Exchanges of Water between the Upper Floridan Aquifer and the Lower Suwannee and Lower Santa Fe Rivers, Florida

By J.W. Grubbs and C.A. Crandall

Prepared in cooperation with the
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Preface

Many areas of the United States have experienced water shortages as a consequence of increased water use due to population pressures, industrial growth, and changes in agricultural irrigation practices. As a result of these increasing demands on water resources, many states have established, or are considering, instream-flow protection programs to ensure that the water requirements for ecosystem maintenance will be met. The State of Florida in 1972 adopted legislation directing the water-management districts to establish minimum flows and levels (MFLs) for all watercourses, and minimum levels for aquifers and surface water, in their respective regions. Section 373.042 of the Florida Statutes specifies that a minimum flow for a watercourse is the flow at which further withdrawals would be substantially harmful to the water resources or ecology of the area. Similarly, the Statute defines the minimum level as the level of water in an aquifer, or level of surface water, at which further withdrawals would be substantially harmful to the water resources of the area. The Statute also allows the development of minimum flows and levels using the “best information available” and the recognition of seasonal variation in setting the flows and levels.

The Suwannee River Water Management District (SRWMD) in the north-central part of the State is one of five regional water-management districts in Florida. The District’s first priority is to set MFLs for the lower Suwannee River, from its confluence with the Santa Fe River to the Gulf of Mexico. The SRWMD began the process for setting MFLs in 1994 with a series of long-term cooperative studies with the U.S. Geological Survey (USGS) that included data collection, analysis, and interpretation. The USGS program culminated in the completion of three major studies, which were conducted to understand the effects that reduced flow in the river could have on the forested floodplain and the mixing of freshwater and saltwater in the estuary, as well as the effects that ground-water withdrawals could have on flows in the river. These studies are reported in Chapters A, B, and C of this Professional Paper 1656 series. Additionally, a summary of the program is presented in Chapter D, which includes a discussion of how the results from these three studies can be used collectively by the SRWMD.

Chapter A of the series describes the hydrology, vegetation, and soils of the forested floodplain of the Lower Suwannee River. The chapter further describes the relation of forest types and other floodplain characteristics to long-term river flow, and estimates potential impacts on the floodplain if river flows were reduced. Chapter B focuses on flow and the mixing of freshwater and saltwater in the lower river and estuary. Salinity and other hydrologic data collected during a period of unusually low flow were used to calibrate a three-dimensional hydrodynamic and transport model that simulates time-varying water levels, currents (lateral, longitudinal, and vertical), and salinity conditions. This chapter includes important discussions of modeled scenarios and hydrologic changes that could result from a reduction of flow in the river. Reductions in streamflow could come from changes in climatic conditions or from direct withdrawal, but may also come from ground-water pumpage adjacent to or many miles from the river. Chapter C presents a discussion of hydrologic conditions governing the interaction between ground water and surface water, an evaluation of the magnitude and timing of water exchanges between the Lower Suwannee River and the Upper Floridan aquifer using historical data, and the models that were used to simulate the exchanges. Also presented in this chapter is a discussion of how a hydrologic model could be used to evaluate hypothetical water-use scenarios, and the ground-water and surface-water exchanges that could result from these hypothetical conditions. Chapter D summarizes the cooperative program and highlights the importance of this multidisciplinary program to our understanding of the hydrology in the Lower Suwannee Basin—an understanding borne out of an extensive data-collection program and complex interpretive studies. Chapter D provides a “roadmap” for water managers to make better use of the integrated results of these studies.

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Conversion Factors, Abbreviations, and Datum

Multiply	By	To obtain
	Length	
inch (in.)	25.4	millimeter (mm)
foot (ft)	0.3048	meter (m)
mile (mi)	1.609	kilometer (km)
	Area	
square mile (mi^2)	2.590	square kilometer (km^2)
	Volume	
gallon (gal)	3.785	liter (L)
	Flow rate	
cubic foot per second (ft^3/s)	0.02832	cubic meter per second (m^3/s)
inch per day (in/d)	0.0254	meter per day (m/d)
inch per year (in/yr)	25.4	millimeter per year (mm/yr)
	Hydraulic gradient	
foot per mile (ft/mi)	0.1894	meter per kilometer (m/km)
	Transmissivity*	
foot squared per day (ft^2/d)	0.09290	meter squared per day (m^2/d)

**Transmissivity:* The standard unit for transmissivity is cubic foot per day per square foot times foot of aquifer thickness [(ft^3/d)/ft^2]ft. In this report, the mathematically reduced form, foot squared per day (ft^2/d), is used for convenience.

Temperature in degrees Fahrenheit (°F) may be converted to degrees Celsius (°C) as °C = (°F - 32)/1.8.

Vertical coordinate information is referenced to the National Geodetic Vertical Datum of 1929 (NGVD 1929), and *horizontal coordinate information* is referenced to the North American Datum of 1983 (NAD 83). NGVD 1929 is approximately equal to sea level.

Acronyms

ADCP	acoustic Doppler current profiler
PEST	parameter estimation program
SMA	soil-moisture deficit accounting algorithm
SRWMD	Suwannee River Water Management District
USACE	U.S. Army Corps of Engineers
USGS	U.S. Geological Survey

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Abstract

Exchanges of water between the Upper Floridan aquifer and the Lower Suwannee River were evaluated using historic and current hydrologic data from the Lower Suwannee River Basin and adjacent areas that contribute ground-water flow to the lowest 76 miles of the Suwannee River and the lowest 28 miles of the Santa Fe River. These and other data were also used to develop a computer model that simulated the movement of water in the aquifer and river, and surface- and ground-water exchanges between these systems over a range of hydrologic conditions and a set of hypothetical water-use scenarios.

Long-term data indicate that at least 15 percent of the average annual flow in the Suwannee River near Wilcox (at river mile 36) is derived from ground-water discharge to the Lower Suwannee and Lower Santa Fe Rivers. Model simulations of ground-water flow to this reach during water years 1998 and 1999 were similar to these model-independent estimates and indicated that ground-water discharge accounted for about 12 percent of the flow in the Lower Suwannee River during this time period.

The simulated average ground-water discharge to the Lower Suwannee River downstream from the mouth of the Santa Fe River was about 2,000 cubic feet per second during water years 1998 and 1999. Simulated monthly average ground-water discharge rates to this reach ranged from about 1,500 to 3,200 cubic feet per second. These temporal variations in ground-water discharge were associated with climatic phenomena, including periods of strong influence by El Niño-associated flooding, and La Niña-associated drought. These variations showed a relatively consistent pattern in which the lowest rates of ground-water inflow occurred during periods of peak flood levels (when river levels rose faster than ground-water levels) and after periods of extended droughts (when ground-water storage was depleted). Conversely, the highest rates of ground-water inflow typically occurred during periods of receding levels that followed peak river levels.

Introduction

The Suwannee River drains an area of about 9,930 mi^2 in southern Georgia and northern Florida (fig. 1). In much of the river, the flow is intimately linked to ground-water conditions. This is evident in the lower reaches of the river and its main tributary, the Santa Fe River, where some of the highest concentrations of springs in Florida are found (fig. 2). The ground-water inflow that sustains these springs accounts for most of the average annual flow of the Lower Suwannee River, and supplies nearly all of the flow in the river during periods of low flow in the dry seasons and during droughts.

The Lower Suwannee River and adjacent Upper Floridan aquifer are an important water resource in this region of Florida. Freshwater flow in the Lower Suwannee River helps maintain the proper balance of freshwater and saltwater that sustains estuarine life in the Gulf of Mexico near the mouth of the Lower Suwannee River. The river also provides numerous recreational opportunities and important aesthetic benefits to the region. Ground water from the Upper Floridan aquifer supplies all of the water for drinking, irrigation, and commercial uses in the Lower Suwannee River Basin.

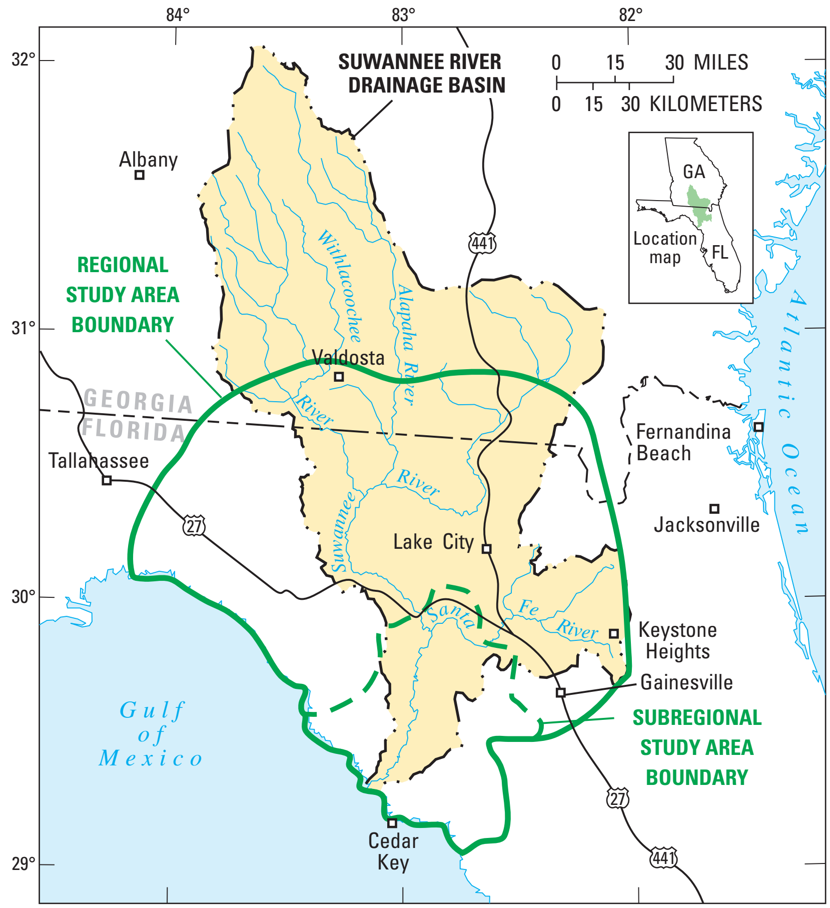


Figure 1. Location of regional and subregional study areas and the Suwannee River Drainage Basin.

The flow of water in the Lower Suwannee River and its springs depends on the ground-water level in the contiguous Upper Floridan aquifer. When ground-water levels are high relative to the river stage, ground-water inflow rates to the river are also high and flow in the river is sustained. As ground-water levels decline relative to river stages, ground-water inflow to the river declines and the flow in the river declines accordingly. Ground-water levels depend on several factors: rainfall that recharges the aquifer, the amount of ground water that is consumed for various uses, and river stage (for locations close enough to the river). Because of these linkages between the river and Upper Floridan aquifer, a good understanding of their hydraulic interaction and the corresponding hydraulic and hydrologic characteristics is essential for managing water in the Lower Suwannee River Basin and assessing the potential consequences of future water development on streamflows.

In 1996, the U.S. Geological Survey (USGS), in cooperation with the Suwannee River Water Management District (SRWMD), initiated a study to evaluate the ground-water and surface-water interactions between the Lower Suwannee River and the contiguous Upper Floridan aquifer. The specific objectives of the study were twofold. The first objective was to integrate historic and current ground- and surface-water data to better understand the hydrology of the river and aquifer and their interaction. Development of a hydrologic model of the river and aquifer systems was an essential element of this integration. The second objective was to use this model to assess the effects of future water-withdrawal scenarios on streamflow.

The primary (subregional) study area covers about 2,300 mi^2 in northern peninsular Florida (figs. 1 and 2), and includes the lowest (most downstream) 76 mi of the Suwannee River from the town of Branford to the Gulf of Mexico (defined here as the Lower Suwannee River) and the lowest 28 mi of the

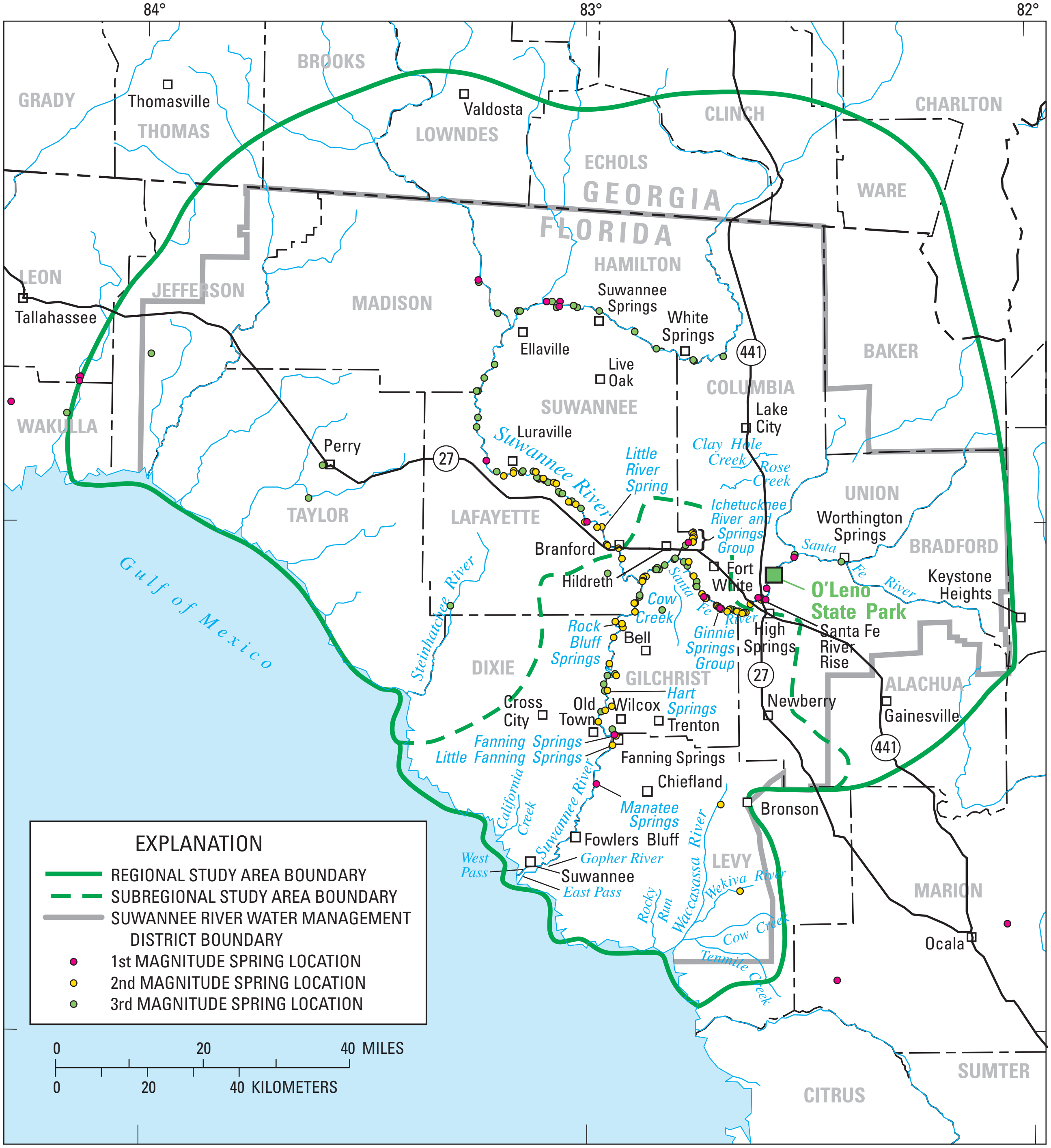


Figure 2. Location of rivers and springs in the subregional and regional study areas and the Suwannee River Water Management District.

Santa Fe River from U.S. Highway 441 near High Springs to the mouth of the river at its confluence with the Suwannee River (defined here as the Lower Santa Fe River). The subregional study area also includes places where the Upper Floridan aquifer contributes ground water to these two river reaches, all of the Lower Suwannee River Basin, and adjacent areas in Lafayette, Dixie, Gilchrist, Alachua, and Levy Counties (fig. 2). Hydrologic and hydrogeologic characteristics of a larger regional study area, encompassing all of the SRWMD (fig. 2), are also examined to provide the regional hydrogeologic setting.

Purpose and Scope

The purposes of this report are to: (1) describe the surface- and ground-water hydrology of the subregional study area; (2) evaluate the magnitude and timing of water exchanges between the Lower Suwannee River and Upper Floridan aquifer using historical and current data and a hydrologic model; and (3) describe how a hydrologic model for simulating these exchanges could be used to evaluate changes in streamflow under hypothetical water-use scenarios. This report describes the analysis of ground-water and surface-water data collected from August 1996 to September 1999, and available historic data. The report also describes the development and application of a model to simulate streamflow, ground-water flow, and river-aquifer exchanges.

Description of Study Areas

The regional study area and all of Florida lie within the Coastal Plain physiographic province (Fenneman, 1938). Puri and Vernon (1964) presented a detailed map of the physiographic divisions within Florida, and identified three major physiographic divisions within or adjacent to the study area: the Northern Highlands, Central Highlands, and Gulf Coastal Lowlands. The Northern Highlands and Gulf Coastal Lowlands make up most of the regional study area (fig. 3).

The Northern Highlands typically has broad, gently sloping, and generally continuous high elevation plateaus in its interior regions, and marginal slopes that are well drained by dendritic streams. The Central Highlands is also characterized by broad, generally coast-parallel high elevation areas, some of which have been divided into distinct areas of elongated ridges, separated by low elevation uplands and broad valleys (Puri and Vernon, 1964). The Brooksville Ridge is the most prominent example of the Central Highlands ridges in the regional study area.

Most of the subregional study area is within the Gulf Coastal Lowlands, a region of generally coast-parallel terraces and ancient shorelines that slope gently from the Northern Highlands and Central Highlands toward the coast. Relict barrier islands that form sand ridges, such as Bell Ridge, and that are commonly underlain by karst limestone are present in the Gulf Coastal Lowlands (Puri and others, 1967). Limestone is at or near land surface over much of this area, and karst topographic features are common. Other features of the Gulf Coastal Lowlands include: (1) extensive areas of poorly drained swamps and wet-pine flatwoods; (2) Lower Suwannee River and Lower Santa Fe River valleys which, apart from the two main rivers and the numerous springs that feed them (fig. 2), are nearly devoid of surface drainage; and (3) coastal areas that are drained by a network of sluggish streams, coastal swamps, and salt marshes.

The boundary between the Northern Highlands and Gulf Coastal Lowlands is defined by the Cody Scarp, which is the most persistent topographic break or escarpment in Florida (Puri and Vernon, 1964). This escarpment is also approximately coincident with the northeastern boundary of the subregional study area and the boundary between confined and unconfined areas of the Upper Floridan aquifer (Miller, 1986). Many of the streams draining the Northern Highlands are captured by sinkholes near the margins of the Northern Highlands and reemerge below the Cody Scarp (Burnson and others, 1984). Two prominent examples of stream capture and reemergence occur in the vicinity of the subregional study area. The first is the Santa Fe River, which drains into a sinkhole at O'Leno State Park and reemerges about 3 mi southwest of the sinkhole at the Santa Fe River Rise (fig. 2). The second example is the springfed Ichetucknee River (fig. 2), which occurs at the downstream end of an abandoned drainage way (Ichetucknee Trace). Intermittent runoff from two ephemeral streams (Clay Hole Creek and Rose Creek) flows into two large sinkholes within this drainage way, about 7 to 9 mi northeast of the Ichetucknee River Springs Group (fig. 2).

The climate of the study areas is humid subtropical (Raulston and others, 1998). Monthly average temperatures typically range from 54 to 57 °F in the winter and from 79 to 91 °F in the summer (National Oceanographic and Atmospheric Administration, 2002). Average annual precipitation ranges from about 51 to 59 in/yr, with about half of this amount typically occurring from June to September (National Oceanographic and Atmospheric Administration, 2002). Summer precipitation is generally associated with localized thunderstorm activity that can produce intense rainfall. Winter precipitation is generally associated with the passage of cold fronts and is more evenly distributed geographically. Average-annual evapotranspiration estimates in the SRWMD range from about 35 to 41 in/yr (Bush and Johnston, 1988, pl. 9; Knowles, 1996).

Most of the subregional study area is sparsely populated. The most densely populated areas are the towns of Cross City (1,800), Chiefland (2,000), Newberry (3,300), High Springs (3,900) Trenton (1,600), and Bronson (1,000) shown in figure 2. Primary economic activities in the study area are silviculture, the manufacture of forest products, and agriculture. Accordingly, forest and agricultural lands account for most of the land use in the study area, although wetlands also cover a large part of the study area. Agricultural land use is found in the better drained areas of eastern Lafayette, eastern Dixie, Gilchrist, northeastern and northwestern Levy County, southern Suwannee, southern Columbia, and western Alachua Counties.

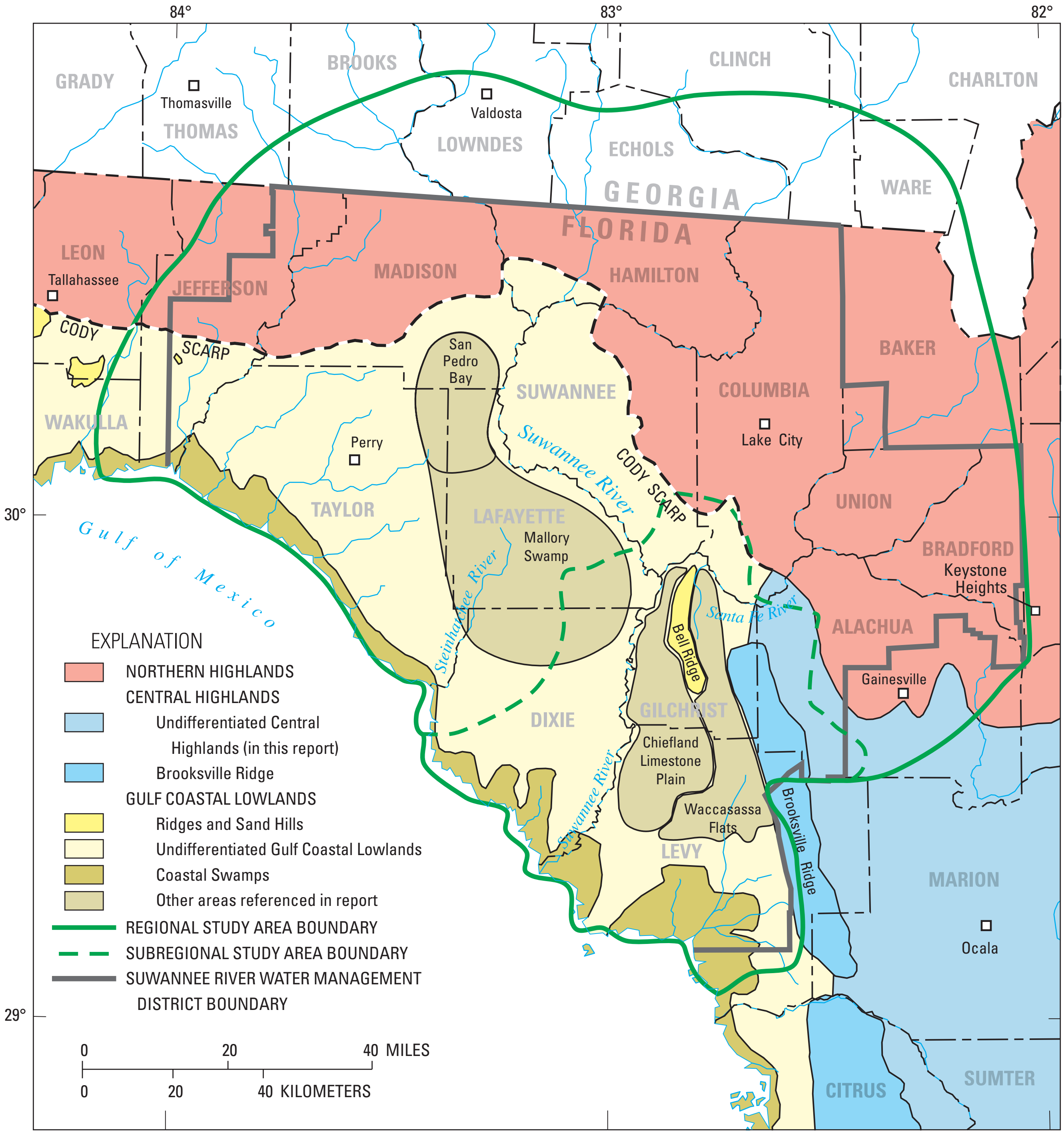


Figure 3. Physiographic areas in and adjacent to the Suwannee River Water Management District.

Previous Studies

Clarke (1965) studied the relation between ground-water inflow, bank and channel storage, and change in streamflow along a 6.8-mi reach of the Lower Santa Fe River between gaging stations near High Springs (map index no. 13, fig. 4 and table 1) and near Fort White (map index no. 14, fig. 4 and table 1). In this reach, streamflow typically increases from about 850 ft^3/s at the upstream station near High Springs to about 1,600 ft^3/s at the downstream station near Fort White, respectively. Clarke (1965) noted that the difference in streamflow (streamflow pickup) at the upstream and downstream ends of this river reach changes slowly over seasonal and longer time scales. Clarke (1965) observed that the streamflow pickup in this reach can decrease abruptly, and even reverse to a difference greater than 1,000 ft^3/s, over shorter time scales when water levels in the Santa Fe River increase rapidly following periods of increased rainfall. After this abrupt decrease, the streamflow pickup subsequently increases to a rate greater than that occurring before the rise in river levels. Clarke (1965) did not determine how much of these short-term changes in streamflow pickup were due to changes in the surface storage (water temporarily stored in the river channel and overbank areas) and changes in the rates of water exchanges between the river and aquifer in this reach.

Pittman and others (1997) evaluated spring contributions to water quantity and nitrate loads in a 33-mi reach of the Suwannee River just upstream from the subregional study area between Ellaville and Branford (fig. 4) during a period of low flow in the river. They concluded that all of the 950-ft^3/s increase in flow in the reach was derived from ground water. Of this total increase in flow, 41 percent was derived from 11 springs that were measured. When unmeasured springs are considered (those known at the time of their study and those subsequently discovered), 75 to 85 percent of the flow increase that was measured in this reach was probably due to springflow contributions.

The chemical composition of surface and ground waters has also been used to evaluate interactions between ground and surface waters in the study areas. Crandall and others (1999) found evidence for the migration of water from the Suwannee River into the Upper Floridan aquifer by observing changes in the chemical composition of water obtained from wells, sinkholes, and springs during a period of rising water levels in the Suwannee River. Evidence for this mixing included decreased concentrations of calcium, silica, and radon-222 (^{222}Rn, a naturally occurring isotope of radon) and increased concentrations of dissolved organic carbon, tannic acid, and chloride. The fraction of river water in ground water ranged from 13 to 65 percent at a sinkhole located about 1 mi from the Suwannee River. Tannic-colored river water was also observed flowing into Little River Spring (fig. 2), which discharges into the Suwannee River at rates from 50 to 230 ft^3/s at lower river stages (Crandall and others, 1999).

