percent, not sufficiently to resolve the difference in measured flow rates. Installation of the 8-in. liner was estimated to have decreased the flow rate by only 3.6 percent. These results indicated that consideration of the well modifications alone did not resolve the discrepancy between measured 1944 and 1964 flow rates.

Parameters characterizing properties of the wellbore and casing that determined the amount of friction loss (the roughness of the borehole and interior of the casing and the diameter of the wellbore and casing) were next varied to determine their effect on the simulated flow rate. Increasing the roughness coefficient to 0.01 ft in the simulation representing 1944 conditions lowered the flow rate by 17.5 percent. A roughness coefficient of 0.01 ft could have represented a highly corroded casing and rugose borehole. The roughness coefficient of 0.0000125 ft, used in the control for sensitivity testing, assumed that the casing and borehole had the roughness of smooth pipe.

When the roughness coefficient of 0.01 ft and the well modifications of the 1950's were both represented in the simulation of the 1944 flow rate, the computed rate dropped to 1,510 gal/min, close to the estimated total rate of 1,400 gal/min in 1964. However, because the 1,180 ft separating the Upper Floridan aquifer and land surface contained only 485 ft of casing, it was unlikely that the composite roughness coefficient for casing and borehole could have changed to this extent as a result of deterioration of the casing. Other analyses indicated that slight inaccuracies in estimating the average borehole diameter did not significantly affect the analysis. Therefore, simultaneous simulation of all of the possible causes of the flow-rate discrepancy amenable to model analysis failed to fully resolve the apparent discrepancy.

Because the measurements in the 1960's and the 1944 measurement were not believed to have been substantially in error, the most likely resolution of the discrepancy in measured rates was through a combination of three factors: (1) changes in the design and physical condition of the well, (2) a slight degree of measurement error, and (3) underestimate of losses through the casing. A possible fourth factor could be a slight increase in the salinity (density) of the flowing water, though existing water-quality data are insufficient to substantiate this possibility. The model analysis, though highly useful in identifying significant factors, did not provide a precise quantitative resolution of the rate discrepancy. The flow estimate of 1,400 gal/min is accepted as the most probable flow rate following the well improvements.

If the multiple-layer representation of the aquifer systems and confining units from land surface to the Boulder Zone is conceptually realistic, it indicates that leakage through the middle confining unit of the Floridan aquifer system substantially mitigates the hydraulic effect of stresses applied to the Upper Floridan aquifer. Drawdowns or head buildups from brackish water withdrawal or subsurface storage of freshwater would tend to remain highly localized.

The more realistic flow-rate estimates made with the simulation modeling approach are substantially different from estimates made with the constant drawdown formula because the simulation approach accounts for friction losses in the well and casing and the effect of recharge through leaky confining layers or from aquifer boundaries located at a finite distance from the well. The tradeoff is that simulation analysis requires the application of a sophisticated model code and the construction of a reasonably accurate model of the source aquifer. The explication of the numerical simulation approach demonstrates a potentially useful tool for estimating the transmissivity of the Upper Floridan aquifer in southern Florida from measured artesian well flow rates.

REFERENCES CITED

Adler, Lynne, 1973, A comparison of the hydrophytes growing in artesian and nonartesian water at Grossman Hammock State Park, Dade County, Florida: Master's Thesis, University of Miami, Department of Biology, June 1973.

Applin, P.L., and Applin, E.R., 1967, The Gulf Series in the subsurface in northern Florida and southern Georgia: U.S. Geological Survey Professional Paper 524-G, 34 p.

Beaven, T.R., and Meyer, F.W., 1978, Record of wells in the Floridan aquifer in Dade and Monroe Counties, Florida: U.S. Geological Survey Open-File Report 78-881, 30 p.

Causaras, C.R., 1987, Geology of the surficial aquifer system, Dade County, Florida: U.S. Geological Survey Water-Resources Investigations Report 86-4126, 240 p., 3 sheets.