Ellins and others (1991) and Kincaid (1998) used ^{222}Rn and other tracers to evaluate mixing of water between the Lower Santa Fe River and the Upper Floridan aquifer. Kincaid (1998) found that appreciable quantities of river water can rapidly infiltrate into the aquifer. The degree of mixing of river and aquifer waters was also highly variable. An analysis of samples collected from the Devil's Ear cave system (fig. 2, part of the Ginnie Springs Group), which discharges about 300 ft^3/s to the Lower Santa Fe River, indicates that river water represented from less than 10 percent to slightly greater than 60 percent of the sample. Kincaid (1998) also described a relation between the degree of mixing and the distribution of rainfall within the Santa Fe River Basin. When large rainfall events occur primarily in the confined areas of the Northern Highlands (Upper Santa Fe River Basin), the resulting storm runoff from the Upper Santa Fe River Basin creates a downward (river-to-aquifer) head gradient and more mixing of river and aquifer water. Conversely, when rainfall occurs primarily in the Lower Santa Fe River Basin, an upward (aquifer-to-river) head gradient and less mixing of river and aquifer water results.

Crane (1986) studied the degree of ground-water/surface-water interactions in the Suwannee River by evaluating the: (1) relation between streamflow and specific conductance in river water; and (2) activity ratios of uranium isotopes ($^{234}U/^{238}U$) in river, spring, and aquifer waters. Crane (1986) found an inverse relation between streamflow and specific conductance in the Suwannee River at and downstream from White Springs, and further noted that the relation became "...progressively evident downstream as more ground water enters the river and runoff contributions to the river become negligible at Suwannee Springs, Branford, and Wilcox." The "increasing downstream importance of ground-water contributions to the flow of the Suwannee River" was also evident in his analysis of $^{234}U/^{238}U$ ratios. Crane (1986) found that from White Springs southward, "...ground-water flow from areas of moderate to high recharge completely dominates the character of the river water." Similar conclusions were made in an earlier study by Hull and others (1981), reporting downstream trends in pH, total organic carbon, specific conductance, and inorganic chemical constituents.

Grubbs (1997) used streamflow and specific conductance measurements of streamflow, direct (storm) runoff, and ground water to evaluate the ground-water/surface-water exchanges in a reach of the Lower Santa Fe River between gaging stations near Worthington Springs and Fort White. Grubbs concluded that ground-water discharge to the river accounts for all, or nearly all, of the increase in flow in this reach of the Santa Fe River on an average annual basis. Evaluations of the changes in streamflow, which occur along reaches of the Suwannee and Santa Fe Rivers, also indicated that most of the increase in the streamflow pickup, which occurs along the Suwannee River (below Ellaville) and the Lower Santa Fe River, is derived from ground water discharging to these river reaches.

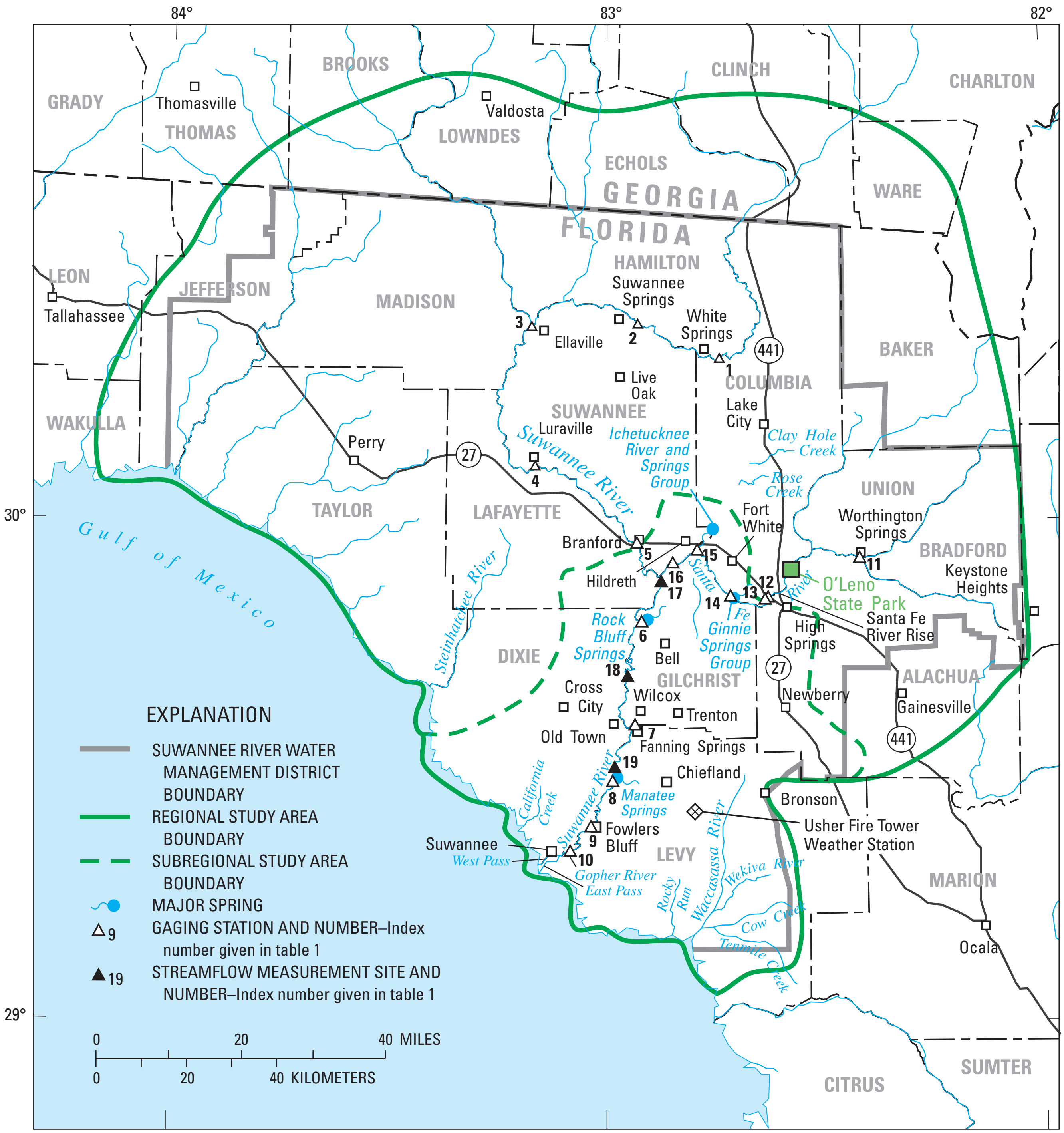


Figure 4. Selected streamgaging stations and streamflow measurement sites on the Suwannee and Lower Santa Fe Rivers.

Table 1. Selected streamgaging stations and streamflow measurement sites on the Lower Suwannee and Santa Fe Rivers.

Site No. (this report only)	Station number	Station name
1	02315500	Suwannee River at White Springs
2	02315550	Suwannee River at Suwannee Springs
3	02319500	Suwannee River at Ellaville
4	02320000	Suwannee River at Luraville
5	02320500	Suwannee River at Branford
6	02323000	Suwannee River near Bell
7	02323500	Suwannee River near Wilcox
8	02323570	Suwannee River near Old Town
9	02323590	Suwannee River at Fowlers Bluff
10	02323592	Suwannee River above the Gopher River
11	02321500	Santa Fe River at Worthington Springs
12	02321975	Santa Fe River at U.S. Highway 441 near High Springs
13	02322000	Santa Fe River near High Springs
14	02322500	Santa Fe River near Fort White
15	02322703	Santa Fe River at Three Rivers Estates
16	02322800	Santa Fe River near Hildreth
17	295309082523801	Suwannee River below the mouth of the Santa Fe River
18	294041082571701	Suwannee River upstream from Hart Springs Run near Wilcox
19	292940082590301	Suwannee River above Manatee Springs near Chiefland

Acknowledgments

The authors wish to thank the numerous volunteers from the SRWMD and the hydrologic technicians from the Orlando, Tampa, and Tallahassee USGS offices for their hard work during several field trips when streamflow was measured continuously (in some instances, for one or more days at a time) at several locations on the Lower Suwannee and Lower Santa Fe Rivers and associated springs. We also acknowledge the work performed by the SRMWD field staff and by Darlene Blum and Lavetra Sloan of the USGS, who measured ground-water levels during the study. Ron Ceryak provided essential information concerning the details and history of data collection by the SRWMD, as well as valuable insight into the hydrogeology of the SRWMD, particularly in the Waccasassa Flats area. Ron Ceryak and Donald Monroe also provided ground-water level and other data from the SRWMD database that was essential to the success of this project. Marvin Franklin gave important information about the history of the interpretation and collection of streamflow and water-level data in the Lower Suwannee and Lower Santa Fe Rivers. The Division of Forestry of the Florida Department of Agriculture was helpful, by granting data-collection access at fire tower and work center sites. The Natural Resources Conservation Service contributed timely information about the soils in Dixie County, as well as crop acreage estimates that were required to estimate irrigation withdrawals of ground water. Keith Halford provided valuable assistance by evaluating some of the historic aquifer test results. John Doherty of Watermark Numerical Computing conveyed timely and valuable suggestions, regarding use of parameter estimation software for calibrating the ground-water flow model. The authors appreciate the work of Richard Marella of the USGS, who compiled much of the water-use data used in the project. Finally, the authors are grateful for the contributions that the technical reviewers made in improving the project analyses and final report. The primary project reviewers, Bob Faye (USGS), John Good (SRWMD), Eve Kuniansky (USGS), Gary Mahon (USGS), Eric Swain (USGS), and Dann Yobbi (USGS), provided invaluable guidance on all aspects of the study, particularly in refining the data-collection plans, improving the way the boundary conditions were treated in the numerical model, and calibration of the simulation models.

Surface-Water Hydrology

Perhaps the most important aspect of the surface-water hydrology of the subregional study area is that much of the study area is devoid of channelized surface drainage and most of the drainage occurs instead through the subsurface. Subsurface drainage occurs because of the: (1) karst topography of the area, which is generally flat and contains numerous sinkholes and closed topographic depressions; and (2) highly permeable rocks of the Upper Floridan aquifer, which are generally either exposed at the surface or overlain by a relatively shallow layer of permeable surficial sediments. These two characteristics allow rainfall to easily move into and through the subsurface and discourage the formation of organized surface drainage networks.

Despite the dominance of subsurface drainage in the subregional study area, some surface drainage does occur. The most important surface drainage features are the Lower Suwannee and Lower Santa Fe Rivers. Smaller streams are also present in some areas, especially along the coast, and some can convey the discharge of springs. Much of the subregional study area lacks good surface or subsurface drainage; these areas are typically covered by broad wetlands and wet-pine flatwoods that are commonly bordered by shallow lakes.

Lower Suwannee and Lower Santa Fe Rivers

The Lower Suwannee and Lower Santa Fe Rivers are the major rivers that drain the subregional study area. In this report, the Lower Suwannee River is defined as the 76-mi reach of the Suwannee River that begins upstream at Branford and extends downstream to the mouth of the river at the Gulf of Mexico (fig. 4). The Lower Santa Fe River is defined in this report as the 28-mi reach of the Santa Fe River, beginning just upstream from U.S. Highway 441 about 2 mi downstream from the Santa Fe River Rise (figs. 1 and 4) where it emerges from the Upper Floridan aquifer, extending downstream to its confluence with the Suwannee River, about 10 river miles downstream from Branford. The subsequent sections describe the geometry of the channels and floodplains of the Lower Suwannee and Lower Santa Fe Rivers and their flow and stage characteristics. Flow and water-level data were collected from gaging stations on the Suwannee and Lower Santa Fe Rivers and from additional streamflow-measurement sites that were established during this study (fig. 4 and table 1).

Channel and Floodplain Characteristics

Channel and floodplain geometry affect the relation between flow and stage in various reaches of the Lower Suwannee and Lower Santa Fe Rivers. This relation is important because both rivers represent an important boundary within the ground-water flow system in the subregional study area. Variations in streamflows and the resulting changes in river stages play a major role in the timing of ground-water/surface-water exchanges along these two rivers. Temporal and spatial patterns of river stages are also an important influence on ground-water levels (for example, the configuration of the potentiometric surface) and directions of ground-water flow.

Most of the river channel and floodplain geometry data for the present study were obtained from previous research and data files (U.S. Army Corps of Engineers, 1989; John Good, U.S. Army Corps of Engineers, written commun., 1996). The U.S. Army Corps of Engineers (USACE) used these data to develop a HEC-2 water-surface profile model (Hydrologic Engineering Center, 1991) of the Lower Suwannee River and its largest tributaries. Other data were obtained from field measurements taken at a few sites during the present study and from discharge measurement notes, which contain depth and stage data. These data were used to estimate the elevation of the channel bottom, changes in the channel-bottom elevation along the axis of the channel thalweg (the deepest part of the river channel), channel top width at bankfull stage, and width of the floodplain at flood stages.

The elevation and slope of the channel bottom, as measured along the channel centerline, are important variables that influence the slope of the water surface. These variables also help determine whether the elevation of the water surface at a given location is affected by downstream water conditions, such as tides or flooding at a downstream confluence. When this condition occurs, the location is said to be affected by “backwater.” Data for the Lower Suwannee River indicate that the elevation of the deepest part of the channel (the thalweg elevation) near the Branford gaging station is close to sea level or NGVD 1929 (fig. 5). The thalweg elevation is highly variable. In some instances, elevations range from about 30 to 5 ft below NGVD 1929 over short distances. Despite this variability, a trend of generally decreasing thalweg elevation in the downstream direction is evident in the data, and is on the order of 0.24 ft/mi (fig. 5). This thalweg elevation slope is mild, and as a result, the flow of the Santa Fe River can affect the stage in the Lower Suwannee River upstream from the mouth of the Santa Fe River. Additionally, the low elevation of the thalweg, relative to NGVD 1929, allows the tides in the Gulf of Mexico to affect the stage in much of the Lower Suwannee River downstream from the mouth of the Santa Fe River, especially during periods of low flow.

The thalweg profile of the Lower Santa Fe River is generally steeper than that of the Lower Suwannee River, with the reach upstream from the gaging station near Fort White having an average channel-bottom slope of about 1.4 ft/mi. The thalweg elevation at the upstream end of the Lower Santa Fe River at U.S. Highway 441 is about 26 ft above NGVD 1929, and the thalweg elevation near the mouth is about 10 ft below NGVD 1929 (fig. 5). Under average flow conditions, most of the Santa Fe River downstream from the gaging station near Fort White is affected by backwater from the mouth of the Suwannee River. Some of the reaches at and upstream from the Fort White gaging station have sections with shoals at low flows; however, all of the Lower Santa Fe River may be subject to backwater from the Suwannee River during periods of flooding.

The Lower Suwannee River has an average channel top width at bankfull stage (when all of the river channel is submerged) of about 600 ft. The bankfull channel top width is as narrow as 250 ft near the Branford gaging station and as wide as 1,000 ft just upstream from the bifurcation of the river into East Pass and West Pass. The bankfull channel top width of the Lower Santa Fe River averages about 200 ft and ranges between 125 and 750 ft. Analysis of the channel geometry data also indicates that the width of the 10-year floodplain of

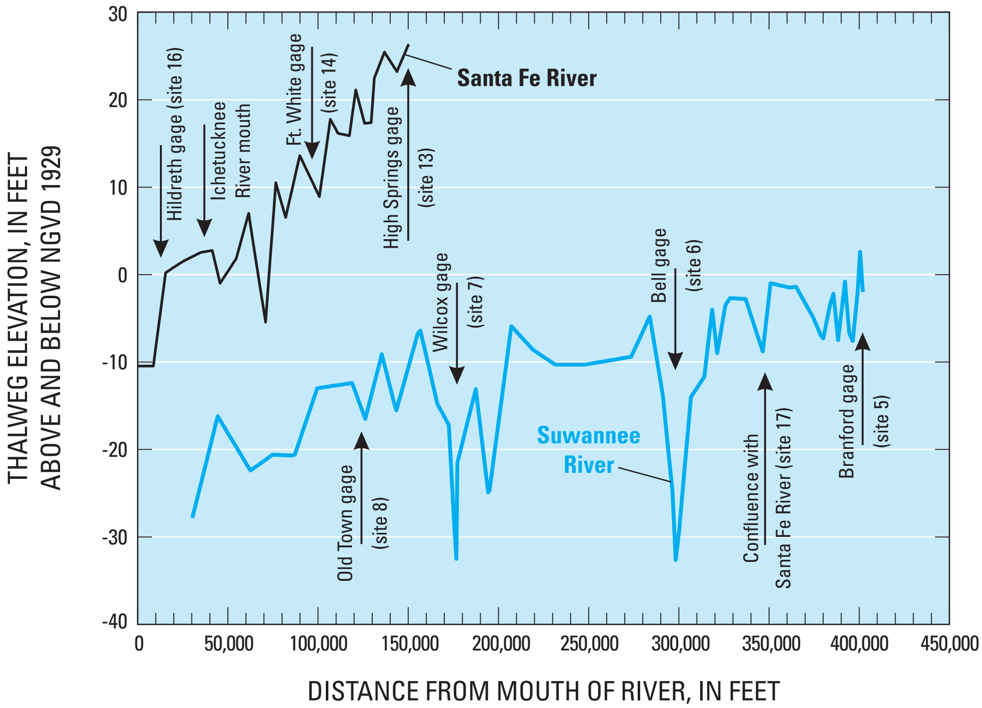


Figure 5. Thalweg profiles of the Lower Suwannee and Lower Santa Fe Rivers. Site locations are shown in fig. 4.

the Lower Suwannee River ranges from about 3,000 ft in the narrower parts of the reach upstream from the mouth of the Santa Fe River to about 21,000 ft about 12 mi upstream from the mouth of the Suwannee River. The width of the 10-year floodplain of the Lower Santa Fe River ranges from about 450 ft in the narrower reaches of the river upstream from the Fort White gaging station to about 5,800 ft near the mouth.

Flow Characteristics

Long-term records of river stages and flow are available at several gaging stations on the Lower Suwannee, Lower Santa Fe, and (to a lesser extent) Ichetucknee Rivers. The gaging stations on the Suwannee River at Branford (map index no. 5, fig. 4) and the Santa Fe River near High Springs (map index nos. 12 and 13, fig. 4) were used to estimate the amount of upstream streamflow entering the Lower Suwannee and Lower Santa Fe Rivers. These data indicate that the flow at the upstream boundary of the Lower Suwannee River ranged from 1,400 to 84,000 ft^3/s, with an average annual flow of about 7,000 ft^3/s during 1931-99. Flow at the upstream boundary of the Santa Fe River ranged from 30 to 20,000 ft^3/s, with an average annual flow of about 850 ft^3/s during the period of record (1931-71) for the High Springs gaging station (map index no. 13, fig. 4).

There is a noticeable relation between streamflow and time of year, and the nature of this relation varies within the subregional study area. In the Suwannee River at Branford, flows are typically higher from February to April than during the remaining months of the year (fig. 6). The highest monthly mean flows typically occur during March and April. The timing of these high flow periods is consistent with climatic data, which indicate that March is the last month in a period (beginning in June) when precipitation typically exceeds potential evapotranspiration in the northern peninsula. Although rainfall rates are high during the summer-early fall (June-September) period, streamflows are typically not as high as during the late winter-spring period (February-April) because of low net precipitation rates (precipitation minus evapotranspiration) from April to June.

A different seasonal pattern is evident in the streamflow data for the Lower Santa Fe River. For example, the monthly mean streamflows in the Santa Fe River near Fort White are highest from February to April and from August to October (fig. 6). The surface- and ground-water contributing areas to the Lower Santa Fe River are south of the contributing areas to the Suwannee River at Branford, and therefore, less influenced by rainfall from cold fronts, which weaken as they move southward into peninsular Florida. The higher mean flows in August and September coincide with the summer wet season. The Santa Fe River streamflows are most variable during the high-flow periods from February to April and from August to October.

The seasonal patterns in the streamflow data for the Suwannee River downstream from the mouth of the Santa Fe River are a mixture of the seasonal patterns of the Suwannee River at Branford and the Lower Santa Fe River. The seasonal distribution of monthly mean discharges in the Suwannee River near Wilcox (fig. 6) is most similar to the Suwannee River at Branford, with the highest flows generally occurring from February to April. This distribution is to be expected because average flow in the Suwannee River at Branford is about 70 percent of that in the Suwannee River near Wilcox, and average flow in the Santa Fe River near Fort White is about 15 percent of that in the Suwannee River near Wilcox.

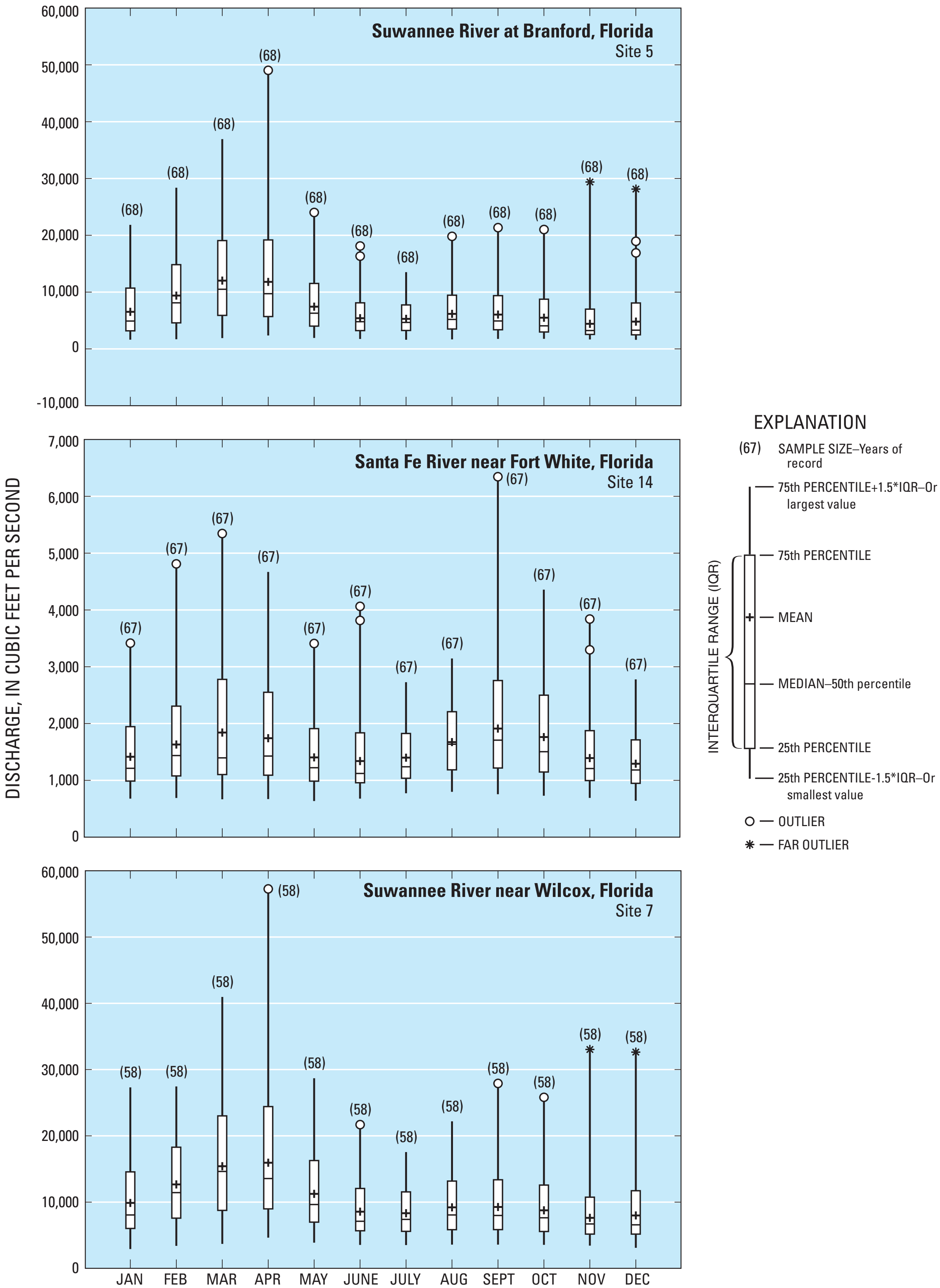


Figure 6. Monthly mean discharge in the Lower Suwannee and Lower Santa Fe Rivers. Site locations are shown in fig. 4.

The remaining 15 percent is derived from ground-water discharge to the Lower Suwannee and Lower Santa Fe Rivers.

Year-to-year variations in the streamflows that occur in a given month or season are even greater than the previously described seasonal variation in streamflows. This interannual variability occurs, in part, because streamflows in the Suwannee and Santa Fe River Basins are subject to wetter-than-normal and drier-than-normal periods associated with climatic phenomena, such as the El Niño-La Niña Southern Oscillation (National Oceanic and Atmospheric Administration, 2005) and possibly the Pacific Decadal Oscillation (Nigam and others, 1999). During the present study, for example, there was a transition from El Niño (from 1997 to mid-1998) to La Niña (from mid-1998 continuing into the first part of 2000) conditions. These two disparate climatic conditions resulted in different flow conditions in the subregional study area during the "high-flow" months. The El Niño period from February to April 1998 was characterized by extreme flooding on the Suwannee and Santa Fe Rivers (fig. 7). The peak daily streamflow in the Suwannee River near Wilcox was close to 50, 000 ft^3/s — a flow that typically occurs only once every 20 to 25 years. During La Niña conditions of the following year, however, flows during these "high-flow" months were as low or lower than the values that typically occur during the dryer months of the year. Tropical weather systems occur frequently in the study area; there is a 24- to 36-percent chance of a tropical storm or hurricane passing over the study area in a given year. These weather systems also increase the variability of monthly streamflow during the hurricane season (June to November).

Short-term variations in streamflow in the lower reaches of the Lower Suwannee River are associated with the ebbing and flooding of tides in the Gulf of Mexico. At the mouth of the Suwannee River, the Gulf of Mexico typically has a mixed semidiurnal pattern in which two high tides of unequal height, and two low tides of unequal height typically occur each day. The mean tidal range (between the higher high tide and lower low tide) of water levels recorded at the nearby Cedar Key (fig. 1) tidal gage was 3.7 ft from October 1, 1997, to September 30, 1999 (data obtained from National Oceanographic and Atmospheric Administration, 2004). The tidal range is generally higher during the "spring tides" that occur twice each month, coinciding with the new and full phases of the moon. During some of these periods, the tidal range can be close to 5 ft. The tidal range is lower (close to 2 ft) during the "neap tides" that occur twice each month, coinciding with the first and third quarter phases of the moon.