Chen, C.S., 1965, The regional lithostratigraphic analysis of Paleocene and Eocene rocks of Florida: Florida Geological Survey Bulletin 45, 105 p.

Dames & Moore, Inc., 1972, Research test well, Floridan aquifer, Turkey Point, Florida: Engineering report prepared for the Florida Power and Light Company, November 1972.

----- 1975, Floridan aquifer water-supply investigations, Turkey Point area, Dade County, Florida: Engineering report prepared for the Florida Power and Light Company, August 1975.

Daugherty, R.L., Franzini, J.B., and Finnemore, E.J., 1985, Fluid mechanics with engineering applications: McGraw Hill, 598 p.

Granger, R.A., 1985, Fluid mechanics: Holt, Rinehart, and Winston, 884 p.

Hoffmeister, J.E., Stockman, K.W., and Multer, H.G., 1967, Miami Limestone of Florida and its recent Bahamian counterpart: Geological Society of American Bulletin, v. 78, p. 175-190.

Hydrologic Associates U.S.A., Inc., 1991, Aquifer performance test: Engineering report submitted to the Miami-Dade Water and Sewer Authority Department, October 1991.

INTERA Environmental Consultants, Inc., 1979, Revision of the documentation for a model for calculating effects of liquid waste disposal in deep saline aquifers: U.S. Geological Survey Water-Resources Investigations Report 79-96, 73 p.

INTERCOMP Resource Development and Engineering, Inc., 1976, A model for calculating effects of liquid waste disposal in deep saline aquifers; Part I--Development, Part II--Documentation: U.S. Geological Survey Water-Resources Investigations Report 76-61, 253 p.

Jacob, C.E., and Lohman, S.W., 1952, Nonsteady flow to a well of constant drawdown in an extensive aquifer: American Geophysical Union Transactions, v. 33, p. 559-569.

Kipp, K.L., Jr., 1987, HST3D: A computer code for simulation of heat of the solute transport in three-dimensional ground-water flow systems: U.S. Geological Survey Water-Resources Investigations Report 86-4095, 517 p.

Labowski, J.L., 1988, Water quality monitoring and results of surface geophysical surveys for a mineralized artesian plume, southwest Dade County, Florida, 1982-1987: Metro-Dade County Department of Environmental Resources Management Technical Report 88-7, 40 p.

Lawrence, F.E., and Braunworth, P.L., 1906, Fountain flow of water in vertical pipes: American Society of Civil Engineers, Transactions, v. 57, p. 265-306.

Lohman, S.W., 1979, Ground-water hydraulics: U.S. Geological Survey Professional Paper 708, 70 p.

Maher, J.C., 1971, Geologic framework and petroleum potential of the Atlantic Coastal Plain and Continental Shelf: U.S. Geological Survey Professional Paper 659, 98 p.

Merritt, M.L., 1984, Digital simulation of the regional effects of subsurface injection of liquid waste near Pensacola, Florida: U.S. Geological Survey Water-Resources Investigations Report 84-4042, 73 p.

----- 1995a, Digital simulation of flows in the Biscayne aquifer of southern Dade County, Florida, from 1945 to 1989, incorporating transient surface inundation and a developing water-management system: U.S. Geological Survey Water-Supply Paper (in review).

----- 1995b, Digital simulation of a plume of brackish water in the Biscayne aquifer originating from a flowing artesian well, Dade County, Florida: U.S. Geological Survey Water-Supply Paper (in review).

----- 1995c, Tests of subsurface storage of freshwater at Hialeah, Dade County, Florida, and digital simulation of the salinity of recovered water: U.S. Geological Water-Supply Paper 2431 (in press).

Meyer, F.W., 1989a, Hydrogeology, ground-water movement, and subsurface storage in the Floridan aquifer system in southern Florida: U.S. Geological Survey Professional Paper 1403-G, 59 p.

----- 1989b, Subsurface storage of liquids in the Floridan aquifer system in southern Florida, 1943-83: U.S. Geological Survey Open-File Report 88-477, 25 p.