The effects of tides on streamflows are shown in figures 8 and 9. The two examples occurred within 7 days of each other, during spring tide and neap tide conditions, and the mean daily flow in the river (averaged over a day) was nearly the same during both periods. During spring tide conditions, the tidal range was 3.8 ft near the mouth of the Suwannee

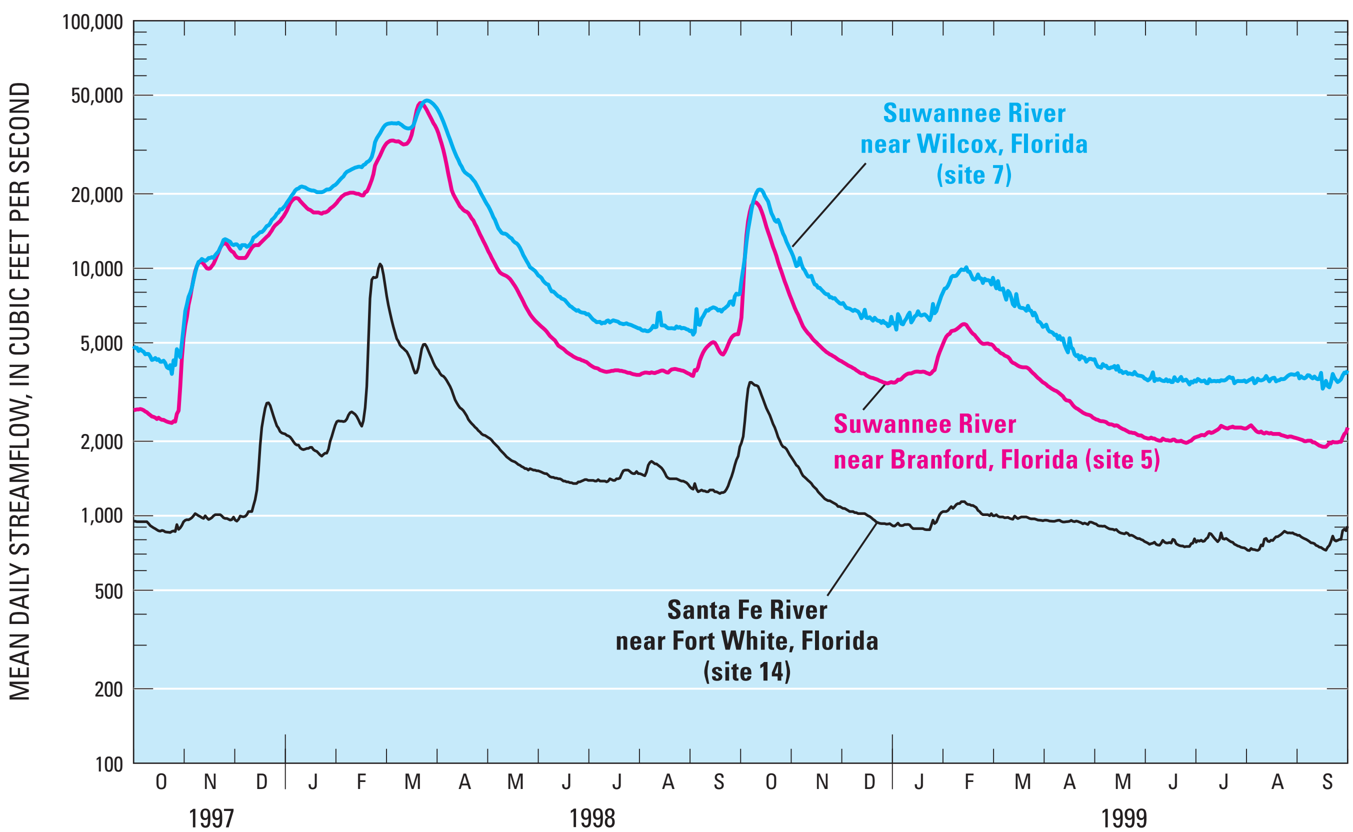


Figure 7. Streamflows in the Lower Suwannee and Lower Santa Fe Rivers from October 1, 1997, to September 30, 1999. Site locations are shown in fig. 4.

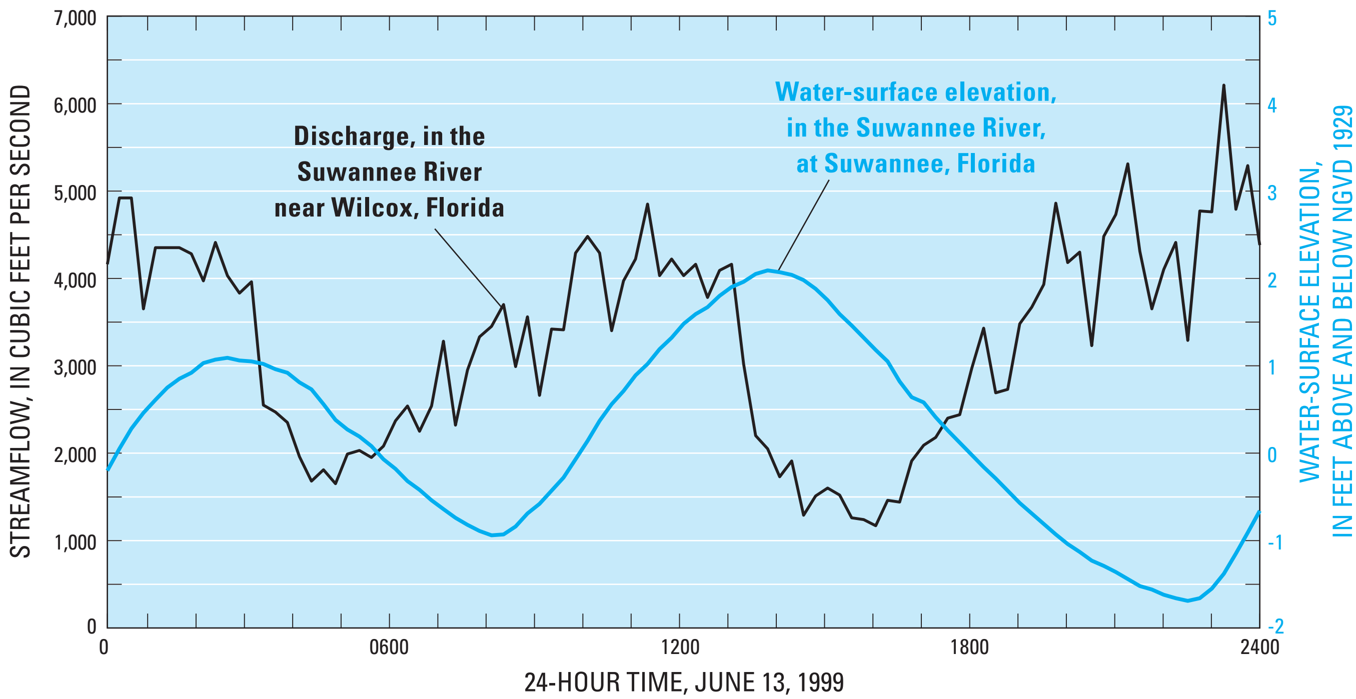


Figure 8. Tidally affected water levels and flows in the Suwannee River during a spring tide.

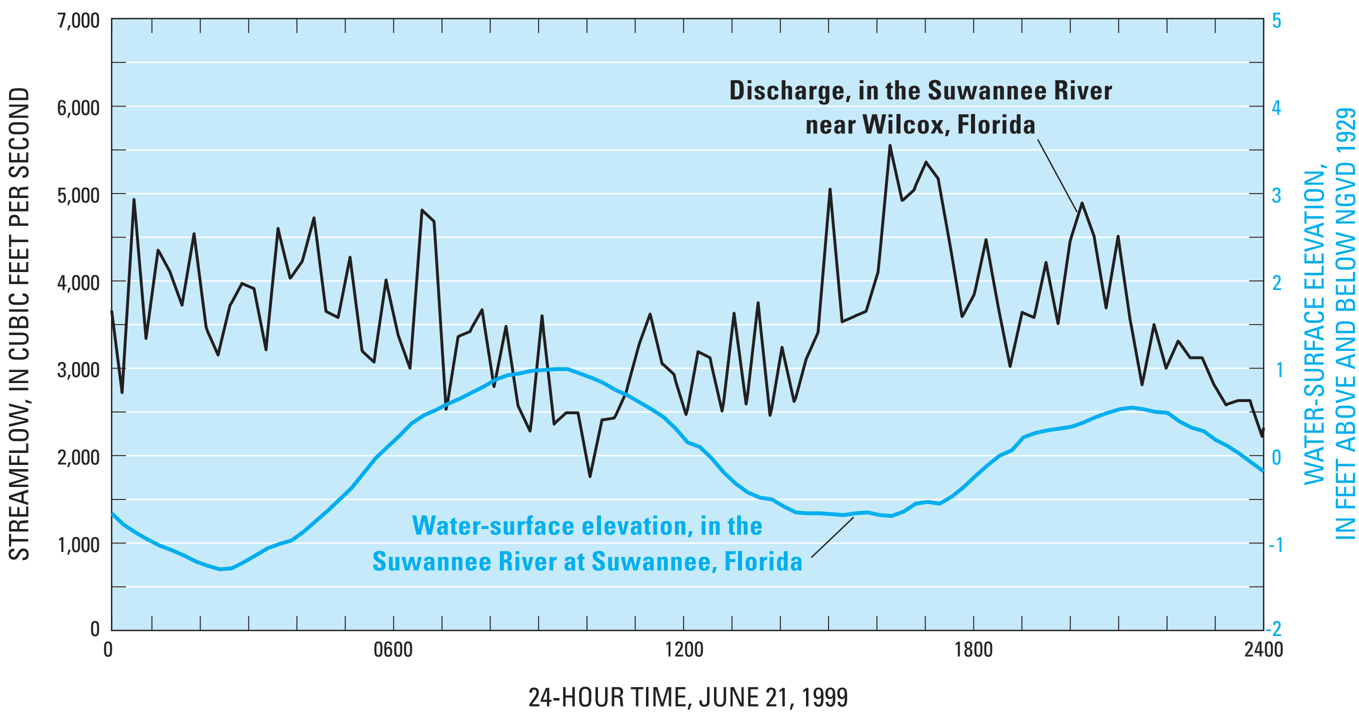


Figure 9. Tidally affected water levels and flows in the Suwannee River during a neap tide.

River, and the observed flow range in the river upstream near Wilcox was 5,000 ft^3/s (from about 1,200 to 6,200 ft^3/s) over the tidal cycle. A week later during neap tide conditions, the tidal range was only 2.3 ft, and the observed flow range in the river upstream near Wilcox was 3,800 ft^3/s (from about 1,800 to 5,600 ft^3/s) over the tidal cycle.

The effects of tides on flow in the Lower Suwannee River can also be enhanced or mitigated by the amount of water flowing in the Suwannee River. For example, the difference between the minimum and maximum flows in the Suwannee River near Wilcox is typically 2,000 to 3,000 ft^3/s over the course of a tidal cycle during low-flow conditions. This range of fluctuation, however, decreases as the average daily discharge increases in the river (fig. 10). When the mean daily discharge is greater than 18,000 ft^3/s in the Suwannee River near Wilcox, the difference between the minimum and maximum flows during a tidal cycle is generally small or negligible (fig. 10). Sustained winds, especially when blowing in an offshore or onshore direction, can also enhance or mitigate the effects of tides on short-term variations in the flow of the Lower Suwannee River.

The effects of these tidal variations also vary with distance to the Gulf of Mexico. For example, during the extreme low-flow period of late September 1999 (when mean daily discharge at the Wilcox gaging station was about 3,800 ft^3/s, measured flow in the Suwannee River above the Gopher River (map index no. 10, river mile 7.5, fig. 11) ranged from -5,000 ft^3/s (where the negative value indicates net upstream flow) to 15,000 ft^3/s over the course of the tidal cycle. The difference between the minimum and maximum flows, however, was substantially smaller upstream during this same period. For example, flows ranged from 1,600 to 5,000 ft^3/s in the Suwannee River at the gaging station near Wilcox (map index no. 7, river mile 33.5, fig. 11) and from 3,700 to 4,100 ft^3/s farther upstream at the gaging station near Bell (map index no. 6, river mile 56.5, fig. 11). The extent of this tidal influence is reduced at higher flows. For example, during mid-August 1997 when the mean daily discharge in the Suwannee River near Wilcox was about 10,000 ft^3/s, a tidal signal was difficult to detect in the streamflows measured at a site just upstream from Hart Springs (near river mile 43, fig. 11).

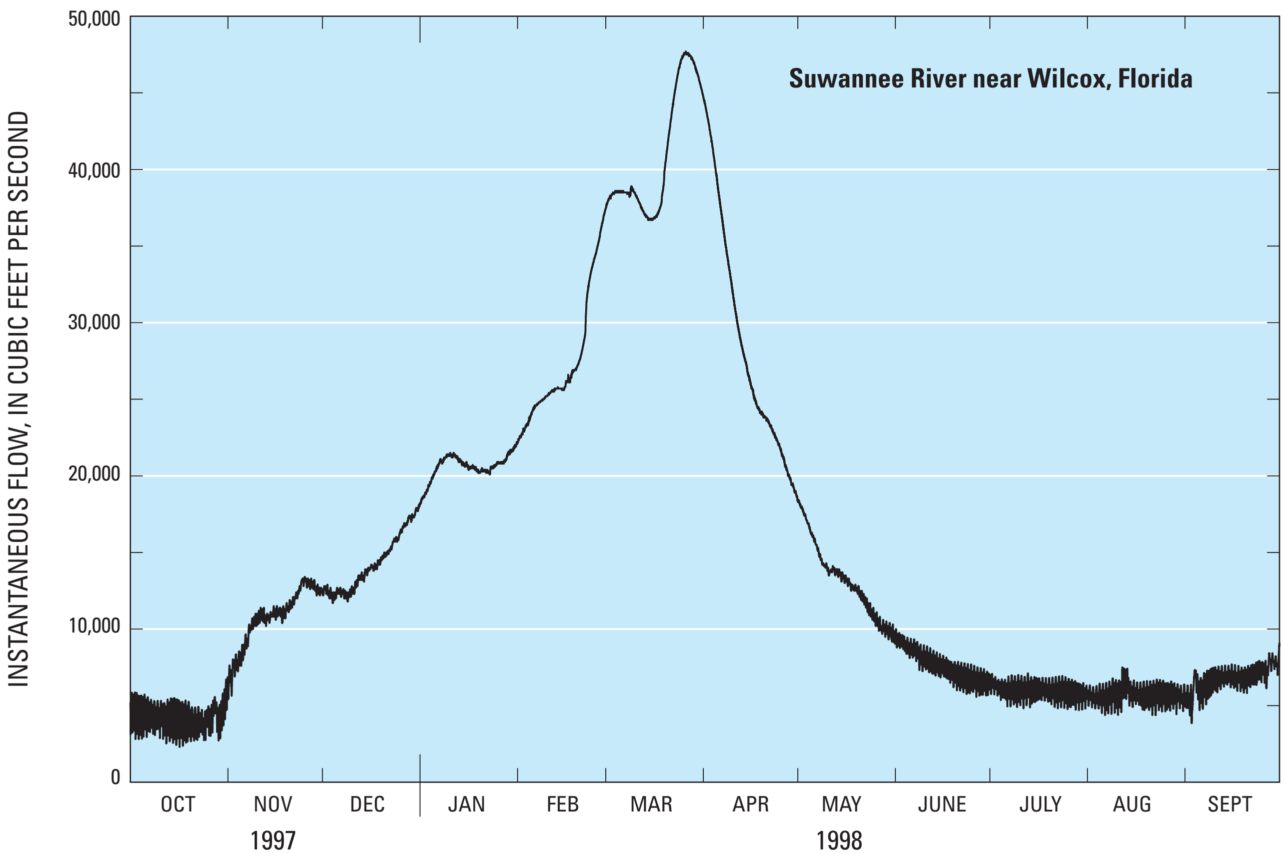


Figure 10. Flow in the Suwannee River near Wilcox, Florida, during water year 1998. Tidally induced discharge fluctuations are greatest at low flows and dissipate as flow in the river increases.

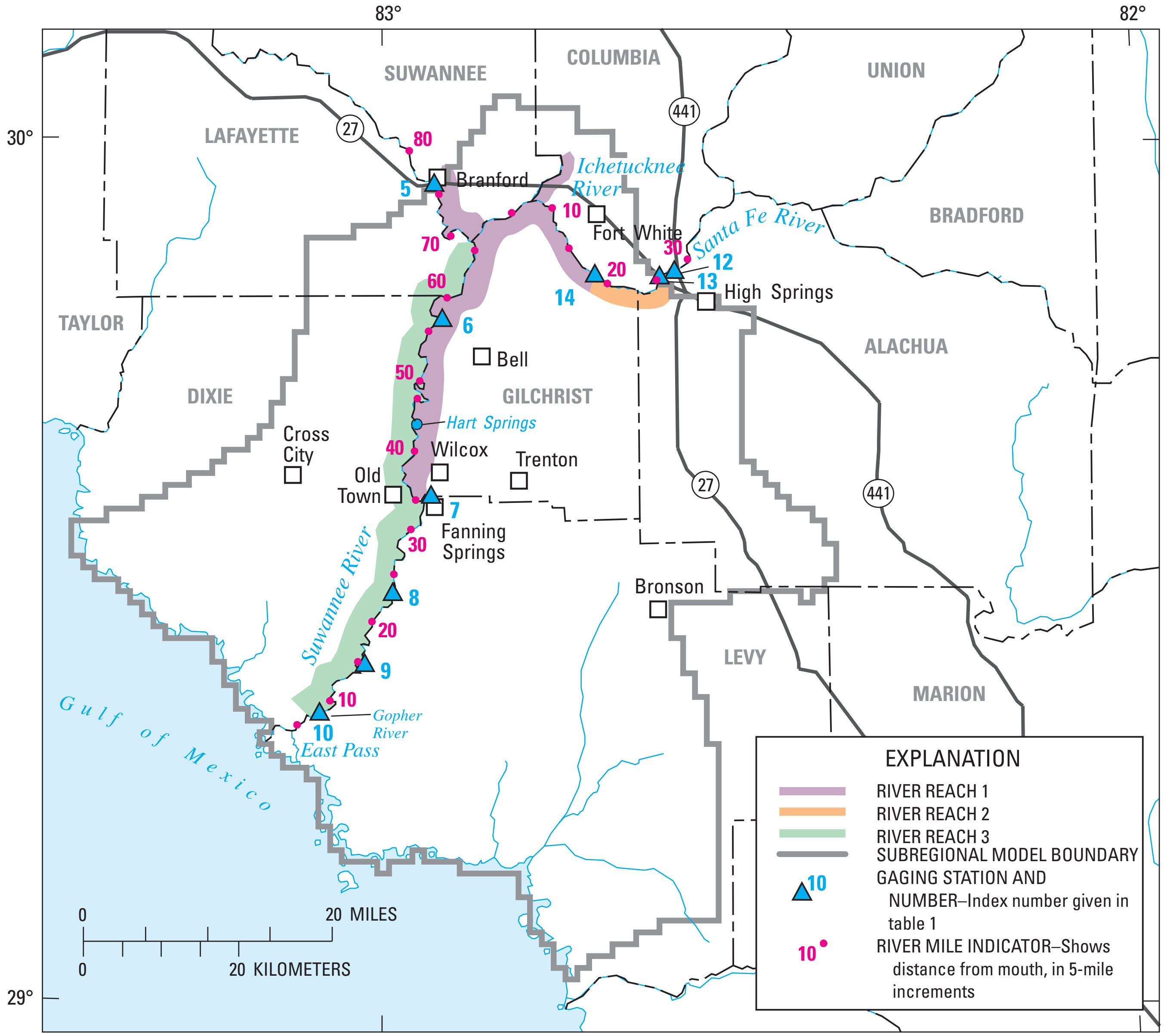


Figure 11. Definition of river mileage system and selected river reaches in the subregional study area.

Water-Level Characteristics

Water levels in the Lower Suwannee and Lower Santa Fe Rivers are associated with interannual, seasonal, and (for the lower reaches of the Lower Suwannee River) tidal variations in flow. Water-level characteristics are also affected by the geometry of the river channels and floodplains that convey the water flowing in the Lower Suwannee River.

The correlation between water levels and streamflow is evident in the seasonal distribution of water levels (fig. 12). The pattern of the monthly average water levels on the Lower Suwannee and Lower Santa Fe Rivers is similar to the pattern of the monthly average flows on these rivers. The Lower Suwannee River sites have the highest water levels from February to April, and the Lower Santa Fe River has a pattern of high water levels from February to April and from August to October. In addition to their association with higher flow in the Santa Fe River, high water levels in the Santa Fe River are also caused by high water levels on the Suwannee River because of the aforementioned backwater effects.

As previously described, flow in most of the Lower Suwannee River is affected by tides in the Gulf of Mexico. Water-level changes associated with these tides propagate upstream into the Lower Suwannee River, causing changes in the slope of the water surface, and thus, streamflow during a given tidal cycle. The magnitude of the water-level changes decreases as the distance from the Gulf of Mexico increases and as streamflow increases. For example, during spring tide

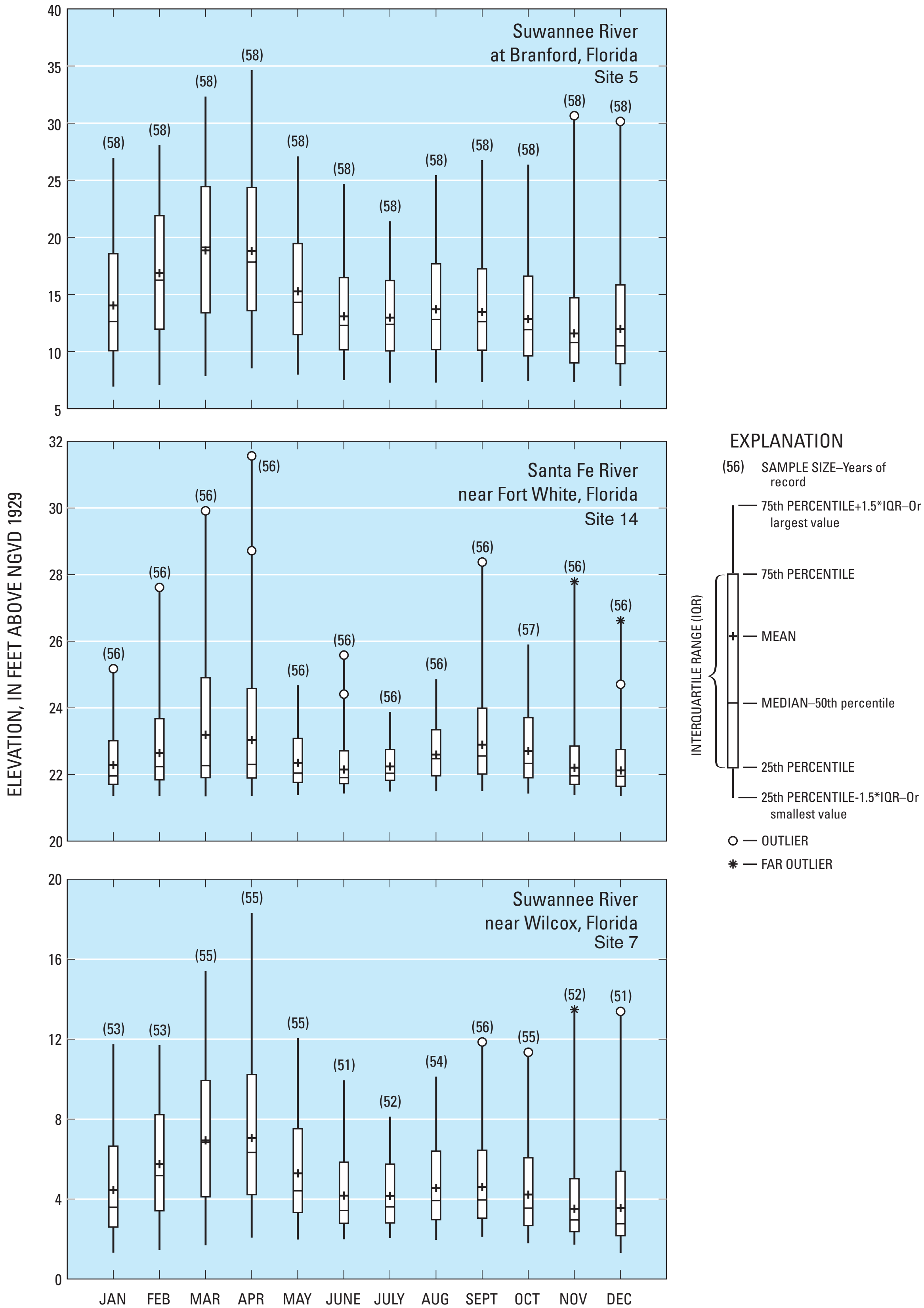


Figure 12. Monthly water-surface elevation at selected locations on the Lower Suwannee and Lower Santa Fe Rivers. Site locations are shown in figs. 4 and 11.

and low-flow conditions in June 1999, the tidal range near the divergence of the Suwannee River near the town of Suwannee at river mile 4.2 was about 3.8 ft, but decreased to 1.8 ft at the gaging station near Old Town (river mile 23.5, map index no. 8, figs. 4 and 11) and to 1.10 ft at the gaging station near Wilcox (river mile 33.5, map index no. 7, figs. 4 and 11). Tidally induced water-level changes occurred at and downstream from the Suwannee River at the mouth of the Gopher River for all streamflows. However, tidally induced water-level changes did not occur upstream at Fowlers Bluff (map index no. 9, figs. 4 and 11) when flow at the Wilcox gaging station exceeded about 23,000 ft^3/s, nor at the gaging station near Old Town when Wilcox flows were greater than about 21,000 ft^3/s.

Smaller Rivers and Streams

Although nearly all of the drainage to the Lower Suwannee and Lower Santa Fe Rivers occurs through the subsurface, smaller rivers and streams drain much of the remainder of the subregional study area that does not contribute runoff to the Lower Suwannee River. Most of this part of the study area has a flat to gently sloping topography and is covered by wet-pine flatwoods. These smaller rivers and streams that drain this area either discharge directly into the Gulf of Mexico or form drainage networks that discharge to the Gulf of Mexico. On the northwestern (Dixie County) side of the Suwannee River, most of the smaller streams are within several miles of the coast. Although the runoff characteristics of these streams are not well understood because of limited streamflow data, these small tidal streams convey tidal currents in and out of marshes as well as some local runoff. Runoff rates from smaller rivers and streams that drain toward the Gulf of Mexico and Lower Santa Fe River were computed from historic streamflow measurements stored in the USGS National Water Information System (U.S. Geological Survey, 1998).

The Waccasassa River (fig. 4) system of streams represents the largest coastal stream network in the subregional study area southeast of the Suwannee River, draining an area of about 617 mi^2. Runoff rates from this system generally range from zero during dry conditions, 4 to 8 in/yr during average conditions, and 0.1 to 1 in/d during peak wet conditions (although some smaller subbasins have higher runoff rates). The California Creek (fig. 4) system of tributaries and distributaries, the largest coastal stream network on the northwestern (Dixie County) side of the Suwannee River, drains a 77-mi^2 area of wetlands south of Cross City. Streamflow data for the California Creek system are limited to measurements taken on September 14, 1964, which coincided with the period of record flood (with an approximate 400-year recurrence interval) on the nearby Steinhatchee River. Runoff rates from the California Creek system were about 0.6 in/d on this same date. During more typical conditions, runoff rates are probably slow or negligible because of the flat topography, high evapotranspiration rates due to the shallow water table, and wetland storage of potential runoff.

A few small rivers and streams drain directly into the Lower Santa Fe River Basin (fig. 1). The largest of these is the spring-fed Ichetucknee River (fig. 4), with a surface drainage area of 213 mi^2 (only 55 mi^2 of which has a defined drainage network). The Ichetucknee River contributes an average flow of about 350 ft^3/s to the Lower Santa Fe River, as estimated from 405 discharge measurements taken at a site about 1.5 mi upstream from the mouth of the Ichetucknee River from 1917 to 1999. Minimum and maximum measured flows were 240 and 580 ft^3/s, respectively, with about 50 percent of the measured flows ranging between 320 and 400 ft^3/s. The Lower Santa Fe River also receives some channelized inflow from several streams draining the northern end of the Waccasassa Flats in Gilchrist County (fig. 3). Cow Creek is the largest of these streams (fig. 4), with runoff rates at about 0.1 in/yr during low-flow (7-day, 10-year) conditions and 5 in/yr during average conditions (data were not available for high-flow conditions).

Ground-Water Hydrology

The ground-water flow system plays a critical role in the overall hydrology of the subregional study area because of the dominance of subsurface drainage and because ground-water discharge sustains the flows of the Lower Suwannee River, Lower Santa Fe River, and numerous springs. To understand the hydrology of the ground-water system, several characteristics must be evaluated including the hydrogeologic framework, hydrologic boundaries, patterns of ground-water levels and flows, hydraulic properties, and the spatial and temporal distribution of sources and sinks of water to the ground-water flow system. Collectively, these characteristics define a conceptual model of the ground-water flow system.

Hydrogeologic Framework

The three principal hydrogeologic units that are present within and adjacent to the study areas are the surficial aquifer system, intermediate aquifer system and intermediate confining unit, and the Floridan aquifer system (fig. 13). The unconsolidated surficial sediments of the surficial aquifer system are present throughout the Northern Highlands area (fig. 3). In the Gulf Coastal Lowlands (fig. 3), the surficial sediments are more locally distributed, but may yield usable quantities of water where the sediments are of sufficient thickness. The surficial aquifer system, where present, is contiguous with the land surface and is composed principally of unconsolidated to poorly indurated siliciclastic deposits (Southeastern Geological Society Ad Hoc Committee on Florida Hydrostratigraphic Unit Definition, 1986). The surficial sediments consist of undifferentiated sands, silts, and clays that are Pliocene and younger age. These sediments are generally less than 40 ft thick, but may be as much as 80 ft thick or greater in areas of the high elevation sand ridges and depressions in the top of

Era	System			Stratigraphic unit	General lithology	Hydrogeologic unit		Notes
Cenezoic	Quarternary	Holocene		Undifferentiated sands and sandy clays	Fine to coarse, clayey quartz sands and sandy clays	Surficial aquifer system		The surficial aquifer system only occurs locally, in areas where these sediments are of sufficient saturated thickness and lateral continuity
		Pleistocene						
	Tertiary	Pliocene		Alachua Formation	Phosphatic clay, occasionally with quartz sand and phosphatic sand			Occurring only locally in the subregional study area
		Miocene		Hawthorn Group	Quartz sand, silt, and clay commonly containing accessory minerals of dolomite, limestone, phosphatic sand, and phosphatic gravel. Interbeds of dolostone and limestone occasionally present	Intermediate aquifer system/ confining unit		Hawthorn Group is generally absent in the subregional study area. It comprises the intermediate aquifer system/ confining unit in the Northern Highlands
				Tampa Limestone	Poorly to well-indurated limestone, with 10- to 35-percent very fine to fine quartz sand	Floridan aquifer system	Upper Floridan aquifer	Occurring only locally in subregional study area (may include part of the Upper Floridan aquifer, where sufficiently permeable)
		Oligocene		Suwannee Limestone	Poorly to well-indurated, crystalline to pelletal limestone			
		Eocene	Upper	Ocala Limestone	Limestone, poorly to well-indurated			
			Middle	Avon Park Formation	Limestone and dolostone*		Middle confining unit	In subregional study area, only present in southern Levy County
			Lower	Oldsmar Formation	Limestone and dolostone; some evaporites and chert		Lower Floridan aquifer	
		Paleocene		Cedar Keys Formation	Dolomite, moderately to highly porous			
					Dolomite with interbedded anhydrite	Lower confining unit of the Floridan aquifer system		Thick anhydrite beds of Cedar Keys form the lower confining unit of the Floridan aquifer system

* Middle part of Avon Park Formation consists of low-permeability gypsum in the southern half of Levy County, where it forms a middle confining unit within the Floridan aquifer system

References: Crane (1986), Florida Geological Survey (2004), Miller (1986)

Figure 13. Geologic and hydrogeologic units present within and adjacent to the regional study area.

the Floridan aquifer system (Hunn and Slack, 1983; Rupert, 1988a; Ron Ceryak, Suwannee River Water Management District, oral commun., 1996). The surficial aquifer system is unconfined and the water table is generally within 10 ft of land surface, but may be deeper in some areas. The water table of the surficial aquifer system is at or near the surface in areas of ground-water discharge (for example, along river and stream corridors) and in the broad wetland areas of Mallory Swamp, San Pedro Bay, and the Waccasassa Flats (fig. 3) where low permeability sediments in the surficial aquifer system (Col and others, 1997) and possibly the Floridan aquifer system impede the vertical flow of ground water. The presence of these sediments in the Waccasassa Flats is consistent with evidence indicating that a closed basin or settling environment existed during the formation of the surficial aquifer system (Col and others, 1997). The saturated thickness of the surficial aquifer system ranges from 10 to 80 ft.

The intermediate aquifer system and intermediate confining unit lie below the surficial aquifer system (where present), and generally consist of fine-grained, unconsolidated deposits of quartz sand, silt, and clay with interbedded limestone of Miocene age (Scott, 1992, p. 55). Regionally, the intermediate aquifer system and intermediate confining unit act as a confining unit that restricts the exchange of water between the overlying surficial aquifer system and the underlying Floridan aquifer system. Accordingly, the term, “intermediate confining unit,” is used in this report to refer to the intermediate aquifer system and intermediate confining unit. The intermediate confining unit is generally present in the Northern Highlands, coinciding with the Hawthorn Group sediments, and is generally absent in the Gulf Coastal Lowlands (figs. 3 and 14). The top of the intermediate confining unit ranges from about 50 to 150 ft above NGVD 1929 (Scott, 1992, p. 44) in elevation, and coincides with the base of the surficial aquifer system. The base of the intermediate confining unit coincides with the top of the Floridan aquifer system and ranges from about 200 ft below to 100 ft above NGVD 1929 (Miller, 1986, pl. 26) in elevation. The thickness of the intermediate confining unit ranges from 0 ft, where the unit pinches out, to greater than 400 ft thick in Baker and Bradford Counties (fig. 14).

The Floridan aquifer system is a thick sequence of carbonate (limestone and dolomite) rocks of mostly Paleocene to early Miocene age. It is subdivided into the Upper Floridan aquifer (Miller, 1986) and in some areas, a middle confining unit and the Lower Floridan aquifer (fig. 13). The Lower Floridan aquifer is present only where a middle confining unit separates the more permeable Upper and Lower Floridan aquifers. The Lower Floridan aquifer is not used for water supply in the study areas, and is only present in the northern part of the SRWMD from Jefferson County east to Columbia County, and in the southern half of Levy County. The Upper Floridan aquifer is present throughout the study area, is highly permeable, and thus typically capable of transmitting large volumes of water. This high permeability mostly is due to the widening of fractures and formation of conduits within the aquifer, caused by dissolution of the limestone from infiltrating water. This process has also produced numerous karst features such as springs, sinking streams, and sinkholes in the study area.

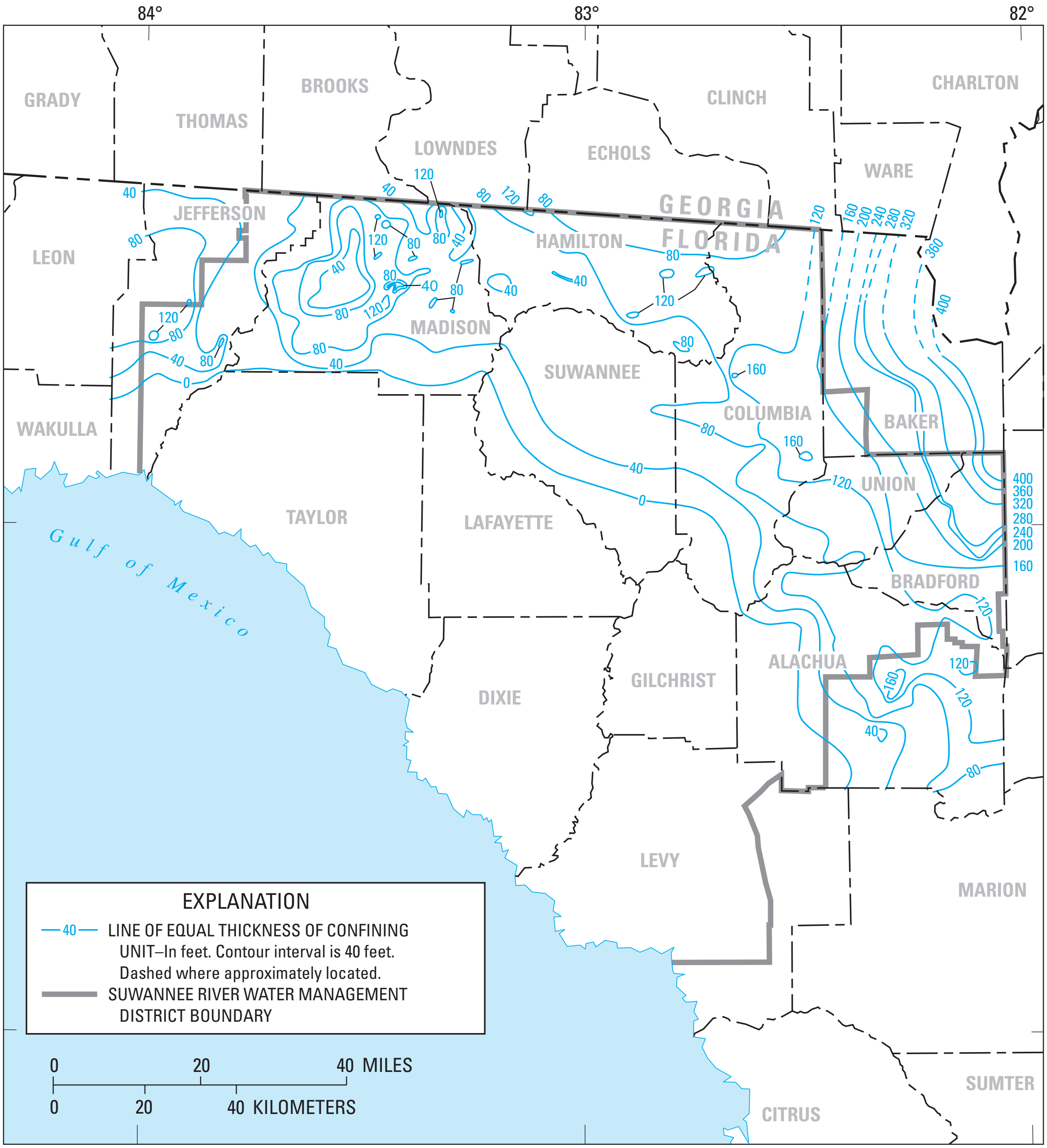
The Upper Floridan aquifer is generally at or near land surface and is unconfined in the Gulf Coastal Lowlands, but is confined or poorly confined in the Northern Highlands. The elevation of the top of the Upper Floridan aquifer (fig. 15) ranges from about 360 ft below to 100 ft above NGVD 1929 (Miller, 1986, pl. 26) in the regional study area, and from NGVD 1929 to 40 ft above NGVD 1929 in the subregional study area. The elevation of the base of the Upper Floridan aquifer in the regional study area (fig. 16) ranges from about 2,100 ft below NGVD 1929 in the coastal areas of Jefferson County to about 400 ft below NGVD 1929 in Lowndes County near Valdosta, Georgia (Miller, 1986, pl. 29). The elevation of the base of the Upper Floridan aquifer (fig. 16) in the subregional study area ranges from about 500 to 1,600 ft below NGVD 1929 (Miller, 1986, pl. 29). The thickness of the Upper Floridan aquifer in the regional study area (fig. 17) ranges from about 400 ft near Valdosta, Georgia, to 2,300 ft in the coastal areas of Jefferson County (Miller, 1986, pl. 28) and from about 500 to 1,600 ft in the subregional study area. Where present, the top of the middle confining unit is the base of the Upper Floridan aquifer, resulting in the apparent discontinuity in the base and thickness of the Upper Floridan aquifer (figs. 16, 17).

Ground-Water Flow System Boundaries

Defining the location and type of boundaries of the ground-water flow system is essential for developing a conceptual model of this system. The subsequent sections describe the location of the horizontal (lateral) and vertical (upper and lower) boundaries of the ground-water flow system in the subregional study area. These sections also describe the direction of ground-water flow along and across these boundaries, including spatial and temporal flow variability at some of the boundaries.

Lateral Boundaries

The lateral boundaries of the ground-water flow system in the subregional study area define horizontal limits of the system. These limits were defined, in part, by the geographic scope of the project and by key features of the regional ground-water flow system adjacent to the subregional study area. These features include areas of relatively high ground-water levels on the potentiometric surface (potentiometric “highs”), areas of relatively low ground-water levels near important areas of ground-water discharge (such as major rivers or the Gulf of Mexico), and points and lines (divergence points” and divides,” respectively) where ground-water flow paths diverge. These features collectively define the location of the following key lateral boundaries: the northeastern lateral no-flow boundary, the southeastern lateral no-flow boundary, the western lateral no-flow boundary, and the constant-head



Data Source: Florida Geological Survey, Tallahassee; and Suwannee River Water Management District, Live Oak, written communs., 1997

Figure 14. Extent and thickness of the Hawthorn Group sediments (intermediate confining unit) in the Suwannee River Water Management District.

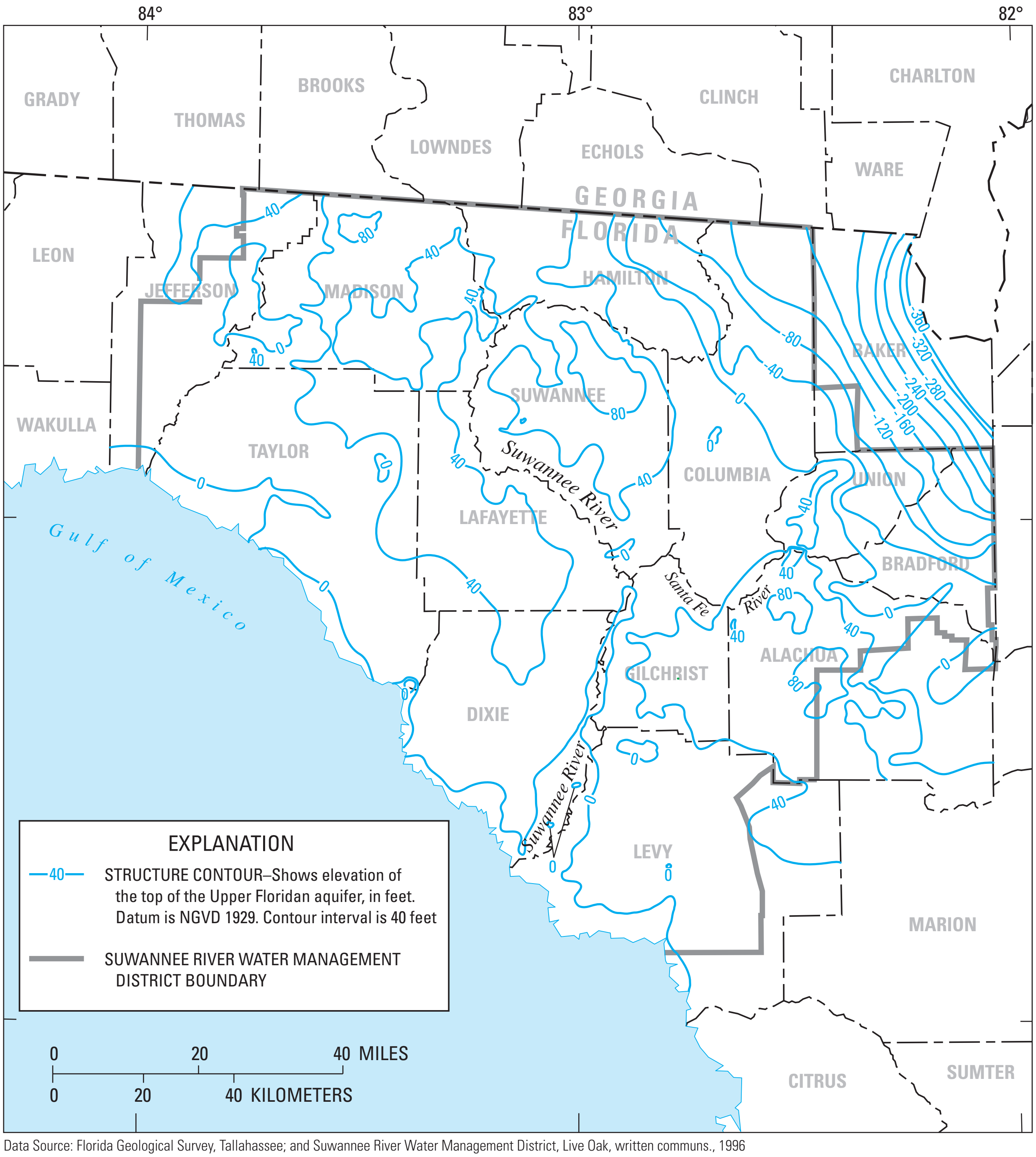
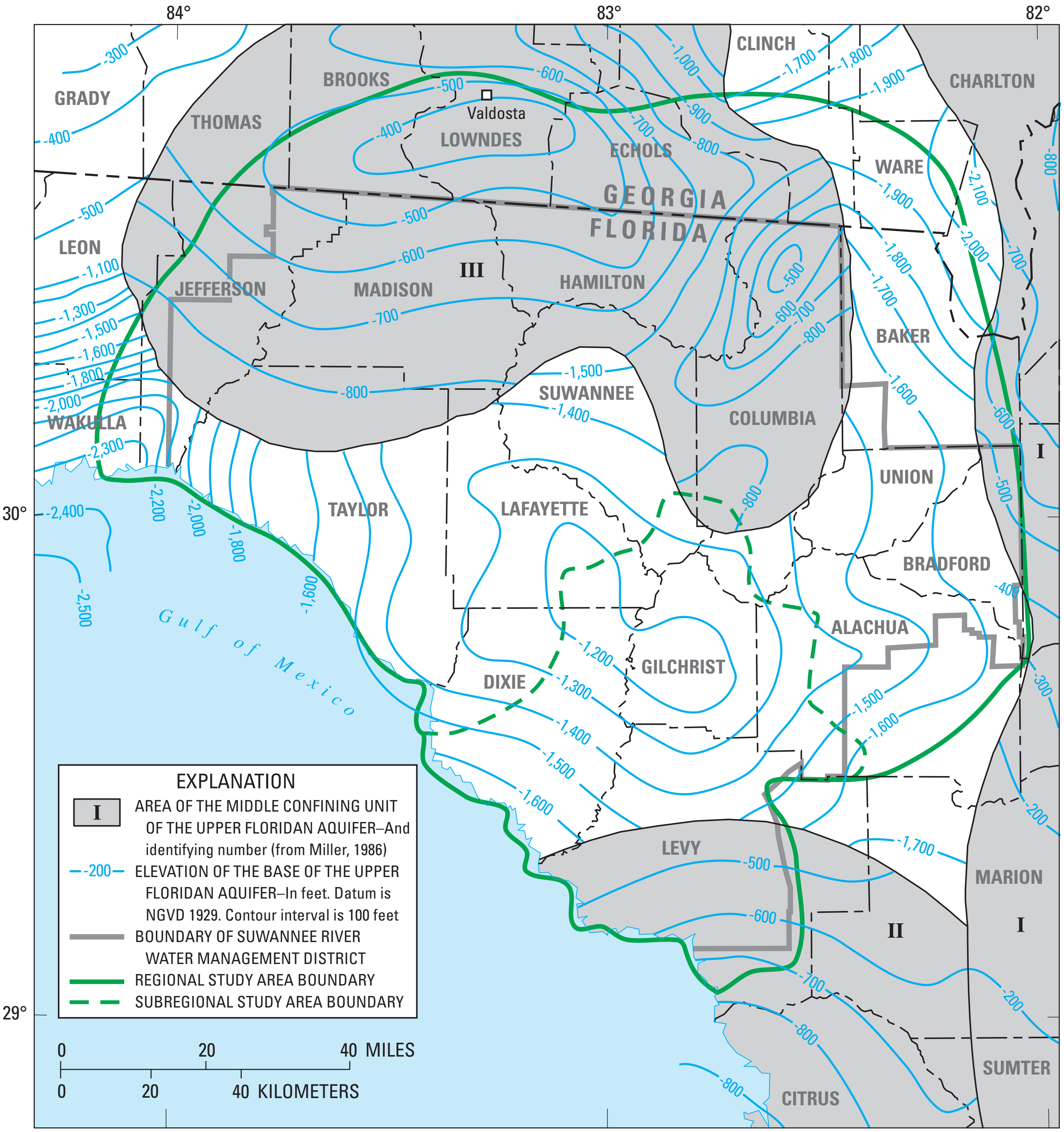
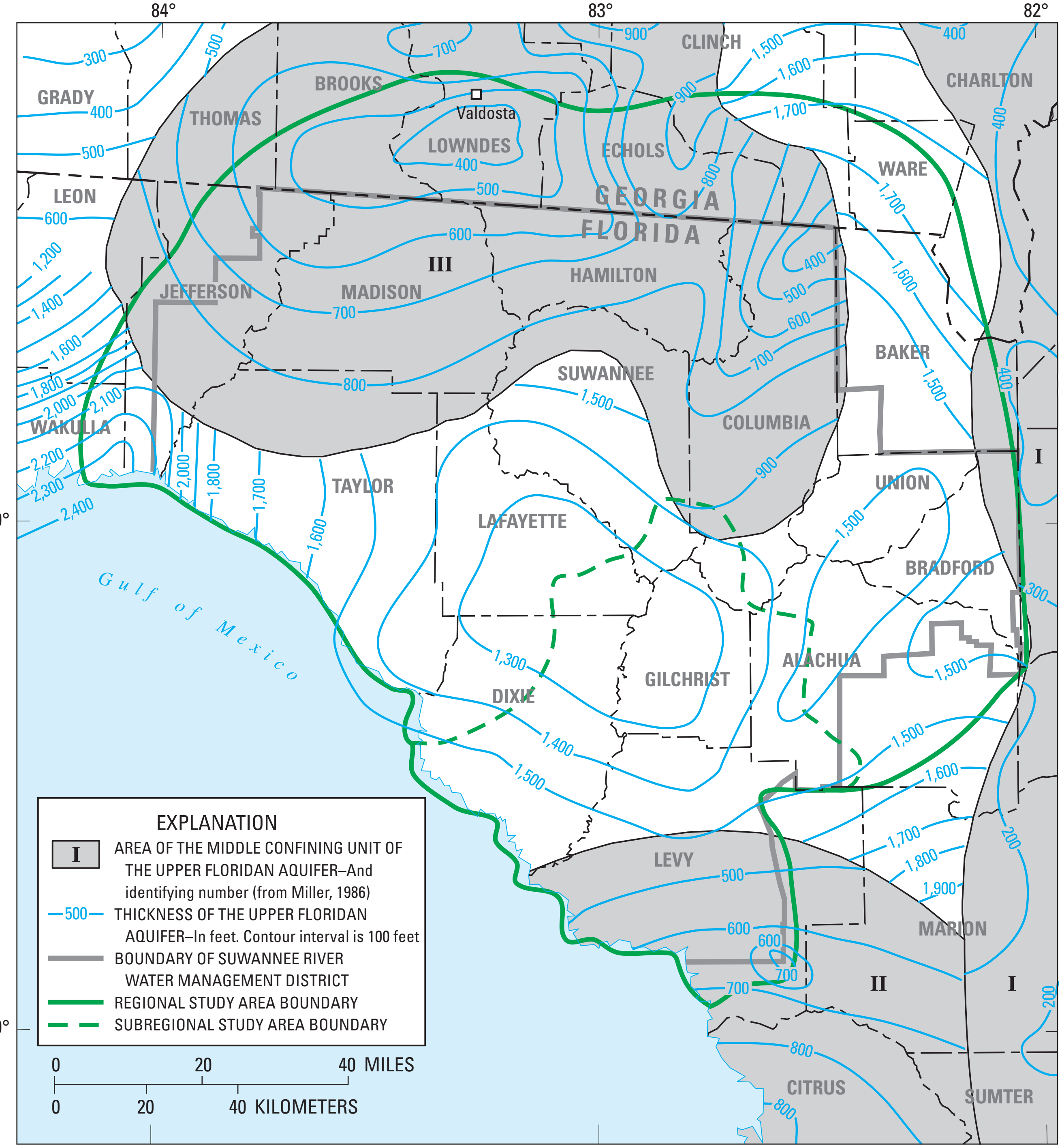


Figure 15. Elevation of the top of the Upper Floridan aquifer.



Source: Miller, 1986.

Figure 16. Elevation of the base of the Upper Floridan aquifer.



Source: Miller, 1986.

Figure 17. Thickness of the Upper Floridan aquifer.

boundary represented by the Gulf of Mexico coastline. These four boundaries define the lateral limits of the area that contribute water to the Lower Suwannee and Lower Santa Fe Rivers, and therefore, represent an important constraint on the amount of water that is available to sustain streamflows and ground-water withdrawals.

The northeastern lateral flow boundary is largely defined by two prominent dome-shaped potentiometric highs that are centered near Valdosta, Georgia (fig. 18), and in the Keystone Heights area in Florida along the border between Clay, Putnam, Alachua, and Bradford Counties (fig. 18). Ground-water flow paths (flow lines) can be drawn that originate at the highest points of the Valdosta and Keystone Heights potentiometric highs and flow toward one or more points of divergence along the boundary. These flow lines change direction abruptly at the points of divergence, either flowing southward toward the Suwannee, Ichetucknee, and Lower Santa Fe Rivers or northward toward the Atlantic Ocean coastline and regional cones of depression centered near Jacksonville and Fernandina Beach. Thus, the flow path between the potentiometric highs and divergence points between Valdosta and Keystone Heights defines the northeastern limit of the area that contributes ground-water flow to the subregional study area.