Miller, J.A., 1986, Hydrogeologic framework of the Floridan aquifer system in Florida and in parts of Georgia, Alabama, and South Carolina: U.S. Geological Survey Professional Paper 1043-B, 91 p.

Puri, H.S., and Winston, G.O., 1974, Geologic framework of the high transmissivity zones in south Florida: Florida Bureau of Geology Special Publication 20, 101 p.

Reese, R.S., 1995, Hydrogeology and the distribution and origin of salinity in the Floridan aquifer system, southeastern Florida: U.S. Geological Survey Water-Resources Investigations Report 94-4010.

Rouse, Hunter, ed., 1949, Engineering hydraulics: New York, John Wiley & Sons.

Smith, C.A., Lidz, Lauralee, and Meyer, F.W., 1982, Data on selected deep wells in south Florida: U.S. Geological Survey Open-File Report 82-348, 144 p.

Uchupi, Elezar, 1966, Map showing relation of land and submarine topography, De Soto Canyon to Great Bahama Bank: U.S. Geological Survey Miscellaneous Geologic Investigations Map I-475, 1 sheet.

U.S. Bureau of Reclamation, 1984, Water measurement manual: Revised reprint, U.S. Department of the Interior, Bureau of Reclamation.

Waller, B.G., 1982, Areal extent of a plume of mineralized water from a flowing artesian well in Dade County, Florida: U.S. Geological Survey Water-Resources Investigations Report 82-20, 20 p.

APPENDIX

LITHOLOGICAL AND PALEONTOLOGICAL DESCRIPTION OF CUTTINGS OBTAINED DURING DRILLING OF THE GROSSMAN WELL, S-524, IN 1944

[Determinations are by Dr. Louise Jordan, Sun Oil Company. Report provided to Dr. J.A. Waters, Sun Oil Company, Dallas, Texas, in a letter dated November 20, 1944]

LOG OF WELL S-524

[This log is included in this report by virtue of permission from the
Oryx Energy Company, Dallas, Texas]

Location: 25-55S-37E

State #1
Miami Shipbuilding Corporation
Dade County, Florida

Depth below land surface (feet)	Description
290 - 300	In the Miocene, fine and coarse polished sand, phosphate granules, fish remains.
300 - 310	Ninety-five percent, fine angular sand as from a clay and 5 percent polished quartz and phosphate pebbles, rare shell fragments.
310 - 320	As above, shell fragments, *Nonion grateloupi, Cibicides floridana, Planulina, Anomalina, Globigerina.*
320 - 330	Same. Shark's teeth; trace, tannish-brown, fine, sandy clay, rare forams.
330 - 340	As above, 10 percent clay, *Discorbis floridana, Cassidulina laevigata var. carinata, Bulimina gracilis* and other Miocene foraminifera.
340 - 350	As above.
350 - 400	Essentially the same.
400 - 410	As above.
410 - 420	As above.
420 - 430	As above.
430 - 440	As above.
440 - 450	As above, very abundant *Globigerina* and other small foraminifera of Miocene age.
450 - 460	As above, abundant *Uvigerina, U. auberiana, U. cf. pigmea*, etc.
460 - 470	As above; trace, brown, fine crystalline dolomite; abundant small Miocene foraminifera.
470 - 480	Top, white, fine, sandy, phosphatic limestone;, large shell fragments; bryozoa; echinoids, *Amphistegina lessoni, Asterigerina miocenica, Asterigerina* sp., *Elphidium* sp.
480 - 490	Same.
490 - 510	Same.
510 - 520	White, fossiliferous, granular limestone; echinoid; shells; bryozoa; and foraminifera as above, Amphistegina and Asterigerina common.
520 - 530	Same.
530 - 540	More sandy.
540 - 550	Same.
550 - 560	Same.
560 - 570	Hard, fossiliferous limestone; *Archais, Sorites*, pelecypods, *Amphistegina chipolensis, Gypsina.*
570 - 580	As above, 20 percent fine sand; limestone more granular.
580 - 590	Same, abundant fossils, poor preservation.
590 - 600	Same, but harder.
600 - 630	Same.
630 - 640	Same.
640 - 650	Same.
650 - 660	Same.
660 - 670	Same.
670 - 740	There is a slight variation in hardness in the samples above.
740 - 790	Limestone appears to be slightly powdery or more chalky than above, abundant *Amphistegina*.
790 - 800	Limestone again becomes granular as above 740 feet.