Historic ground-water level data and the estimated predevelopment potentiometric surface of the Upper Floridan aquifer indicate that the location of the northeastern

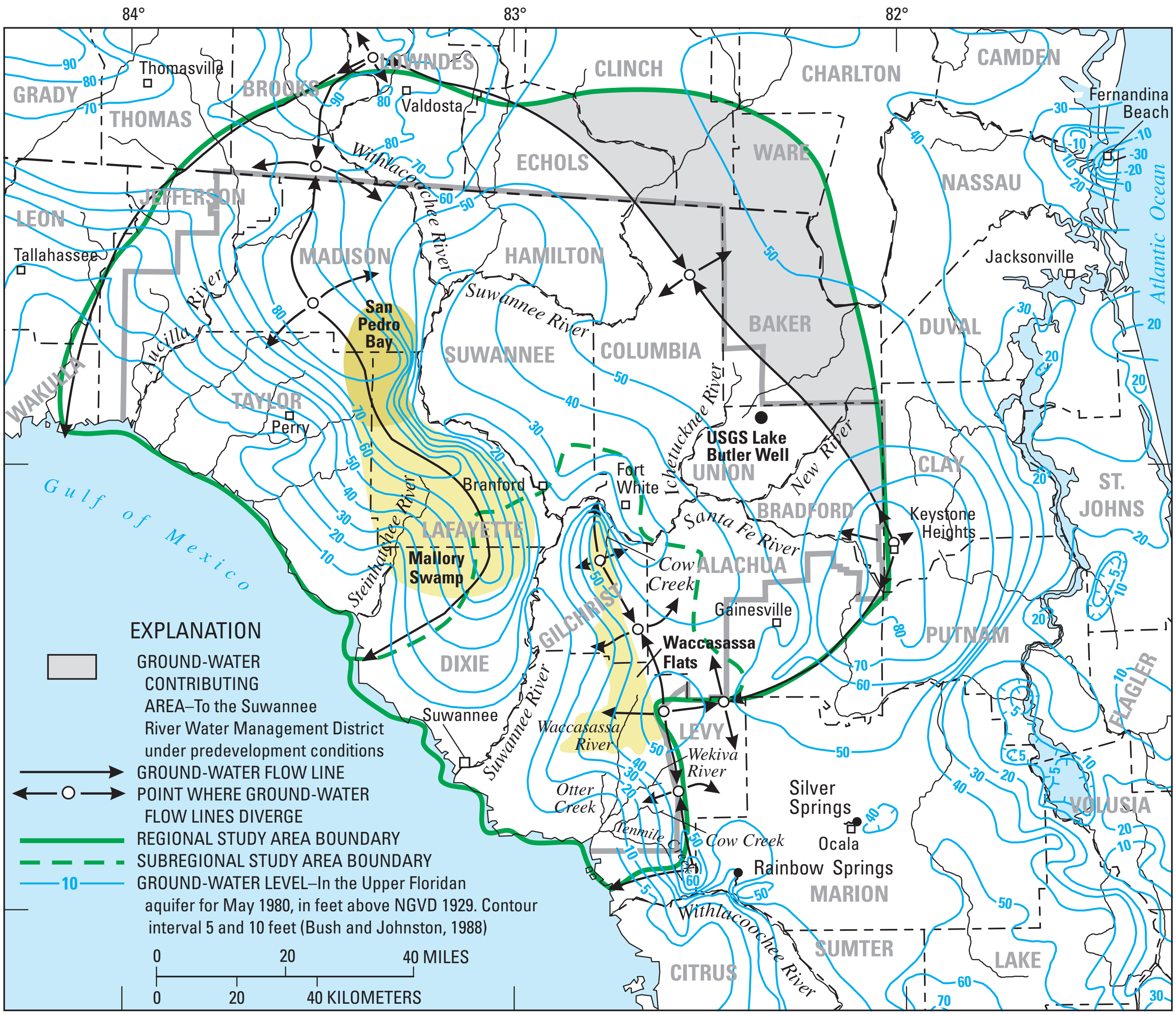


Figure 18. Potentiometric surface of the Upper Floridan aquifer during May 1980, including key features of the ground-water flow system. USGS is U.S. Geological Survey.

lateral flow boundary has migrated to the southeast over the last century (figs. 18 and 19). Before substantial withdrawals occurred from the Upper Floridan aquifer, the SRWMD was almost completely enclosed by the bounding flow lines that originate at the highest points of the Valdosta and Keystone Heights highs (fig. 19). Large ground-water withdrawals from the Upper Floridan aquifer began in the Jacksonville and Fernandina Beach areas of Florida during the late 1800s and have increased in these and other areas northeast of the regional study area. These withdrawals have caused large regional drawdowns in the Upper Floridan aquifer in the northeastern part of the regional study area and the concurrent westward migration of the northeastern part of the flow line boundary. This has resulted in the reduction of the area of the Upper Floridan aquifer that contributes water to the SRWMD and the diversion of water from the SRWMD to the pumping centers near Jacksonville and Fernandina Beach. Historic water-level data from long-term observation wells (fig. 20) along the northeastern lateral flow boundary indicate that: (1) the rate of drawdown has slowed, and perhaps stopped, in recent times; and (2) movement of this boundary has slowed or the boundary may have reached a new equilibrium location (fig. 18).

The southeastern lateral flow boundary is defined by five ground-water flow paths (fig. 18). The first of these flow paths originates on the potentiometric high near Keystone Heights, and is directed downgradient in a southwesterly direction toward eastern Levy County where it intersects a second flow line originating from the potentiometric high along the Waccasassa Flats to the west (described later). This part of the boundary delimits the contributing areas to the Lower Santa Fe River to the north and Rainbow Springs and Silver Springs to the south. The southeastern lateral flow boundary continues along two ground-water flow paths: (1) along the southern end of the ridge of the Waccasassa Flats potentiometric high; and (2) along a path directed northward from the potentiometric high west of Rainbow Springs. The location of this part of the southeastern lateral flow boundary determines whether recharge to the Upper Floridan aquifer flows east toward Rainbow Springs or west toward the Gulf of Mexico. The rest of the southeastern lateral flow boundary is defined by a short ground-water flow line directed westward toward the Gulf of Mexico from the potentiometric high west of Rainbow Springs.

The western lateral no-flow boundary defines the boundary between areas that contribute ground-water flow to the Withlacoochee River and Suwannee River to the east, and those that contribute ground-water flow to the Gulf of Mexico (and streams and rivers draining into the Gulf) to the west. This lateral boundary is defined by three ground-water flow paths. The northern part of the western boundary is defined by a southward-directed flow path originating on the Valdosta potentiometric high and terminating at a point where it intersects a northward-directed flow path originating on another potentiometric high centered in southern Madison County. The latter potentiometric high (the "San Pedro Bay potentiometric high") coincides with an elevated area that has generally flat topography covered with poorly drained flatwoods and swamps, most notably San Pedro Bay and Mallory Swamp (fig. 18). The western lateral no-flow boundary continues along a flow path that originates on the San Pedro Bay potentiometric high and is directed southward toward the Gulf of Mexico.

The lateral boundary near the Gulf of Mexico coastline defines a final key external boundary for the subregional study area, and its location is defined by the line of intersection of two surfaces: (1) the interface between freshwater and saltwater in the Upper Floridan aquifer using the conceptualization of Cooper (1959, p. 464); and (2) the top of the Upper Floridan. In the study areas, this line of intersection represents the westward limit of freshwater discharge from the Upper Floridan aquifer. Only a few direct observations of saline ground water are available in the study areas and are insufficient to map the elevation of the freshwater-saltwater interface. Countryman and Stewart (1997) studied the subsurface electrical resistivity and conductivity of the coastal areas of Dixie and Levy Counties to locate the freshwater-saltwater interface near the mouth of the Suwannee River. Interpretations of harmonic electromagnetic, direct-current resistivity, and transient electromagnetic data placed the interface on the landward side and within several miles of the coastline; however, these results could not be confirmed due to insufficient water-quality and lithologic data (Countryman and Stewart, 1997, p. ix). The data are probably best used to interpret the configuration of the surface of the freshwater-saltwater interface, rather than its precise elevation near the Suwannee River (Tony Countryman, Northwest Florida Water Management District, oral comm., 1997).

Despite the absence of direct observations of the location of the freshwater-saltwater interface, several types of data can be used to infer the location of the western boundary. First, the elevation of the interface under static freshwater and saltwater conditions can be approximated using ground-water level data and the Ghyben-Herzberg principle (Fetter, 1988, p. 151), which place the interface at an elevation below NGVD 1929 that is equal to 40 times the ground-water level above NGVD 1929. Ground-water level data from several wells less than 1 mi from the Gulf coast indicate that average ground-water levels near the coast are typically 1 to 2.5 ft higher than NGVD 1929, which indicates that the freshwater-saltwater interface is on the order of 40 to 100 ft below NGVD 1929 near the coast. Second, recent data from two shallow well nests in the coastal marshes, several miles north and south of the mouth of the West and East Passes of the Suwannee River, indicate a consistent pattern of upward-seeping, brackish ground water (Ellen A. Raabe and others, U.S. Geological Survey, written commun., 2004). Thus, both well nests appear to be located in a zone of mixing that defines a diffuse boundary between fresh and salty ground water discharging on the landward and seaward sides of the boundary, respectively. Finally, the unconfined and highly permeable nature of the Upper Floridan aquifer in the subregional study area limits the seaward extent of the boundary because ground water can easily discharge through the top of the Upper Floridan aquifer.

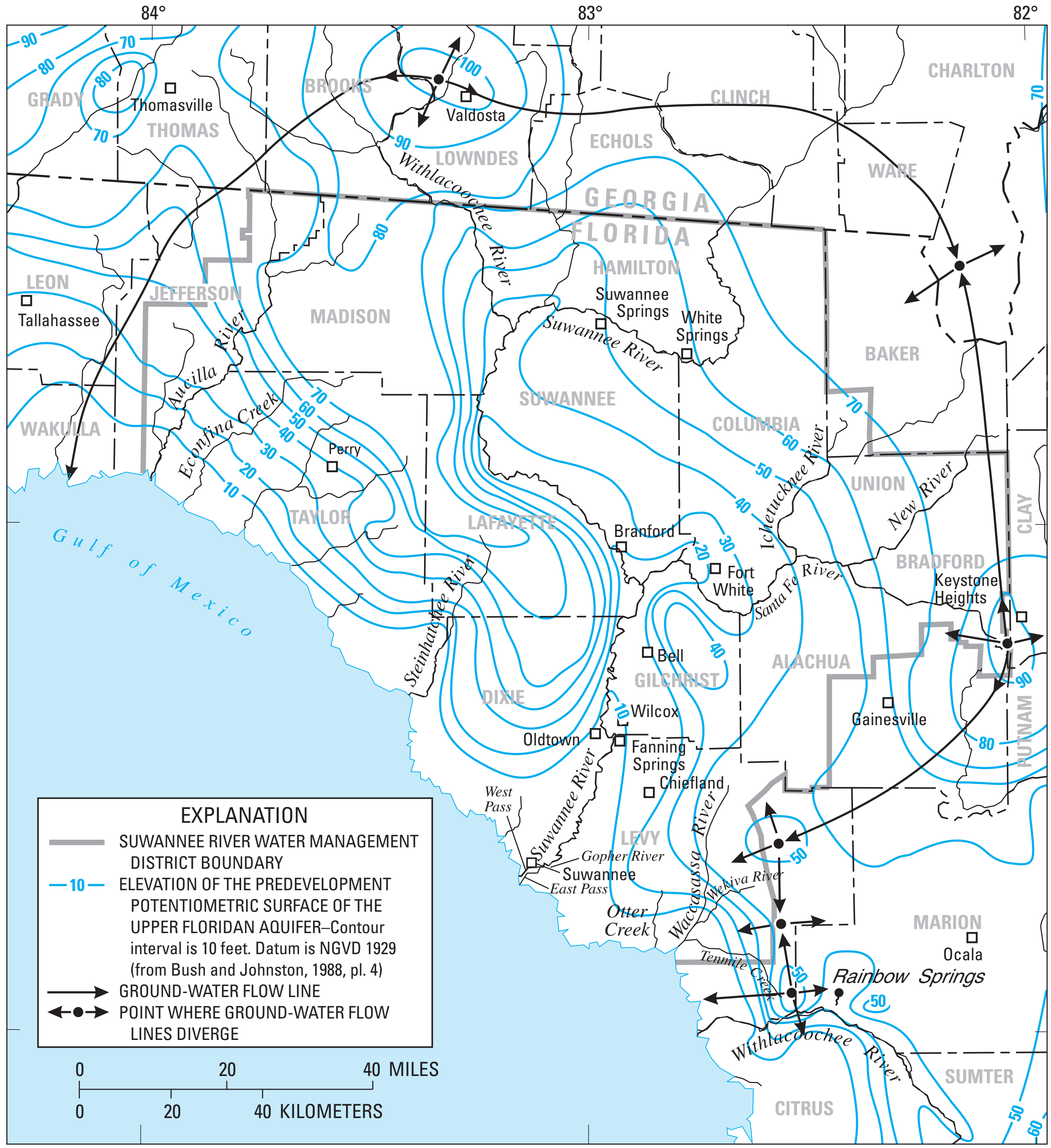


Figure 19. Predevelopment potentiometric surface of the Upper Floridan aquifer, including key features of the ground-water flow system.

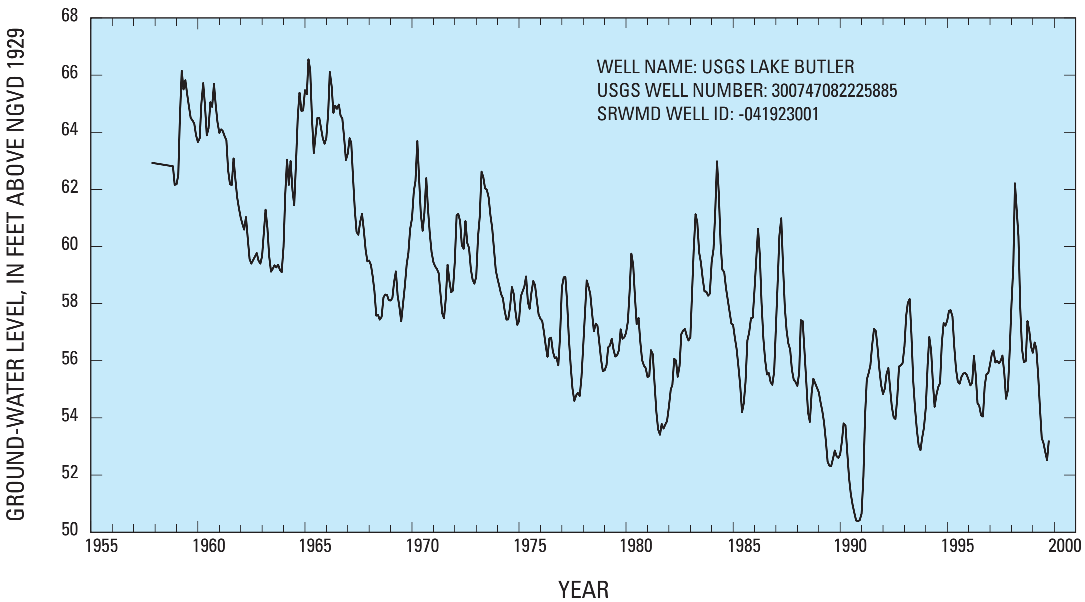


Figure 20. Ground-water levels in the Upper Floridan aquifer near Lake Butler, Florida, from 1957 to 1999. USGS is U.S. Geological Survey, and SRWMD is Suwannee River Water Management District. Well location is shown in fig. 18.

Collectively, the above observations indicate that the freshwater-saltwater interface is probably close to the coastline. Thus, the western boundary is conceptualized as being a constant-head boundary with a head value equal to zero, which is the approximate elevation of the Gulf of Mexico in the subregional study area located within 1 mi offshore of the coastline.

Upper and Lower Flow-System Boundaries

For this study, the top of the ground-water flow system in the subregional area was conceptualized to be the top of the Upper Floridan aquifer or the top of the unconsolidated surficial sediments (where present). Therefore, the Upper Floridan aquifer and any overlying sediments are treated as a single flow system. This conceptualization is supported by data from collocated wells (not shown) near Bell, Florida (fig. 4), tapping both the Upper Floridan aquifer and surficial deposits, which indicate that water levels in these units are similar in magnitude and in their patterns of fluctuations (fig. 21).

The lower boundary of the ground-water flow system is coincident with the base of the Upper Floridan aquifer. This boundary is defined by one of three features, including the: (1) top of a middle confining unit within the aquifer where present; (2) base of the aquifer in areas where a middle confining unit is absent and where freshwater is present throughout the entire thickness of the aquifer; or (3) freshwater-saltwater interface when it occurs within the aquifer. As previously described, the elevation of the freshwater-saltwater interface was estimated using the Ghyben-Herzberg principle, and the resulting surface of the base of the flow system is shown in figure 22.

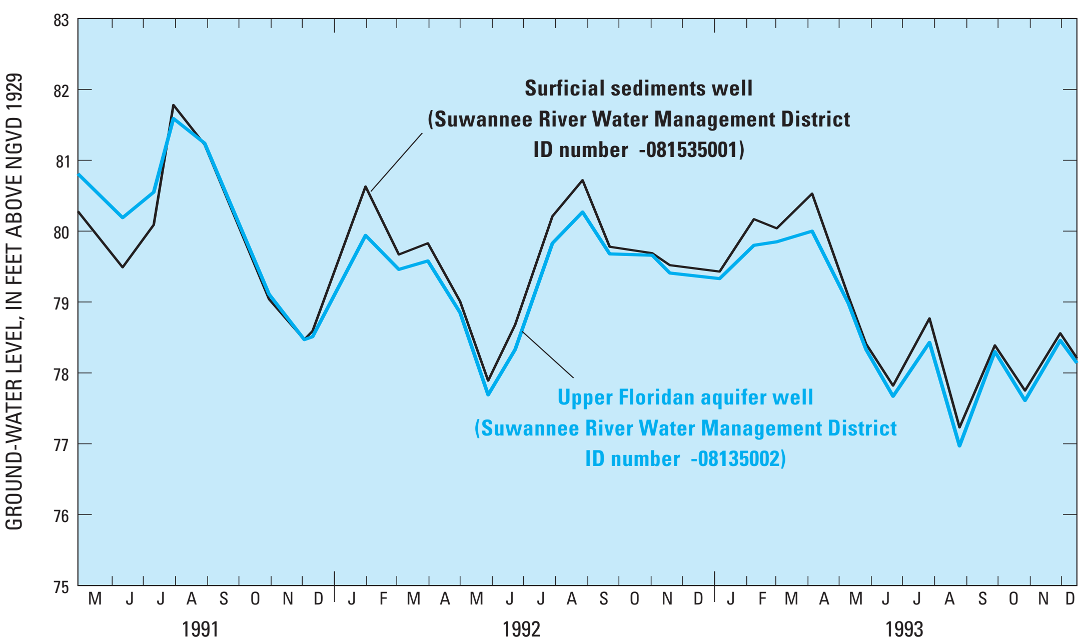


Figure 21. Ground-water levels of the collocated wells tapping the unconsolidated surficial sediments and Upper Floridan aquifer near Bell, Florida, 1991-93. The location of Bell is shown in fig. 4.

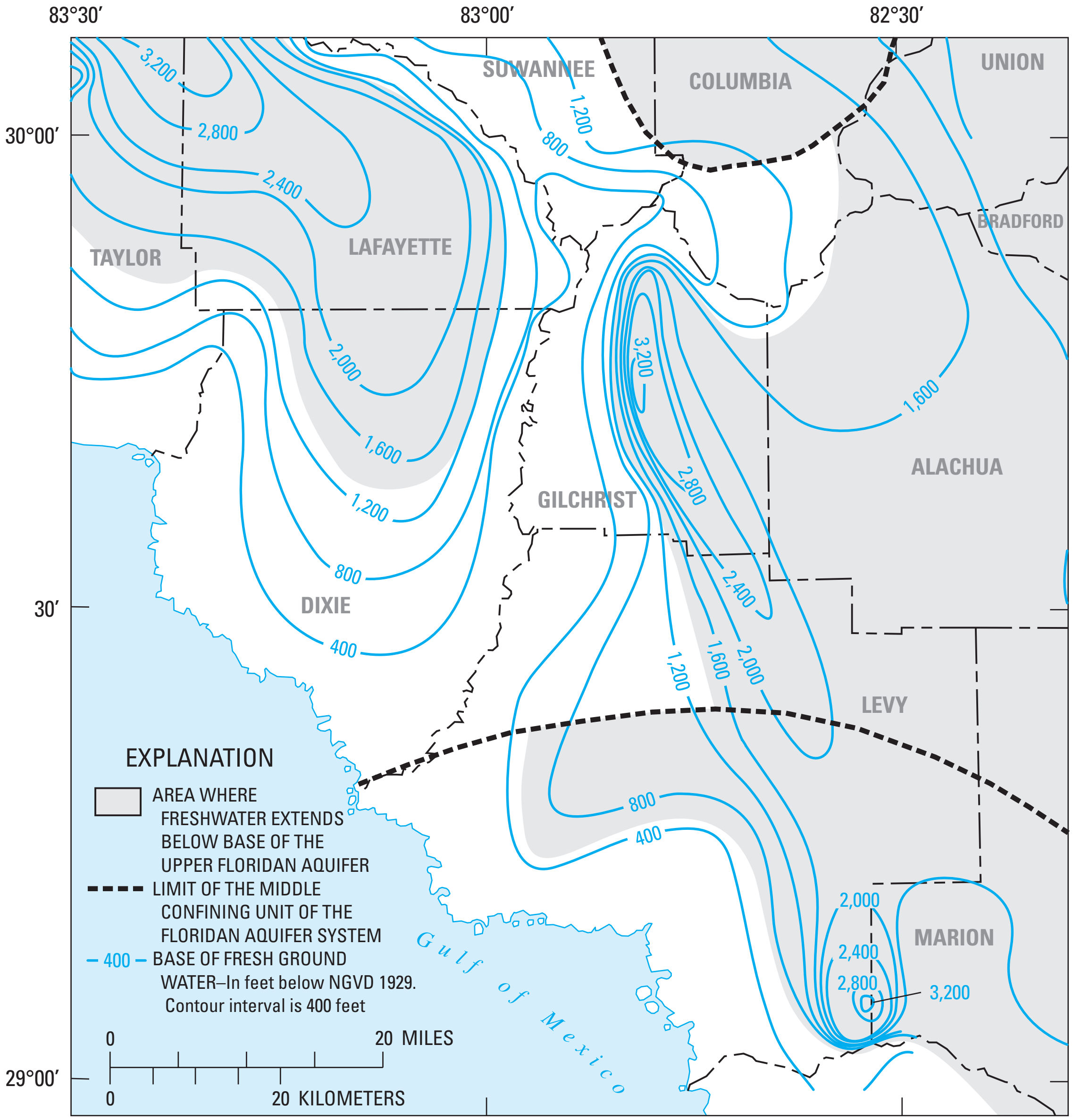


Figure 22. Elevation of the base of freshwater in the Upper Floridan aquifer.

Patterns of Ground-Water Levels and Flow

The ground-water flow system within and adjacent to the subregional study area can be characterized by a number of spatial and temporal patterns. Several of the spatial patterns were discussed in the previous description of the ground-water flow system boundaries. The general spatial pattern of ground-water flow is one in which flow is directed away from western and eastern external boundaries that pass through the San Pedro Bay-Mallory Swamp, Keystone Heights, Waccasassa Flats, and Rainbow Springs potentiometric highs toward three sets of features for ground-water discharge. These three sets of features, in order of decreasing size of contributing area, are the: (1) Lower Santa Fe River-Ichetucknee River-Lower Suwannee River drainage network; (2) Otter Creek-Waccasassa River-Tenmile Creek drainage network; and (3) Gulf of Mexico coastal area. Ground-water flow lines that delineate the approximate location of the boundaries between the contributing areas to these features during March 1998 are shown in figure 23. Analysis of potentiometric surface maps from different periods of time and under a range of hydrologic conditions indicate that the overall configuration of the potentiometric surface of the Upper Floridan aquifer has remained reasonably consistent over time (except where noted previously). Accordingly, the spatial patterns of flow and contributing areas described above and in the previous descriptions of key external boundary conditions have also remained reasonably consistent.

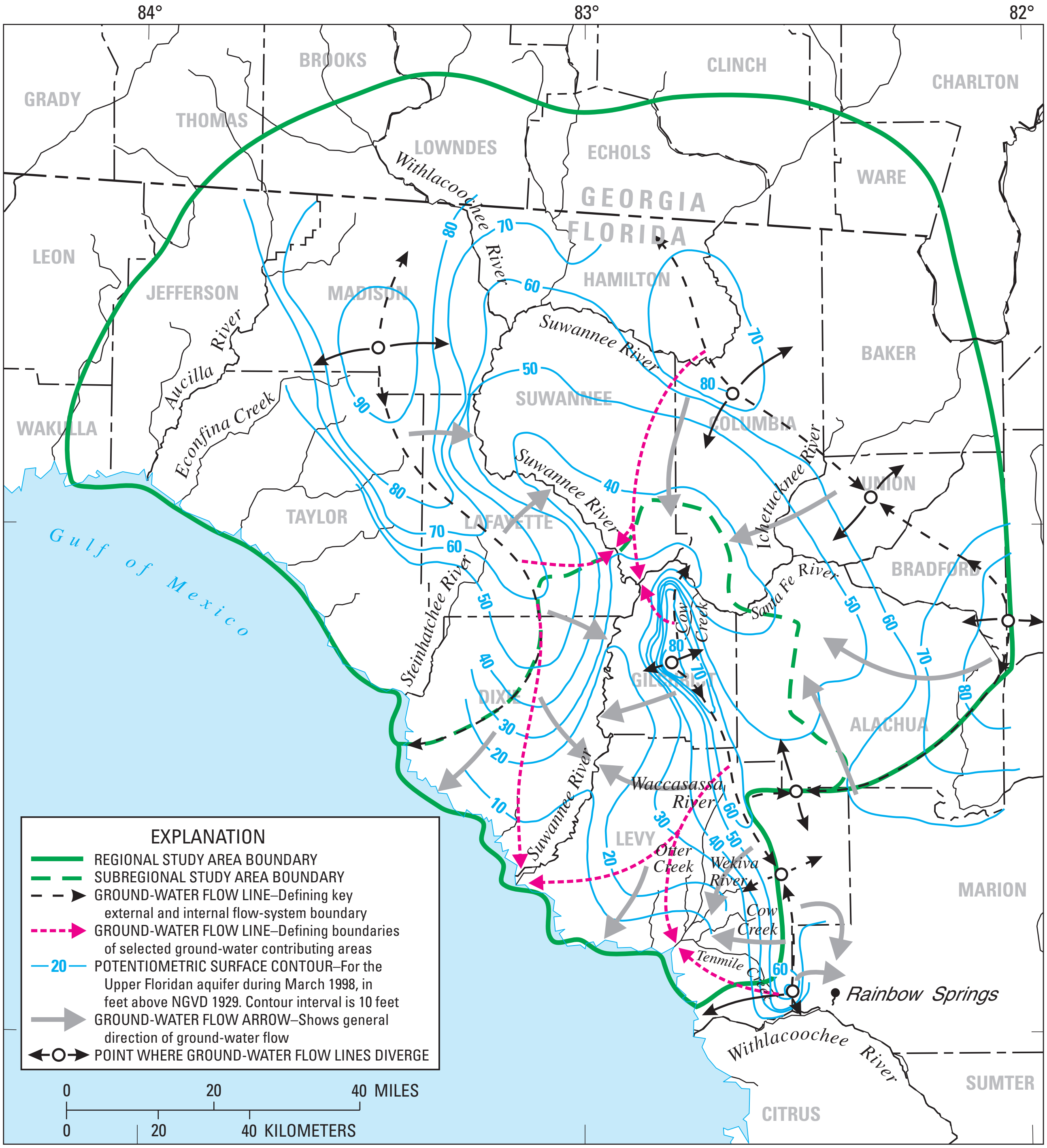


Figure 23. Potentiometric surface of the Upper Floridan aquifer during March 1998 and general spatial patterns of ground-water flow.

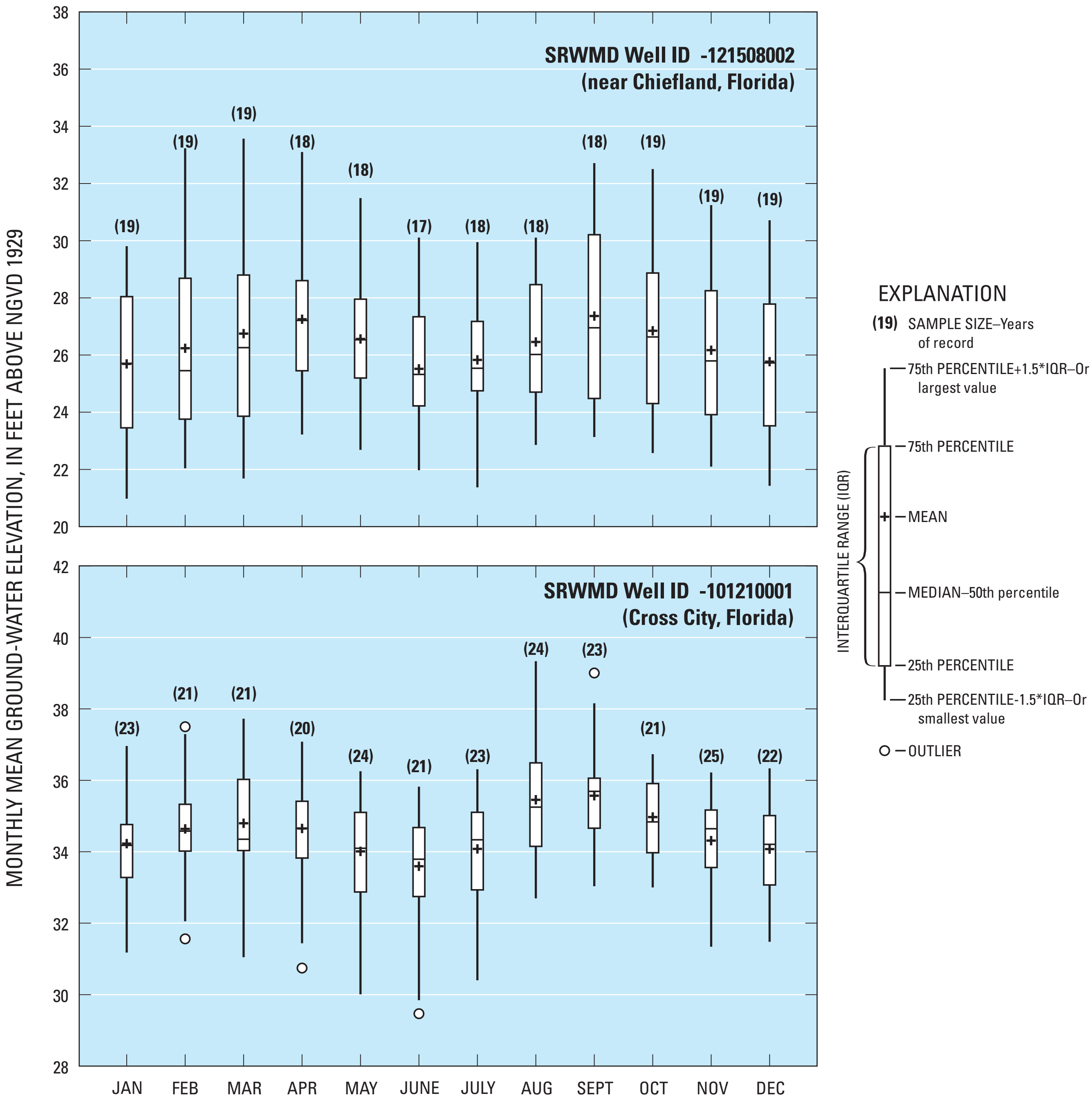


Figure 24. Monthly mean ground-water levels near Chiefland and Cross City, Florida. Locations of Chiefland and Cross City are shown in fig. 4. SRWMD is Suwannee River Water Management District.

Several temporal patterns are also evident in the ground-water system in the subregional study area. High ground-water levels typically occur during two periods: from late summer to mid-fall, and from early to mid-spring (fig. 24). Low ground-water levels typically occur in the intervening periods. The pattern of high water levels is similar to the pattern of monthly rainfall data (fig. 25), although the monthly mean ground-water levels generally lag the monthly mean rainfall totals by about 1 month during the summer.

During the study period, ground-water levels exhibited varying degrees of consistency with the previously discussed seasonal pattern. In water year 1998, ground-water levels exhibited a relatively typical seasonal pattern: peak ground-water levels occurred during March-April and October in most of the observation wells. The October 1998 period represented a particularly sharp peak in ground-water levels, coinciding with a tropical storm that produced heavy rainfall in the subregional study area, with several weather stations

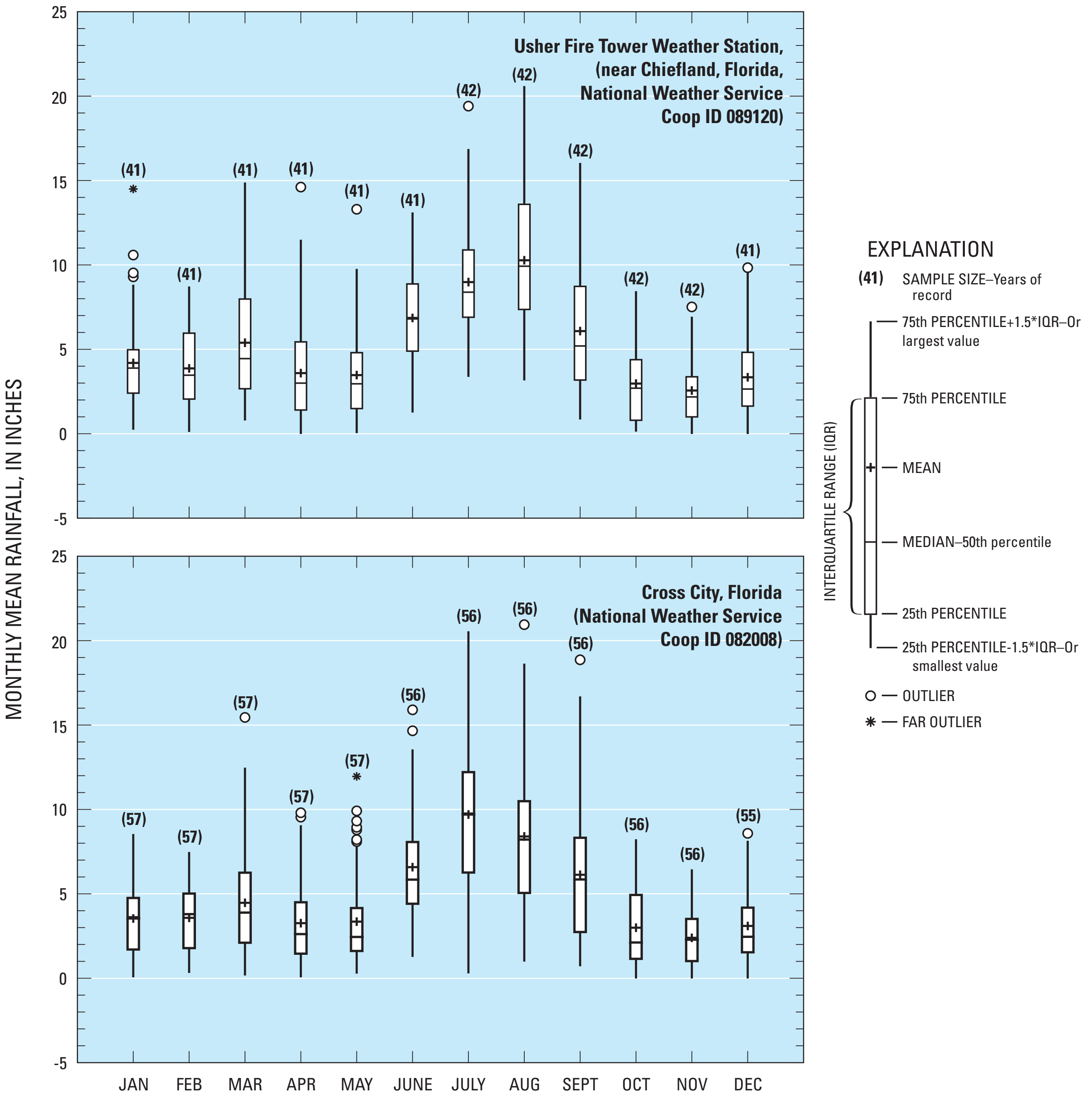


Figure 25. Monthly rainfall near Chiefland and Cross City, Florida. Locations of Chiefland and Cross City are shown in fig. 4.

reporting rainfall totals that were two to three times (6-11 in.) higher than normal for the month. In water year 1999, ground-water levels were less consistent with typical seasonal weather patterns. The typical late winter-early spring peak was absent or minimal in all wells. Many of the wells showed little response to the rains that eventually came in August and September 1999, and ground-water levels were better characterized as being in a relatively steady recession since the October 1998 peak.

Because of the close linkage between the river and aquifer, ground-water levels along the river are sensitive to fluctuations in river levels (fig. 26). Data from wells in these areas confirm this association and also indicate the direction of flow between the Lower Suwannee River and Upper Floridan aquifer. For example, in the vicinity of the gaging station of the Suwannee River near Bell near river mile 56, ground-water levels in nearby well t059 (figs. 26, 27; table 2) indicate that the direction of ground-water flow probably

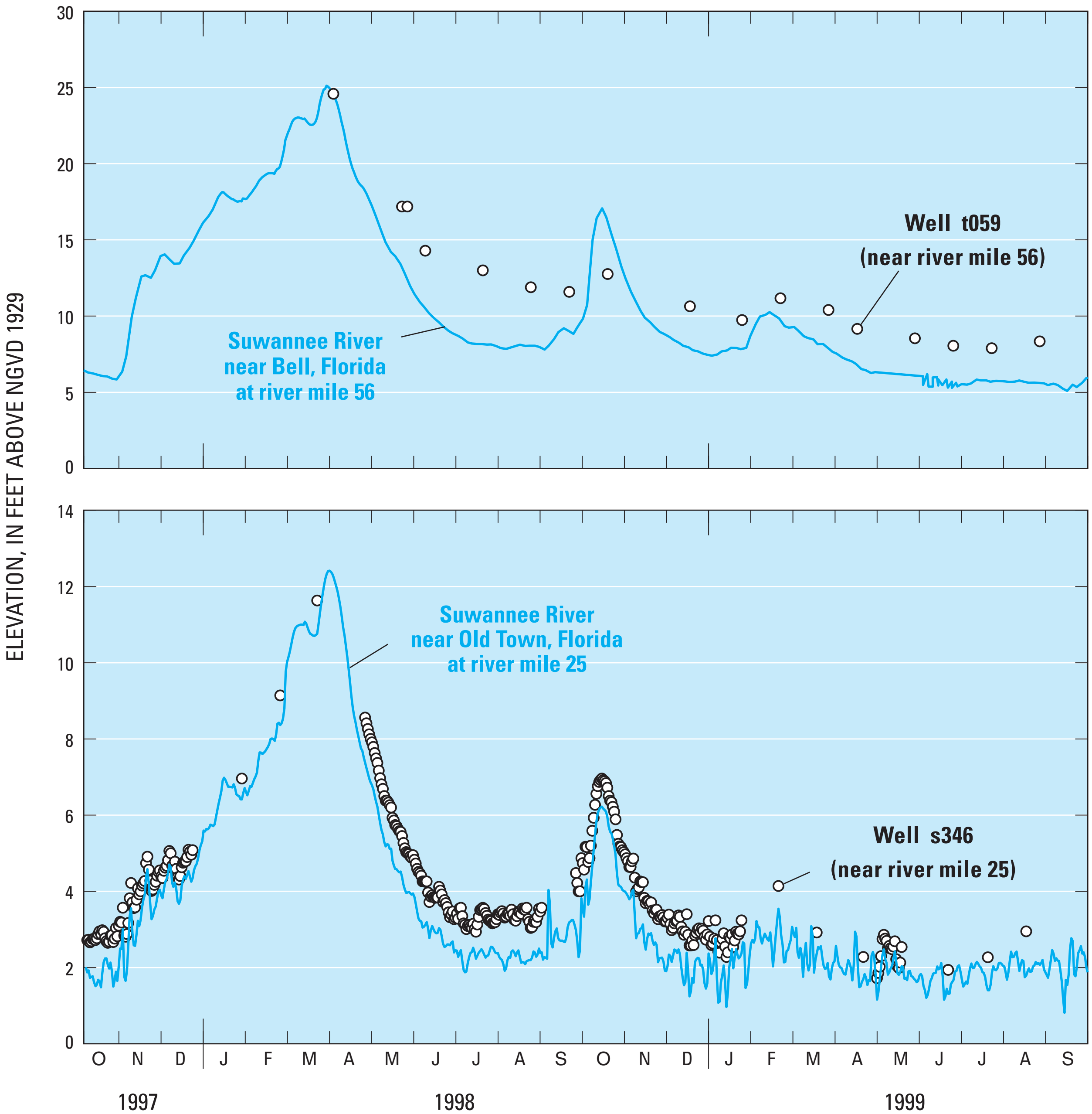


Figure 26. Temporal variations in ground- and surface-water levels near river miles 25 and 56 of the Suwannee River. Well locations are shown in fig. 27.

reversed during periods of peak river levels when well water levels were at or below river levels. Conversely, downstream near river mile 25 and the gaging station (map index no. 9, fig. 4) near Old Town and Manatee Springs, ground-water levels in well s346 (figs. 26, 27) near the river were consistently higher than river levels, indicating a consistent pattern of ground-water discharge to the river. In the lower reaches of the river, tidally induced changes in river levels produce tidally induced changes in aquifer levels. Because of the small lag in the response of the ground-water levels to the tidal fluctuations in the river levels, the difference between river and aquifer levels also fluctuates. As a result, this causes ground-water discharge to the lower river to fluctuate over the course of a tidal cycle. Observations of these short-term fluctuations in ground-water discharge were made during a series of continuous spring-flow measurements taken in Fanning Springs and Manatee Springs, and are briefly described later.

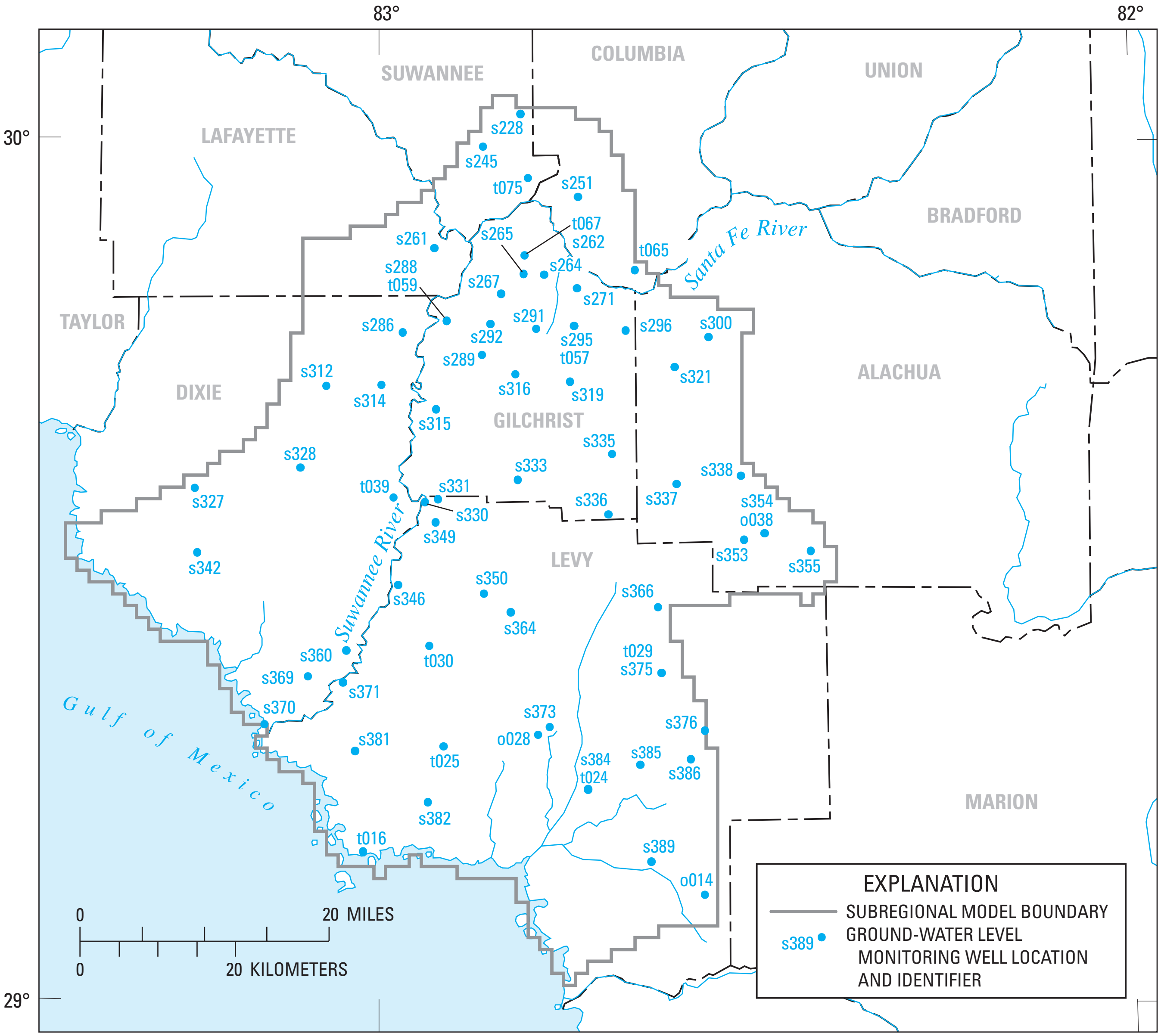


Figure 27. Network of wells used to measure ground-water levels within the subregional study area.

Hydraulic Properties

Two types of hydraulic properties influence ground-water flow in aquifers: conductive and storage properties. Conductive properties influence the rate of ground-water flow for a given hydraulic head gradient and are a function of the degree of intergranular connections and secondary dissolution of the subsurface rocks and sediments. Conductive properties are typically expressed as hydraulic conductivity or transmissivity (conductivity multiplied by aquifer thickness). Storage properties influence how much ground water is released from storage for a given change in hydraulic head and are a function of the compressibility of water and the elasticity and water-retentive characteristics of the subsurface rocks and sediments. Storage properties are typically expressed as: (1) specific yield (volume of water released from desaturation of the pore space of a unit volume of an unconfined aquifer by gravity), (2) specific storage (volume of water released per unit volume of aquifer per unit change in hydraulic head), or (3) storativity (specific storage multiplied by aquifer thickness). Estimates of hydraulic properties can be made from aquifer tests, calibration of numerical (simulation) models, and by making inferences from other hydrologic and hydrogeologic data.

Table 2. Identifiers for the network of wells used to measure ground-water levels within the subregional study area.

[Well locations shown in fig. 27. Well numbers beginning with the letters ‘s’, ‘o’, or ‘t’ indicate that the well was measured by personnel from the Suwannee River Water Management District (SRWMD), U.S. Geological Survey (USGS) Orlando office, and USGS Tallahassee office, respectively. Also note that some of the wells were measured by personnel from more than one office. Therefore, a given well may have more than one well number (for example, USGS well identifier 294743082543901 has two well numbers: s288 and t59). NA indicates that the well does not have the corresponding well number]

Well No. (this report only)	SRWMD well identifier	USGS well identifier	Well No. (this report only)	SRWMD well identifier	USGS well identifier
s228	-051521001	300205082484401	s350	-111436001	292843082514201
s245	-061401003	295950082514685	s353	-111809001	NA
s251	-061629001	295618082440985	s354	-111811001	293252082292385
s261	-071417001	295246082553885	s355	-111920001	NA
s262	-071515001	295214082482501	s360	-121330002	NA
s264	-071526001	295055082465201	s364	-121508002	292713082493685
s265	-071528001	295057082483485	s366	-121708005	NA
s267	-071532001	294931082501685	s369	-131203001	NA
s271	-071632001	NA	s370	-131219001	291940083090101
s286	-081313005	294654082581085	s371	-131306001	NA
s288	-081416001	294743082543901	s373	-131526001	NA
s289	-081425001	294521082514901	s375	-131705001	NA
s291	-081515002	294709082473001	s376	-131736001	291910082341185
s292	-081518005	NA	s378	-131821001	NA
s295	-081618001	294721082443001	s381	-141305001	NA
s296	-081624001	294701082402201	s382	-141429001	291414082560985
s300	-081724001	NA	s384	-141620001	291508082432901
s312	-091212003	294311083041085	s385	-141707002	NA
s314	-091311001	NA	s386	-141711001	NA
s315	-091420001	294135082553485	s389	-151624001	NA
s316	-091504001	294400082491385	o014	NA	290743082341501
s319	-091607001	294330082445085	t016	NA	291048083011801
s321	-091704001	294428082362901	t022	NA	291241082300101
s327	-101120001	NA	t024	NA	291508082432901
s328	-101210001	293731083061885	t025	NA	291806082545601
s330	-101428001	NA	o028	NA	291855082472601
s331	-101429011	NA	t029	NA	292310082373701
s333	-101516017	293822082483285	t030	NA	292507082560201
s335	-101603001	NA	o038	NA	293252082292385
s336	-101634001	293414082415285	t039	NA	293525082585301
s337	-101722001	293619082362385	t057	NA	294721082443001
s338	-101816001	NA	t059	NA	294743082543901
s342	-111117007	293137083143085	t065	NA	295114082393801
s346	-111326004	292921082583285	t067	NA	295214082482501
s349	-111405001	NA	t075	NA	295737082480801

Aquifer Tests

Aquifer tests have been used to estimate hydraulic properties of the Upper Floridan aquifer at selected sites within and adjacent to the regional study area (fig. 28 and table 3). Results from these tests indicate that transmissivity in the Upper Floridan aquifer ranges from about 1,600 to 1,000,000 ft^2/d in the regional study area and from about 9,100 to 2.7 million ft^2/d in areas adjacent to the study area. Additionally, storage coefficient estimates from these tests range from about 4×10^{-4} to 0.5 (dimensionless). A review of some of the tests also indicates that some of the storage coefficient values may have wide confidence intervals, and therefore, limited significance.

Several limitations in the aquifer tests make it difficult to evaluate the accuracy of individual transmissivity estimates or to identify meaningful geographic patterns in Upper Floridan aquifer transmissivity within the regional study area. One limitation arises from various interpretations of the results of the aquifer tests, which can yield estimates of transmissivity that differ by orders of magnitude in some instances. For example, the transmissivity value derived from the Andrews Nursery aquifer test (site 8, fig. 28 and table 3) was based on a Theis curve analysis in which the curve was a poor fit to the measured drawdown data. A transmissivity estimate that is close to 10 times higher than the value shown in figure 28 can be obtained if the Jacob straight-line method is used to obtain a better fit of data from the later stages of the test period (Keith Halford, U.S. Geological Survey, written commun., 1997).

Another source of uncertainty arises because none of the tests conducted in the regional study area employed fully penetrating pumping or observation wells, and the effects of partial penetration were not considered in any of the analyses. Additionally, because most of the observation wells used in the tests penetrated less than 200 ft of the Upper Floridan aquifer, no data were available to evaluate whether vertical differences in hydraulic conductivity exist within the aquifer and whether such differences could have affected the hydraulic property estimates that were inferred from the tests (even if partial penetration effects were considered).

Another limitation is that the results of an aquifer test in a given area may not be representative of the larger area surrounding the test because of heterogeneity present at the site of the test or in the larger area. For example, data were collected at two observation wells for the Piedmont Farms test (site 23, fig. 28 and table 3). Analysis of the drawdown data from the first observation well yielded a transmissivity estimate of about 300,000 ft^2/d, using results from the Neuman method of analysis. Analysis of data from the second observation well yielded a transmissivity estimate of 1 million ft^2/d using the same method. A cursory reanalysis of this test indicated that the latter estimate may actually be greater than the reported value (about 2 million $ft^2/d \pm 1$ million ft^2/d), which indicates that the transmissivity at this site is even more spatially variable or at least that the mean value of transmissivity at this site has a wide confidence interval.

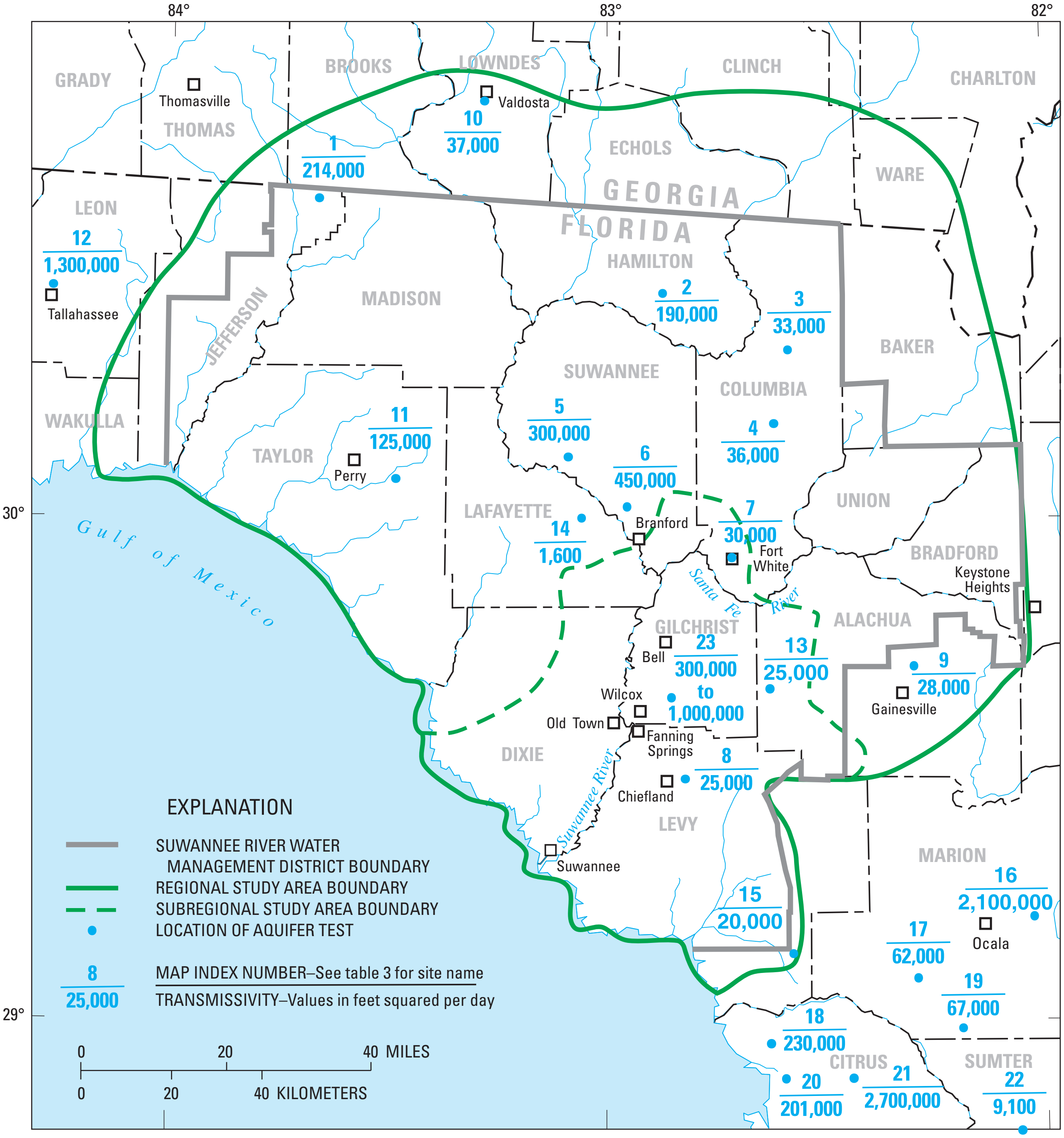


Figure 28. Location of selected sites and estimated transmissivity in the Upper Floridan aquifer based on aquifer tests.