Depth below land surface (feet)	Description
800 - 810	Same, abundant *Amphistegina*, still in the Miocene at 860 feet.
860 - 990	Skip in samples.
990 - 1,000	In Oligocene, white, fossiliferous limestone; *Miogypsina hawkinsi* common; abundant shells; echinoid fragments, etc.
1,000	Pure-white, fossiliferous chalk; thin, oval *Miogypsina*; thick, discoid *Miogypsina* with large papillae.
1,000 - 1,060	Skip.
1,060 - 1,070	Limestone as above, large papillate Oligocene *Lepidocyclina*.
1,080	Same.
1,090	Light-cream, soft, slightly dolomitic chalk; ostracods rather common; *Valvulina floridana*.
1,100	Light-cream, soft, fossiliferous, coquina chalk; abundant *Valvulina floridana*; *Operculinoides dius*.

The next three samples do not carry *Valvulina floridana* but do carry Oligocene fauna not noted above. *Valvulina floridana* occurs in both the top of the Claiborne and in the Suwannee (basal Oligocene).

1,110	Light-cream to white granular chalk, *Heterostegina cf. texana, Camerina*.
1,120	As above, abundant *Miogypsina*.
1,130	Same.
1,140	Sandy, white, dolomitic limestone; ostracods common; *Amphistegina*.
1,150	Suggest top of Claiborne (*Lituonella* zone); sandy, white, dolomitic limestone as above; trace cream-colored dolomite-encrusted coquina; *Valvulina floridana* apparently from sandy phase. Small *Lituonella* from coquina, *Spirolina coryensis*.
1,160	Eighty percent sandy white dolomitic limestone; *Coskinolina* and *Dictyoconus cookei* from cream, dolomite coquina.
1,170 - 1,230	Fine, crystalline, white to cream-white, slightly calcareous dolomite; preservation of fauna different. *Dictyoconus cookei*, algal fragments, *Pseudochrysalidina floridana, Spirolina coryensis, Rotalia avonparkensis* and others.

SUMMARY

290 - 860	Of the Miocene: 290 - 470 feet. Sands and clays with small foraminifers. 470 - 480 feet. Top, white, fine, sandy, phosphatic limestone. 560 - 570 feet. Top, hard, fossiliferous limestone with Archais and Sorites.
890 - 990	Skip in samples.
990 - 1,080	Of the Oligocene, white, fossiliferous limestone with *Miogypsina hawkinsi* and other Oligocene large foraminifers.
1,090 - 1,100	Light-cream, soft, dolomitic chalk with *Valvulina floridana* and first *Operculinoides dius* noted, probably still in Oligocene.
1,110 - 1,130	Cream to white, granular chalk with Oligocene fauna. The 1,140-foot sample is a sandy dolomitic limestone with ostracods.
1,150	Suggest top Claiborne (*Lituonella* zone, middle Eocene).
1,230	In Claiborne. No Ocala fauna was noted. As the basal Oligocene and upper middle Eocene (Claiborne) carry several of the same species, it is difficult to separate them if no Ocala fauna is noted between them. The samples 1,110 to 1,130 feet seem to be out of place as the fauna in samples 1,090 and 1,100 feet was not observed in them. Although it is possible that the well from 1,150 to 1,230 feet is still in the Oligocene (due to this difficulty in separating basal Oligocene and uppermost Claiborne fauna), I believe the beds are Claiborne because of numerous species of foraminifera rather than abundant of two or three species, as is usually the case in the basal Oligocene.