Table 3. Transmissivity estimates from aquifer tests and flow-net analyses of the Upper Floridan aquifer.

[Well locations are shown in fig. 28. SRWMD, Suwannee River Water Management District; USGS, U.S. Geological Survey; SWFWMD, Southwest Florida Water Management District]

Site No. (this report only)	Site name	Transmissivity, in feet squared per day	Source of information
1	Finlayson	214,000	SRWMD files in Live Oak
2	Oxy	190,000	SRWMD files in Live Oak
3	Osceola National Forest	33,000	Miller and others (1978)
4	Lake City	36,000	SRWMD files in Live Oak
5	Boatright	300,000	SRWMD files in Live Oak
6	Wet Farms	450,000	SRWMD files in Live Oak
7	City of Fort White	30,000	SRWMD files in Live Oak
8	Andrews Nursery	25,000	SRWMD files in Live Oak
9	City of Gainesville	28,000	SRWMD files in Live Oak
10	City of Valdosta #4	37,000	Vorhis (1961)
11	Proctor and Gamble Foley Plant	125,000	USGS files in Tallahassee
12	City of Tallahassee #2	1,300,000	Davis (1996)
13	R.D. Williams	25,000	SRWMD files in Live Oak
14	John Folks, Division of Forestry, Midway	1,600	SRWMD files in Live Oak
15	Tidewater	20,000	Bush and Johnston (1988)
16	Silver Springs	2,100,00	Faulkner (1973)
17	Circle Square	62,000	SWFWMD (2000)
18	Florida Power – Crystal River	230,000	SWFWMD (2000)
19	Marion Oaks	67,000	SWFWMD (2000)
20	Crystal River	201,000	SWFWMD (2000)
21	Hampton Hills	2,700,000	SWFWMD (2000)
22	Tompkin Park Romp111	9,100	SWFWMD (2000)
23	Piedmont Farms	300,000 to 1,000,000	SRWMD files in Live Oak

Nearly all of the pumping and observation wells used in the aquifer tests (fig. 28) had open intervals (between the bottom of the well casing and the bottom of the well), which were constructed within the Ocala Limestone or Avon Park Formation. The remaining two tests, City of Tallahassee and Finlayson (sites 12 and 1, respectively, fig. 28 and table 3), used wells that were open to the Suwannee Limestone; the City of Tallahassee test also used observation wells open to the Ocala Limestone. No wells were completed in the Oldsmar Formation. Because of the limited number of tests available within and adjacent to the regional study area, the wide range in the reported transmissivity values, and the uncertainties in the individual estimated values, significant differences in the conductive properties of the Suwannee Limestone, Ocala Limestone, Avon Park Formation, and Oldsmar Formation were not apparent; nor were differences apparent within different depth intervals of the Upper Floridan aquifer.

Results from Previous Modeling Studies

Results from previous ground-water modeling studies represent another source of information about the hydraulic properties of the Upper Floridan aquifer. Several models have been developed for areas that include all or part of the regional or subregional study areas. The hydraulic properties used in these models represent the investigators' interpretation of the most likely values of the conductive and storage properties of the aquifer given a limited amount of water-level, flow, boundary condition, and hydrogeologic data.

Bush and Johnston (1988) developed steady-state models of the entire Floridan aquifer system that simulated predevelopment and "modern conditions" as represented by 1980 ground-water withdrawal rates. Calibrated transmissivity values from the models ranged from 10,000 to greater than 1,000,000 ft^2/d in the SRWMD and adjacent areas (with most values ranging from 250,000 to greater than 1,000,000 ft^2/d). The calibrated transmissivity values from the models differed from aquifer-test values in the SRWMD, sometimes by one order of magnitude. Bush and Johnston (1988) largely attributed these discrepancies to scale effects; for example, the tendency to include highly conductive conduits and fractures as the area being considered increases in size.

The model by Bush and Johnston (1988) was based, in part, on several small-scale models. Ryder (1985) developed a ground-water flow model of the Floridan aquifer system in west-central Florida that overlapped part of the subregional study area in Levy County and adjacent areas to the south and east. The calibrated transmissivity values of the model by Ryder (1985) ranged from 17,000 ft^2/d in western Levy County to 13,000,000 ft^2/d near large springs. Ryder (1985) also attempted a transient simulation from May 1976 to September 1976, using specific yield and specific storage values of 0.20 and $1.0x10^{-6}$, respectively, and reported a good match between simulated and measured heads in most of the model area. Calibration was not completed, however, because of uncertainties in the amount of irrigation pumpage and the values of model parameters in the southern part of the model area. Tibbals (1990) calibrated steady-state and short-term (60-day) transient models of the Floridan aquifer system in a large part of east-central Florida (adjacent to the Ryder (1985) model area). Tibbals (1990) used transmissivity values of 100,000 to 400,000 ft^2/d and a storage coefficient value of $1x10^{-3}$ in the areas of his model that were adjacent to the subregional area of the present study. Tibbals (1990) further noted that the calibration could have been improved in some locations by adjusting the storage coefficient value, but that this was not done because most of the calibration error was caused by uncertainty in the rates and areal distribution of pumping. Krause and Randolph (1989) calibrated a steady-state model of the Floridan aquifer system in southeastern Georgia and adjacent parts of Florida and South Carolina. Calibrated values of transmissivity in this model generally ranged from 100,000 to 250,000 ft^2/d in parts of Baker and Bradford Counties.

Davis (1996) developed a steady-state model of the Upper Floridan aquifer for parts of northern Florida and southern Georgia along the western boundary of the SRWMD. The calibrated transmissivity values for the model were highly variable, ranging from 5,000 ft^2/d in the region of the Apalachicola Embayment-Gulf Trough to greater than 10,000,000 ft^2/d in the area surrounding the large springs in parts of southern Leon, Wakulla, and Jefferson Counties (fig. 3).

A steady-state model was developed for a 340-mi^2 area surrounding a paper mill in Taylor County to evaluate the extent to which unpotable water from the Fenholloway River could affect the water in the underlying aquifers (Lee and Passehl, 1995). The hydraulic conductivity values calibrated for the model generally ranged from 80 to 1,200 ft/d, although a small part of the study area had a high value of 6,000 ft/d. The corresponding transmissivity values generally ranged from 64,000 to 960,000 ft^2/d, but was 4.8 million ft^2/d in the area with high hydraulic conductivity.

Sepúlveda (2002) developed a steady-state model of the intermediate and Floridan aquifer systems in peninsular Florida. This model covered most of the regional study area and subregional study areas near the Suwannee River basin, where transmissivity values ranged from 3,000 to 12,000,000 ft^2/d. The highest transmissivity values were located near springs, and the lowest transmissivity values were located in the poorly drained areas of San Pedro Bay in Lafayette County, the Waccasassa Flats in Gilchrist County, and the Keystone Heights area in eastern Bradford County (fig. 3).

Inferences from Geologic Characteristics and Potentiometric Surface Maps

Qualitative inferences about the transmissivity of one area relative to another can also be made from various other hydrologic and hydrogeologic data. These data include potentiometric surface maps, geological inferences about the depositional and post-depositional environments of different areas, estimates of recharge and runoff rates, geographic distributions of springs, and results of tracer tests. In this study, these data indicate several areas where the transmissivity of the Upper Floridan aquifer is relatively high and two areas where the transmissivity is relatively low.

Relatively high transmissivity values are indicated by the gradual slope of the potentiometric surface in the areas along the Lower Suwannee River corridor, especially east of the river in the Chiefland Limestone Plain physiographic region (fig. 3) where the slope of the potentiometric surface is 5 to 8 ft/mi; in the middle and lower reaches of the Waccasassa River drainage network where the slope is 5 to 8 ft/mi; and in western Alachua County where the slope is 2 ft/mi or less. The slope of the potentiometric surface is also a function of recharge rates, and low recharge rates can result in gradual slopes. Recharge rates in these areas, however, are relatively high, as indicated by the high rates (16-22 in/yr on average) of ground-water discharge and net rainfall (Bush and Johnston, 1988, pls. 7 and 9; Grubbs, 1997, table 5). Thus, a large volume of ground-water moves through these areas and high transmissivities are necessary to convey this water, given the low hydraulic gradient. The high density of karst features, such as sinkholes and springs, and the highly permeable surficial sediments in these areas are consistent with active dissolution of the limestone and associated high transmissivity values. High transmissivities are also indicated for the areas along and upgradient from the Ichetucknee River. This area has one of the highest concentrations of ground-water discharge in Florida, and dye-tracing studies have indicated that the time of travel between the springs and sinkholes 7-mi upgradient may be as fast as 7 days (Butt and others, 2000).

In contrast, the potentiometric surface has a steeper slope and substantially higher levels in the Waccasassa Flats area of Gilchrist County and north-central Levy County. This north-south trending potentiometric “high” coincides with the Waccasassa Flats geomorphic subprovince of the Gulf Coastal Lowlands (Col and others, 1997). Recharge rates in the Waccasassa Flats are probably similar in magnitude to those described above for the Suwannee River corridor and Waccasassa River network because the patterns of rainfall and evapotranspiration are the same, and the potential for direct (storm) runoff is limited because of the flat topography and absent or poorly developed drainage network. Thus, recharge in the Waccasassa Flats should be close to the difference between precipitation and actual evapotranspiration and, therefore, similar to that found in the adjacent areas where transmissivity is probably substantially higher. Recharge rates may be slightly lower in the Waccasassa Flats area because the potential for direct (storm) runoff during wet periods and rates of evapotranspiration should be higher because of a shallower water table, less-permeable surficial deposits, presence of wetland and lake features, and limited surface drainage in these areas (although surface storage in the form of wetlands and lakes reduces the potential for direct runoff). Thus, high elevations and steep gradients in the potentiometric surface in the Waccasassa Flats area relative to the surrounding areas appear to be explained by relatively low transmissivities in these areas rather than relatively high recharge rates.

This inference is consistent with the conclusions from several hydrogeologic studies of the Waccasassa Flats (Yon and Puri, 1962; Puri and others, 1967; Col and others, 1997). These studies and one study by Vernon (1951) describe physiographic and geologic characteristics that are consistent with at least the shallow part of the Upper Floridan aquifer having low permeability under the Waccasassa Flats. Vernon (1951) presented several lines of evidence that the Waccasassa Flats represents a former stream course, including the presence of deltaic sediments at the southern end of the flats in central Levy County. Col and others (1997) note that "... clayey sediments deposited by such a stream would thus be responsible for shielding the underlying carbonates from dissolution and supporting the generally swampy standing water conditions in the flats." Yon and Puri (1962) and Puri and others (1967) conducted detailed studies of the Waccasassa Flats and hypothesized that the silicilastic sediments above the Upper Floridan aquifer were deposited in a low-energy setting between barrier islands on the western boundary of the Waccasassa Flats and on a mainland paleo-shoreline along the eastern boundary of the flats (Col and others, 1997). They also concluded that a graben was present under the southern part of the flats, which was filled with low permeability miocene and Pliocene sediments that minimized karst dissolution because these sediments restricted recharge to the Upper Floridan aquifer (Col and others, 1997). The presence of the graben, however, could not be confirmed because of the limited number of available lithologic logs. Col land others (1997) did conclude that sediment size analyses of various intervals of borehole sediment cores indicated that "... the clastic sediments underlying the flats were deposited in quiet water conditions."

Sources and Sinks of Water

Water is supplied to and removed from the ground-water flow system as ground-water recharge, ground-water discharge, and direct withdrawals from pumped wells. Ground-water recharge and discharge occur naturally as water moves across the upper boundary of the ground-water system, either as a downward flux from land surface to the water table (recharge) or as an upward flux of ground water (discharge) to coastal wetlands and near-shore areas. Ground-water recharge and discharge also occur naturally through the exchange of water between the ground-water system and rivers, streams, and springs. This important mechanism of ground-water recharge and discharge is discussed later. Pumping wells, especially those used for irrigating agricultural crops, represent another important mechanism for withdrawing water from the ground-water system in the subregional study area.

Natural Recharge and Discharge

The ground-water system is recharged by infiltrating rainfall and seepage from streams and wetlands when water levels in the streams and wetlands are higher than the water-table elevation. In areas that lack surface drainage, average annual recharge can be approximated as the difference between precipitation and evapotranspiration and ranges from 18 to 23 in/yr (Bush and Johnston, 1988, pls. 7 and 9); interception and changes in soil moisture storage can also affect recharge, but these terms become negligible as the averaging time increases and the thickness of the unsaturated zone decreases. This range is similar to average annual recharge rates of 14 to 20 in. found using eddy correlation and water-budget methods (Knowles, 1996), 15 in. using a chloride tracer approach (Lee, 1996), and 18 to 24 in. using a water-budget method (Grubbs, 1997). In other areas that have surface drainage, recharge is reduced somewhat by direct runoff. Streamflow data from basins with well developed drainage networks in the Northern Highlands and from Cow Creek (in Gilchrist County) and the Waccasassa River indicate that direct runoff is generally about 3 to 6 in/yr. This result indicates that in some of the wetland areas with limited drainage networks, such as the Waccasassa Flats, Mallory Swamp, and unnamed coastal swamps (fig. 3), recharge to the Upper Floridan aquifer may be as much as 3 to 6 in/yr lower than the difference between annual precipitation and evapotranspiration.

Daily rainfall was estimated over a set of Thiessen polygons, using data collected at 19 stations from October 1996 to September 1999 (water years 1997-99). Daily potential evapotranspiration was estimated by multiplying the daily pan evaporation measured at the National Weather Service station, Gainesville 11 WNW (National Oceanic and Atmospheric Administration coop identifier, 083322) during water years 1997-99 by pan coefficients of 0.62 (April-October) and 0.82 (November-March). These values yield a weighted annual average pan coefficient of 0.70, which is a typical value for calculating potential evapotranspiration in Florida (Jones and others, 1984).

Daily estimates of actual evapotranspiration and recharge were computed for the subregional study area, using the above daily precipitation and potential evapotranspiration data for water years 1996-99 by means of the following daily soil-moisture deficit accounting (SMA) algorithm (E.P. Weeks, U.S. Geological Survey., written commun., 1993):

$$SMD(t) = SMD(t-1) + ET_A(t) - P(t), \qquad (1)$$

where

$SMD(t)$ is the soil-moisture deficit on day t (the current day),

$SMD(t-1)$ is the soil-moisture deficit on the previous day,

$ET_A(t)$ is the actual evapotranspiration on day t, and

$P(t)$ is the precipitation (rainfall) on day t.

The actual evapotranspiration is calculated from a function of potential evapotranspiration and the soil-moisture deficit:

$$ET_A(t) = ET_P(t) \ \text{ if } SMD(t) \le SMD_1, \qquad (2)$$

$$ET_A(t) = ET_P(t)\left[1 - \frac{SMD(t) - SMD_1}{SMD_2 - SMD_1}\right] \ \text{ if } SMD_1 \le SMD(t) \le SMD_2, \text{ and}$$

$$ET_A(t) = 0 \ \text{ if } SMD(t) > SMD_2 ,$$

where

$ET_P(t)$ is the potential evapotranspiration rate on day t,

SMD_1 is the soil-moisture deficit value beyond which $ET_A = ET_P$, and

SMD_2 is the soil-moisture deficit value beyond which $ET_A(t) = 0$.

Equations 1 and 2 were solved iteratively until changes in the values of $SMD(t)$ and $ET_A(t)$ were negligible. On days when rainfall was greater than $SMD(t\text{-}1)$ and $ET_A(t)$, $SMD(t)$ was negative, indicating an excess amount of moisture in the soil zone on that day. Accordingly, recharge was set equal to the absolute value of $SMD(t)$ on these days and $SMD(t)$ was set to zero. This procedure is consistent with the complete drainage of the excess soil moisture from the soil zone during that day and the concomitant flux of this excess moisture to the water table as recharge.

Results obtained from the SMA algorithm were consistent with previous evapotranspiration measurements taken within and near the subregional study area, and with expected recharge rates. For example, the evapotranspiration rates computed with the SMA algorithm were 31.3 and 31.9 in/yr for water years 1998 and 1999, respectively, in the vicinity of the National Weather Service station near Cross City (fig. 4). Actual evapotranspiration at the closest site (about 11 mi southeast of Gainesville) with measured data was 32 in. for water year 1994 (Knowles, 1996). Sumner (2001) measured somewhat higher rates of 36 and 42 in. for calendar years 1998 and 1999, respectively, at a site farther south in central Volusia County, east-central Florida. At the weather station near Cross City, the annual recharge rates computed with the SMA algorithm were 46 and 23 in/yr for water years 1998 and 1999, respectively. The recharge rate calculated at the Cross City weather station for water year 1998 was substantially higher than the difference between long-term average estimates of precipitation and actual evapotranspiration of 17 in/yr (Bush and Johnston, 1988, pls. 7 and 9) because of above average rainfall (78 in.) during water year 1998. Conversely, the recharge rate was similar to the estimated long-term average recharge rate during water year 1999 when rainfall was more typical (55 in.).

Similar patterns in the annual recharge rates for water years 1998 and 1999 were evident in the rest of the subregional study area (fig. 29). During water year 1998, recharge rates were substantially higher than typical values, ranging from 35 to 75 in. (fig. 29) or two to three times greater than the 14- to 24-in/yr range from the estimates of Knowles (1996), Lee (1996), and Grubbs (1997). The maximum value (75 in.) occurred in the Thiessen polygon corresponding to the Wekiva Fire Tower rain gage where rainfall measured 104 in. This rainfall total was close to the record rainfall totals for Florida (107 in.), and for the nearby Usher Fire Tower weather station near Chiefland where 99 in. of rain fell in 1964. Rainfall totals at the other gages were also higher than normal, ranging from 65 to 96 in. During water year 1999, recharge rates were lower, ranging from 13 to 25 in. (fig. 29), and were consistent with the smaller rainfall totals (41 to 55 in.) recorded for the year.

Over shorter time periods, the SMA algorithm produced estimated recharge rates that were also consistent with ground-water level fluctuations. An example of this relation is presented in figure 30 for the Cross City area. The timing and magnitude of ground-water level rises generally coincided with the timing and magnitude of SMA-estimated recharge events. Additionally, rainfall occurred without a corresponding increase in water levels during several periods. Estimated recharge rates were generally equal to zero during these periods, indicating that the SMA algorithm was a reasonably

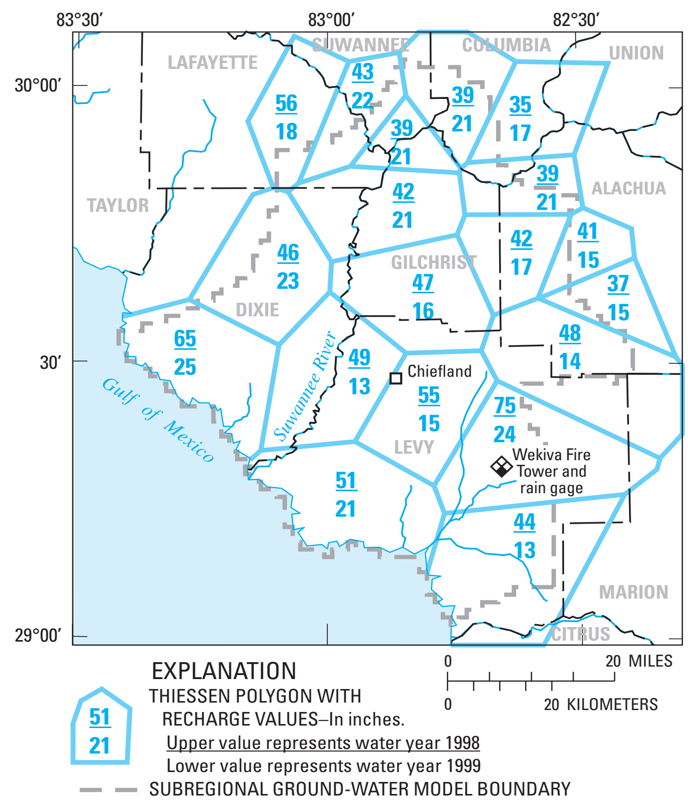


Figure 29. Recharge totals in the subregional study area based on results of the soil-moisture accounting method, water years 1998 and 1999.

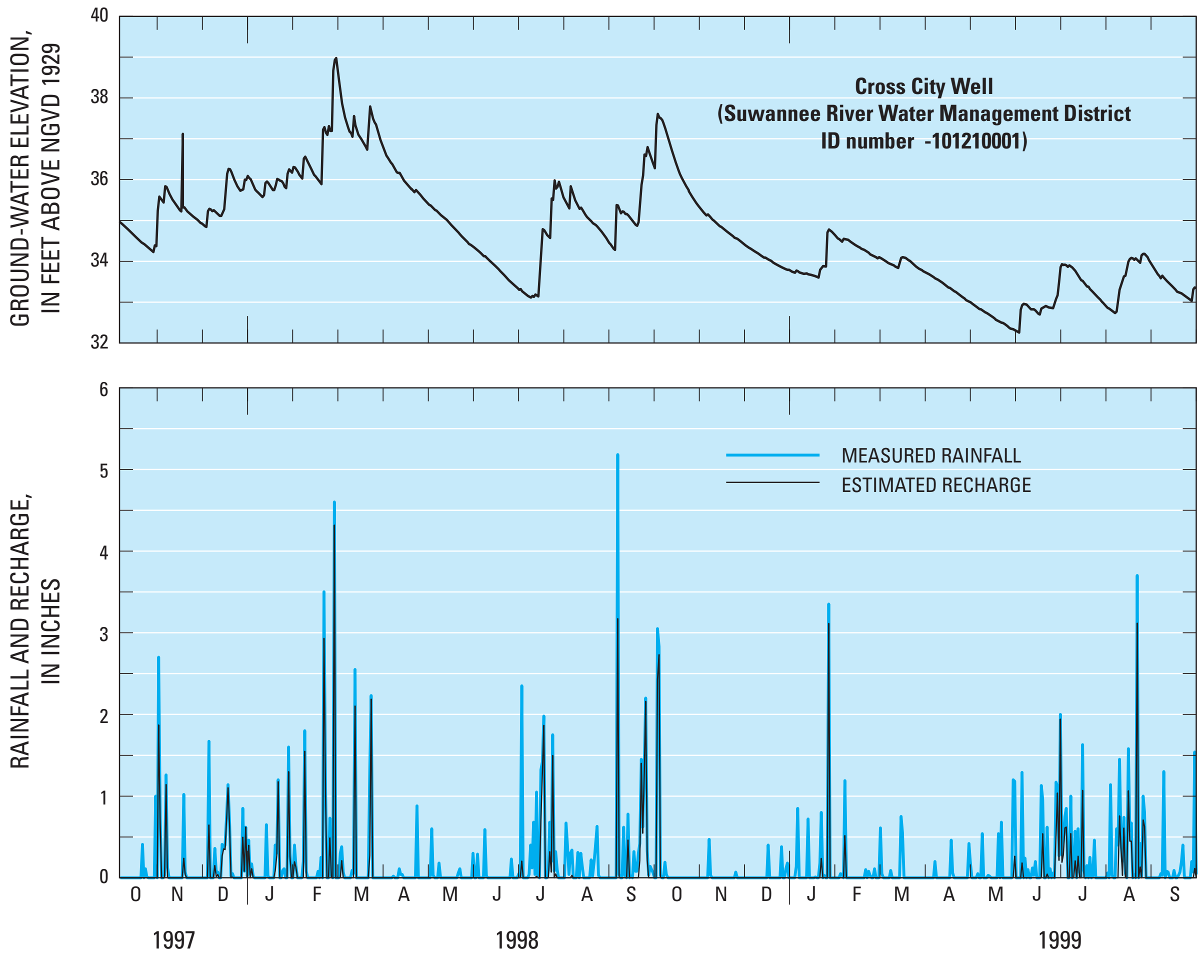


Figure 30. Daily ground-water levels, rainfall, and estimated recharge for Cross City, Florida, during water years 1998-99. Well location is shown in fig. 27 (see map identifier s328).

good predictor of the effects of soil-moisture depletion and replenishment on recharge. The algorithm, however, was not a perfect predictor of recharge. There were a few periods when water-level rises occurred but the SMA algorithm failed to predict recharge (for example, the small water-level rise during March 1999 in the Cross City well), or conversely, when the water levels declined despite SMA-algorithm predictions of recharge sufficient to raise ground-water levels.

Although natural discharge from the Upper Floridan aquifer takes place primarily as ground-water discharge to streams, natural discharge also occurs from the Upper Floridan aquifer to coastal wetlands, tidal creeks, and near-shore discharge to the Gulf of Mexico. Analysis of the potentiometric surface maps of the Upper Floridan aquifer indicates that the ground-water contributing area to these coastal areas is about 700 mi^2. Assuming a long-term recharge rate of about 20 in/yr to this contributing area, the long-term average discharge from the Upper Floridan aquifer to the coastal areas is on the order of 1,000 ft^3/s. During the study period, recharge was higher (about 38 in/yr) than the long-term rate, so discharge from the Upper Floridan aquifer to the coastal areas was probably closer to 2,000 ft^3/s.

Well Pumpage

Data from monthly measurements of ground-water withdrawals for public, self-supplied commercial and industrial, and power-generation uses were obtained from the permit files at the Southwest Florida Water Management District and the SRWMD and from the water-use data bases of the St. Johns River Water Management District, and the Florida Department of Environmental Protection. Additional monthly data were obtained from several ground-water users. These data were then merged with data describing the location of the wells that were the source of the withdrawals. To simplify this process of merging the two types of information (withdrawal and location data), data were not merged if a user withdrew less than a monthly average of 200,000 gal or less than a daily average of 10,000 gal.

Agricultural withdrawals in the subregional study area were estimated using data describing the type, amount, and irrigation requirements of each crop grown during the study period. The crop data were compiled for each county in the study area, using the following crop categories: fruit crops, field crops, vegetables, ornamentals, and grasses. The area

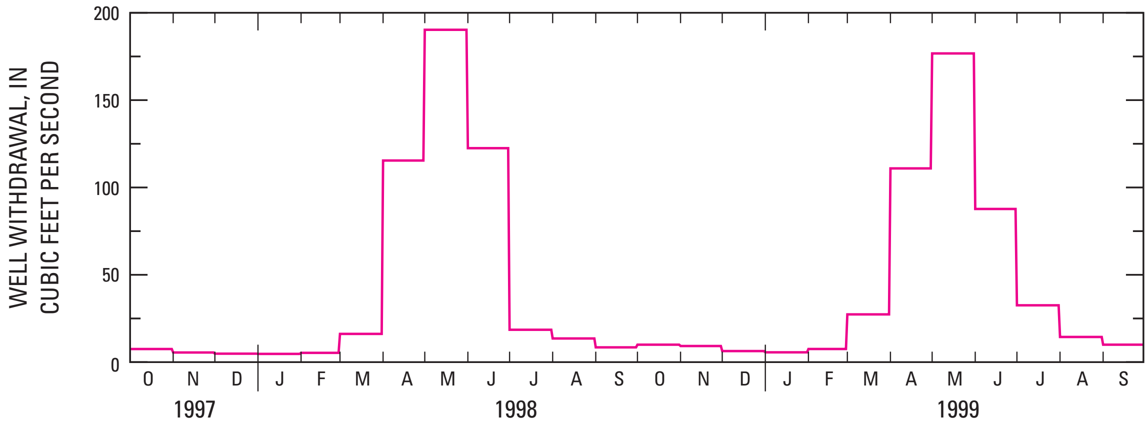


Figure 31. Estimated monthly ground-water withdrawals from wells in the subregional study area during water years 1998-99.

of each crop grown in the study area was obtained from the county extension offices of the University of Florida Institute of Food and Agricultural Sciences. Monthly water-use values were estimated by interpolating between the normal (50-percent probability of occurrence) and dry-period (20-percent probability of occurrence) values of monthly net irrigation requirements (the irrigation required after accounting for rainfall) for each crop, using the following formula:

$$NIR_{i,j}(p) = NIR_{i,j}\,(0.2) - \frac{[NIR_{i,j}\,(0.50) - NIR_{i,j}\,(0.20)]}{(0.50 - 0.20)}\,(0.20 - p); \quad (3)$$

where

NIR is the interpolated net irrigation requirement (NIR);

i is an index representing a particular crop category (for example, $i = 1$ might indicate the fruit crop category, $i = 2$ the field crop category, and so forth);

j is an index representing the months of the year ($j = 1$ indicates January, $j = 2$ indicates February, and so forth until $j = 12$, which indicates December);

p is probability of occurrence of the observed monthly precipitation total for a particular month during the study (for example, June 1999) in National Weather Service Climate Zone 2 for the State of Florida (*p* is calculated by comparing the value for that month with historical values from climate zone 2);

$NIR_i\,(0.20)$ is the dry-season NIR value for crop category *i* and month *j*; and

$NIR_j\,(0.50)$ is the normal-season NIR value for crop category *i* and month *j*.

The agricultural withdrawal for a given crop type during a given month of the study was then estimated by multiplying the area of a crop by the interpolated NIR value, then dividing this value by an estimated irrigation efficiency of 70 percent (Richard Marella, U.S. Geological Survey, written commun. 2000).

During water years 1998-99, the average withdrawal rate from the Upper Floridan aquifer in the subregional study area was estimated to be about 35 ft^3/s. There was considerable variability about this mean value: (1) the minimum monthly withdrawal was about 5 ft^3/s and occurred in January of both water years; (2) the maximum monthly withdrawal rates were about 190 ft^3/s and 175 ft^3/s and occurred in May 1998 and 1999, respectively (fig. 31). Most of this variability was due to seasonal changes in irrigation rates. The crop data indicate that negligible irrigation occurs in December and January and that rates of irrigation withdrawals peak in May, reaching rates of 130 to 140 ft^3/s. Agricultural withdrawals also accounted for most (80 percent) of the withdrawals occurring within the subregional study area during water years 1998-99. The remaining withdrawal categories (public supplies and self-supplied commercial and industrial withdrawals) showed a seasonal pattern of withdrawal rates similar to the agricultural withdrawals, but substantially less variable. The average monthly withdrawal for all of these less variable uses ranged from a minimum of 5 ft^3/s in January 1998 to a maximum of 15 ft^3/s in May 1998.

The geographic distribution of ground-water withdrawals from the Upper Floridan aquifer in the subregional area is shown in figure 32. Most of the withdrawals occur in the Chiefland Limestone Plain of the Gulf Coastal Lowlands and the Central Highlands areas (fig. 3). These areas are typically underlain by well-drained soils that are favorable for agricultural uses and development of cities, towns, and homesteads. The largest withdrawals occur in the towns of Cross City, Trenton, Chiefland, and High Springs and in irrigated areas near Chiefland, Trenton, east of Manatee Springs, and southwest of Williston.

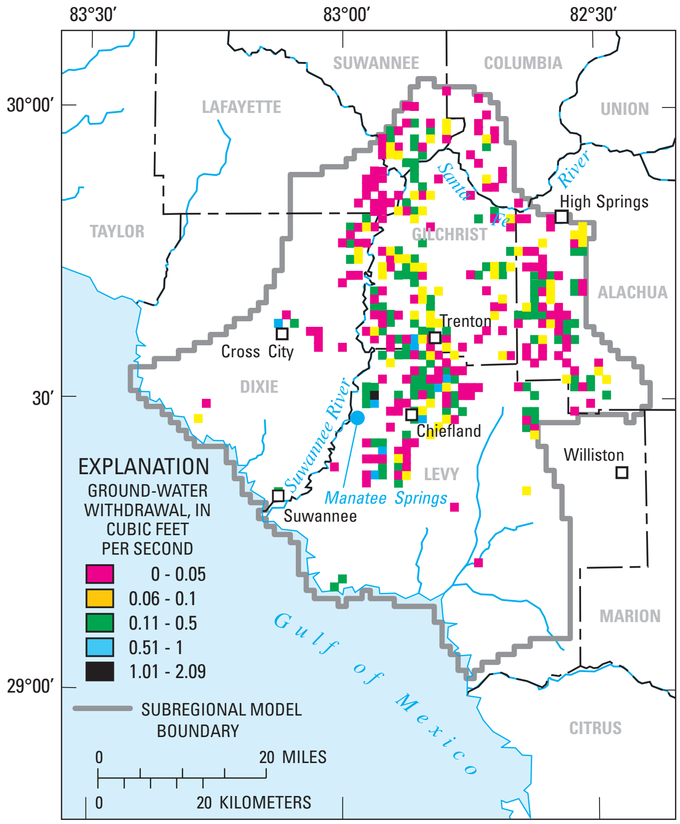


Figure 32. Geographic distribution of ground-water withdrawals from the Upper Floridan aquifer during water years 1998-99.

Evaluation of Ground-Water and Surface-Water Exchanges Using Hydrologic Data

The interactions between the Upper Floridan aquifer and the Lower Suwannee and Lower Santa Fe Rivers represent an important process that affects the flow and water-level characteristics of these rivers. A key element in understanding these interactions is quantifying the magnitude of water exchanges across the river-aquifer interface and how they vary with time and location along these two rivers. A necessary first step for developing a more comprehensive understanding of processes that are affected by streamflows and water levels was to use hydrologic data to evaluate how ground- and surface-water exchanges along the Lower Suwannee and Lower Santa Fe Rivers vary in time and space along different river reaches.

Temporal Characteristics

Ground- and surface-water exchanges in the subregional study area vary across time scales that range from hours to decades and longer. Inferences about these temporal variations were made using data collected at long-term gaging stations and from miscellaneous measurements taken before and during the study. Streamflow data from the long-term gaging stations on the Suwannee River at Branford (map index no. 5, fig. 4), Santa Fe River near Fort White (map index no. 14, fig. 4), and Suwannee River near Wilcox (map index no. 7, fig. 4) indicate that the difference between the combined streamflow from the two upstream gaging stations (Branford and Fort White) and the downstream gaging station (Wilcox) was about 1,500 ft^3/s for flows measured during climate years 1942-2000. (Climate years are defined by the year beginning on April 1 and ending on March 30 of the following year.) The difference (pickup) between the upstream and downstream flow in this reach (referred to as reach 1, fig. 11) represents about 15 percent of the long-term, average annual streamflow at the Wilcox gaging station (10,000 ft^3/s). In about half of these years, the average annual pickup in this reach was between 1,200 and 1,900 ft^3/s, and the minimum and maximum annual pickup values were about 500 and 3,000 ft^3/s, respectively (figs. 33 and 34). Streamflow pickup can be used to estimate the amount of ground water discharged to this reach of the Lower Suwannee and Lower Santa Fe Rivers because: (1) direct runoff to this reach is minimal, and (2) changes in the amount of water stored within the river channels and floodplains (surface storage) are typically small or negligible over an annual averaging period, and therefore, should not substantially affect the annual pickup values. The accuracy of the pickup-derived estimates of ground-water discharge, however, is still affected by the uncertainties of the daily discharge data (from individual gaging stations) that are used to compute streamflow pickup rates.

Inspection of the time-series plots and scatter plots of annual pickup relative to streamflow at the Wilcox gaging station indicates that the magnitude of pickup in reach 1 has a significant (but weak) positive correlation with the streamflow measured at the Wilcox gaging station (fig. 35). Thus, small pickup values tend to occur in years with low flows, and large pickup values tend to occur in years with high flows. This relation is not surprising given that: (1) the annual flow of the Suwannee River near Wilcox is correlated with rainfall, and thus, recharge; and (2) year-to-year differences in the magnitude of this pickup are determined by year-to-year differences in recharge occurring within the contributing area to reach 1. The relation between streamflow and pickup, however, is subject to a fair amount of scatter because flow in the Suwannee River near Wilcox is more strongly related to upstream runoff entering the Suwannee and Santa Fe Rivers from outside of the subregional study area than to recharge occurring within the contributing area of this reach.

Analysis of the time series of the annual pickup values also indicates that there is a substantial amount of interannual correlation in streamflow pickup. Thus, even though annual streamflow pickup along reach 1 tends to increase as flow in the Suwannee and Santa Fe Rivers increases, there may be a lag in this response. For example, annual streamflows at the Wilcox gaging station increased sharply from climate years 1963 to 1964, and then declined sharply from 1964 to 1968 (fig. 34). The annual pickup values also began to increase (albeit less abruptly than the increase in streamflows) in 1963, but continued to increase in the two climate years following

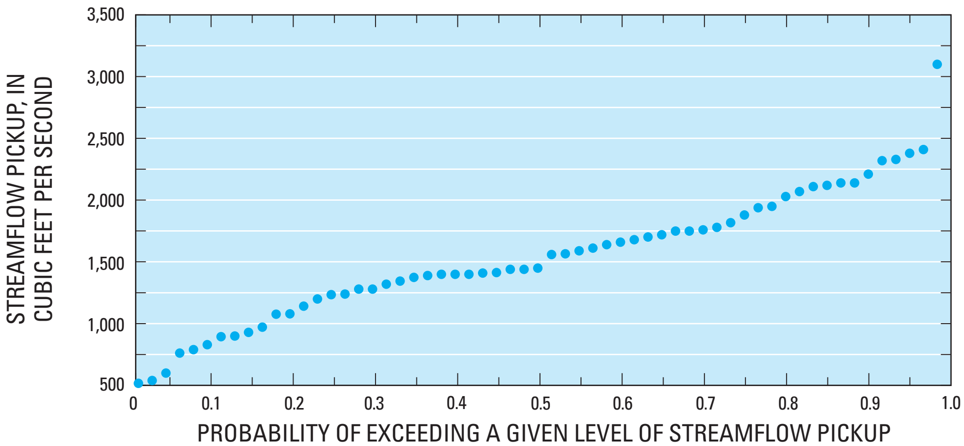


Figure 33. Cumulative distribution of the median annual change in streamflow (streamflow pickup) in reach 1 of the Lower Suwannee and Santa Fe Rivers. Location of river reach 1 is shown in fig. 11.

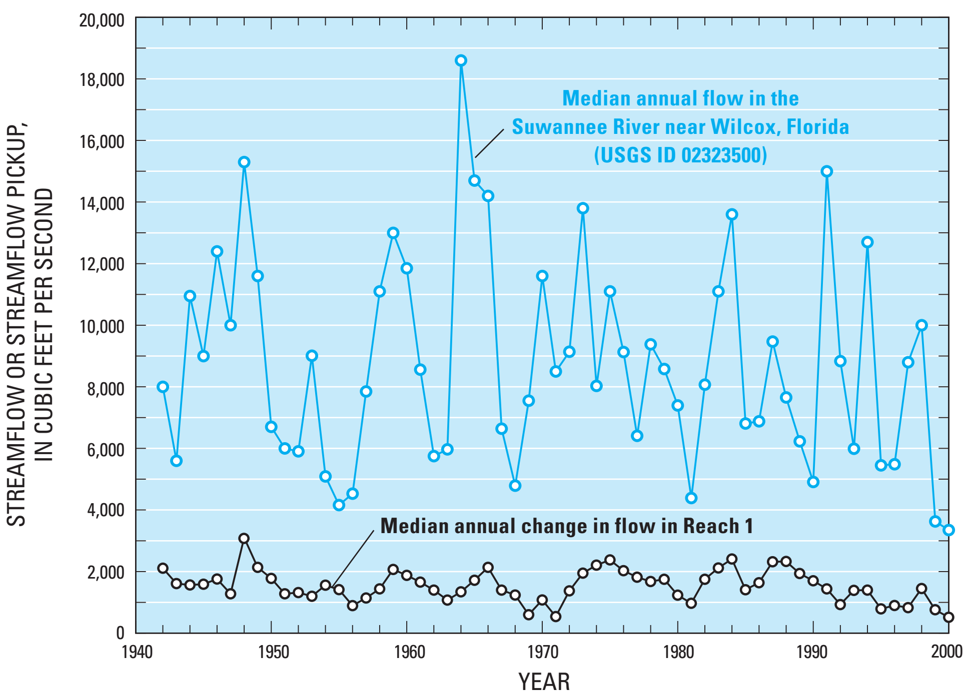


Figure 34. Time series of annual streamflow in the Suwannee River near Wilcox, Florida (map index no. 7 in fig. 4), and the streamflow pickup in reach 1 of the Lower Suwannee and Santa Fe Rivers during 1941-2000. Location of river reach 1 is shown in fig. 11. USGS is U.S. Geological Survey.

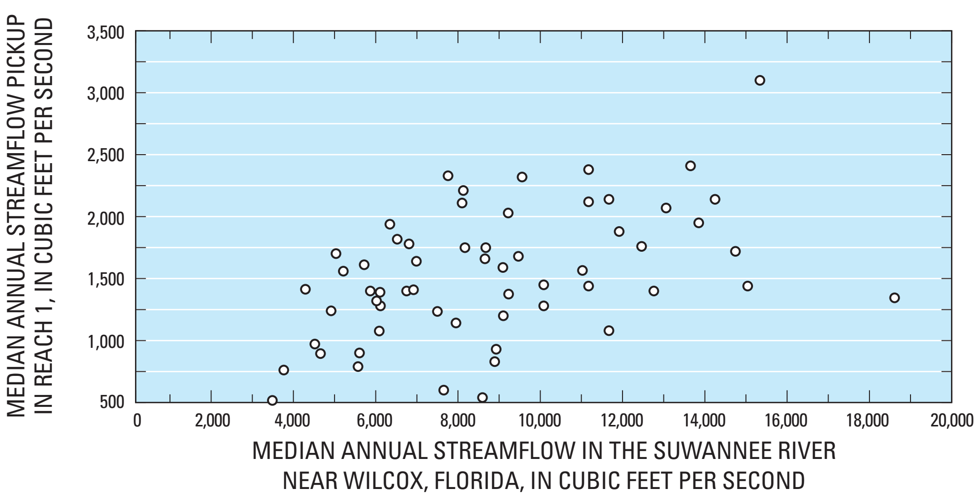


Figure 35. Relation between annual streamflow in the Suwannee River near Wilcox, Florida (map index no. 7 in fig. 4), and the streamflow pickup in the Lower Suwannee and Santa Fe Rivers. Location of river reach 1 is shown in fig. 11.

the peak annual streamflow in 1964. Additionally, the annual pickup increased by a factor of 2 from the trough (in 1963) to the peak (in 1966) in the annual pickup time series. This was substantially smaller than the change in the streamflows at Wilcox where the peak flow in 1964 was more than three times the flow in 1963. These interannual correlations arise because changes in the ground-water flow system occur more slowly than in the river system, and the effects of stresses on the ground-water system are smoothed over longer time periods.

Streamflow-pickup rates can also be used to evaluate variations in the magnitude of the ground- and surface-water exchanges over monthly and seasonal time scales. Although the median monthly pickup values for the 1942 to 2000 period were similar for all calendar months, there was a general pattern of decreasing values from January to March, when the minimum monthly pickup of about 1,000 ft^3/s occurred (fig. 36). This period of decreasing pickup values was followed by a period of increasing values from March to May, when the maximum monthly pickup of about 1,800 ft^3/s occurred (fig. 36). This pattern appears to be associated with conditions that occur during the late winter to early spring flood period. The January to March period typically coincides with rising river levels associated with flooding. Ground-water discharge to reach 1 may be lower during this period because river levels are generally high relative to ground-water levels. From March to May, water levels in the rivers are typically receding from peak flood levels. Ground-water discharge to reach 1 may be higher during this period because river levels are low relative to ground-water levels.

Rates of ground- and surface-water exchanges may not reflect the pattern of decreasing mean and median streamflow pickup rates from January to March followed by increasing rates from March to May (fig. 36) because changes in surface storage affect streamflow-pickup rates. Streamflow pickup is approximately equal to ground- and surface-water exchanges minus the rate of change in surface storage. Therefore, ground-water discharge to reach 1 is typically greater than pickup during the January to March period because surface storage is increasing as river levels rise in the river channel and floodplains. Conversely, ground-water discharge to reach 1 is typically less than pickup during the March to May period because surface storage is decreasing as river levels decline in the river channel and floodplains.

The effects of surface storage on streamflow pickup are even more pronounced over shorter time periods, such as within seasons or during shorter periods of rising or receding river levels preceding or following a peak river level. An attempt was made to account for surface storage effects at this time scale to evaluate the short-term changes in river-aquifer water exchanges before, during, and after flooding events. This was done by: (1) calculating the daily pickup occurring along the Suwannee River between the Branford and Wilcox gaging station (map index nos. 5 and 7, fig. 4) and Santa Fe River downstream from the Fort White gaging station (map index no. 4, fig. 4) during water years 1998-99; and (2) estimating daily ground- and surface-water exchanges in reach 1 by correcting the daily pickup calculations with estimates of the daily rate of change in the amount of water stored in the river channels and floodplains. This "corrected-pickup" estimate of river-ground-water exchange was computed as follows:

$$Q_{gw \leftrightarrow sw}(t) = Q_{pickup}(t) - S_{sw}(t) \qquad (4)$$

where

t is time in days;

$Q_{gw \leftrightarrow sw}$ is the average water exchange between the river reach and the aquifer on day t, in cubic feet per second;

$Q_{pickup}(t)$ is equal to the mean daily flow in the Suwannee river near Wilcox minus the sum of the mean daily flows of the Suwannee River at Branford and the Santa Fe River near Fort White on day t, in cubic feet per second; and

$S_{sw}(t)$ is the rate of change in the amount of water stored in the river channels and floodplains on day t, in cubic feet per second.

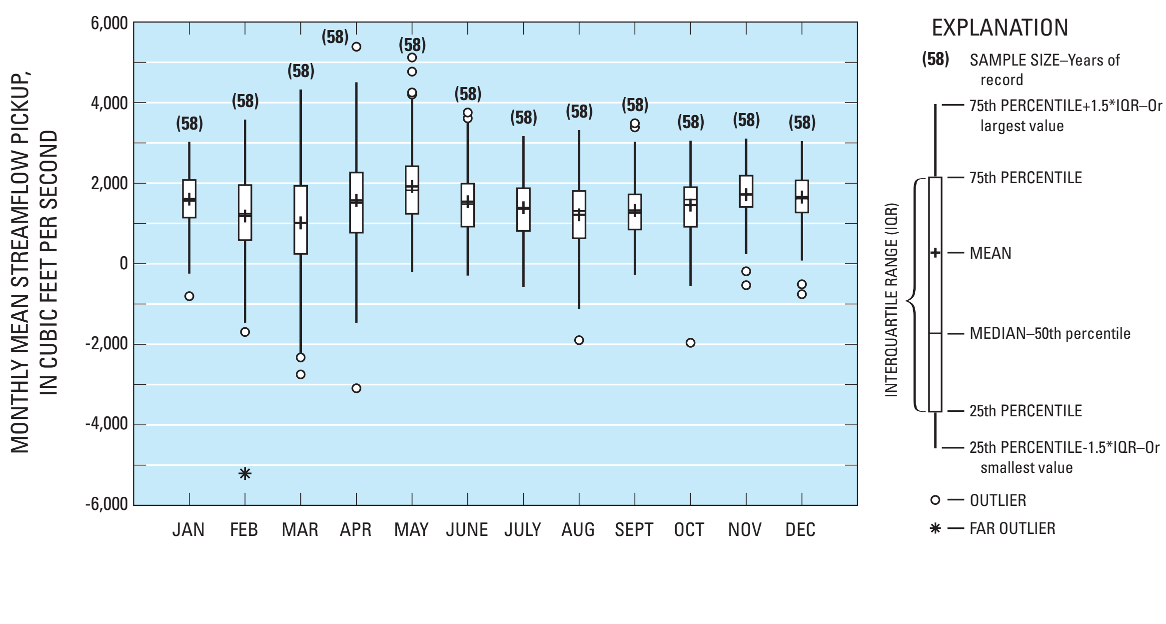


Figure 36. Monthly mean streamflow pickup in reach 1 of the Lower Suwannee and Santa Fe Rivers. Location of river reach 1 is shown in fig. 11.

The variable, $S_{sw}(t)$, was estimated by computing the surface storage for each day, $S_{sw}(t)$, then computing the difference between the surface storage on a given day from the surface storage on the previous day and dividing by 86,400 seconds per day:

$$S_{sw}(t) = \left[S_{sw}(t) - S_{sw}(t{-}1)\right]/(86400) \quad (5)$$

These corrected-pickup estimates of $Q_{gw\leftrightarrow sw}(t)$ should be interpreted with caution because they can be sensitive to errors in the daily flow estimates from the gaging stations and to errors in estimating the surface storage. This is especially true during periods of overbank flow, when errors in the streamflow or surface storage could be much larger than the estimated values of $Q_{gw\leftrightarrow sw}(t)$.

The estimates of the average rate of ground- and surface-water exchanges along reach 1 (using eq. 4) during water years 1998-99 were somewhat lower than the long-term average of 1,500 ft^3/s during 1951-99. The net ground- and surface-water exchanges averaged about 1,000 and 1,100 ft^3/s during water years 1998 and 1999, respectively. The $Q_{gw\leftrightarrow sw}(t)$ data computed using equation 4 also indicated that aquifer-to-river leakage (ground-water outflow) rates into the study reach were as high as 7,600 ft^3/s and followed the patterns in the surface storage, $S_{sw}(t)$, hydrograph (fig. 37). The statistical significance of the above estimates of the calculated rates, however, are limited. For example, the 95-percent confidence interval for the average net ground- and surface-water exchange was from -1,500 to 3,600 ft^3/s during water years 1998-99 (where negative values indicate river-to-aquifer leakage). The uncertainty of the above-estimated value of the maximum rate was even greater (with a 95-percent confidence interval from 1,300 to 13,900 ft^3/s) because of uncertainty in estimating the $Q_{gw\leftrightarrow sw}(t)$ and $S_{sw}(t)$ terms in equation 4.

More accurate estimates from equation 4 are possible if the analysis is limited to conditions when river levels are not changing rapidly. The longest period of relatively stable river levels along reach 1 occurred from November 1998 to September 1999, when the corrected pickup estimate of average ground-water exchange was about 1,200 ft^3/s, with a corresponding 95-percent confidence interval ranging from about 300 to 2,100 ft^3/s.

The peak river-to-aquifer leakage ($Q_{gw\leftrightarrow sw}(t) < 0$) rates were estimated to be about 3,100 and 1,900 ft^3/s in water years 1998 and 1999, respectively (fig. 37). Again, these peak values are insignificant when compared with uncertainties in the underlying data that were used to compute them. The 95-percent confidence intervals for peak river-to-aquifer leakage rates were about ±7,000 and ±3,400 ft^3/s during water years 1998 and 1999, respectively.

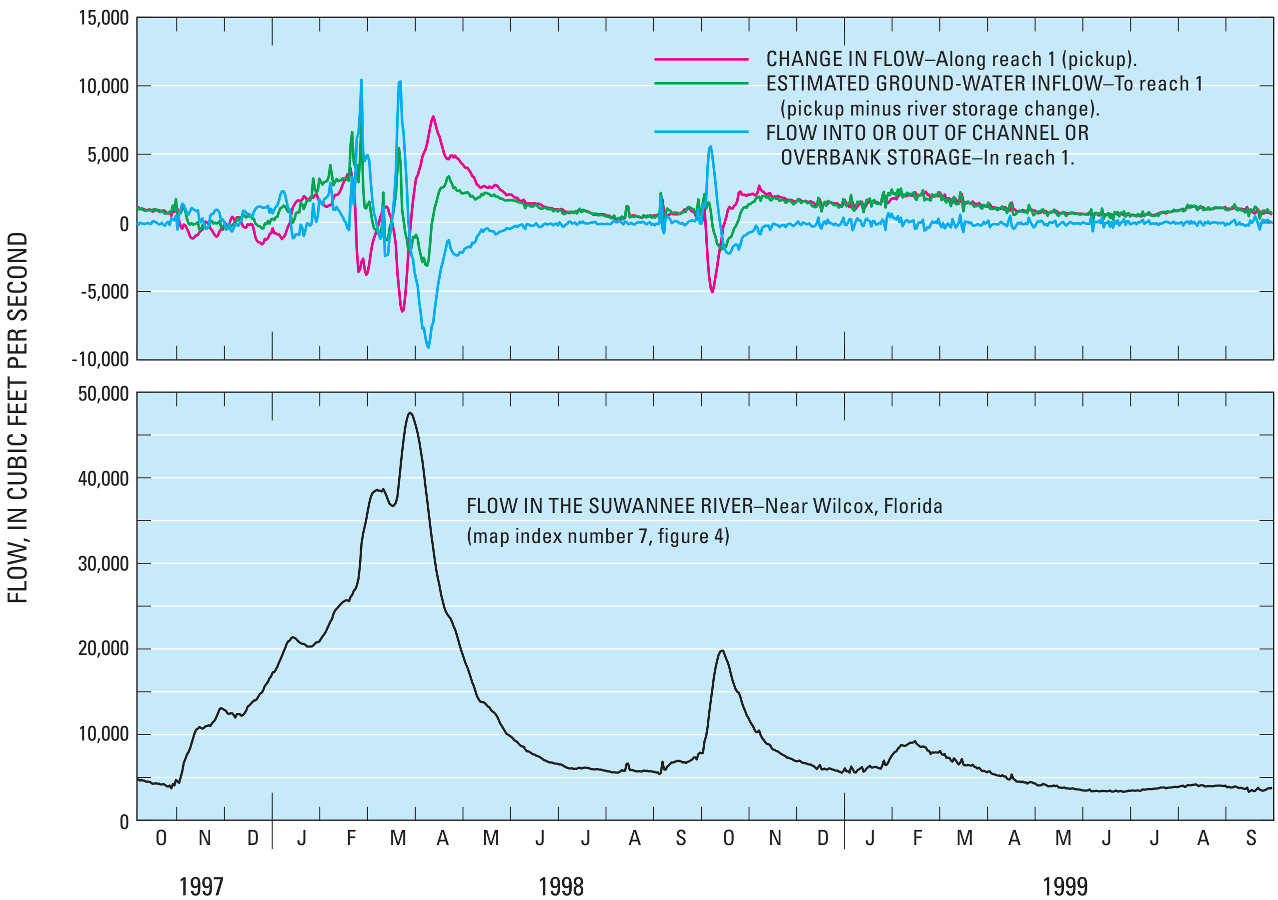


Figure 37. Relation between streamflow, streamflow pickup, surface storage, and river-aquifer water exchanges. Location of river reach 1 is shown in fig. 11.

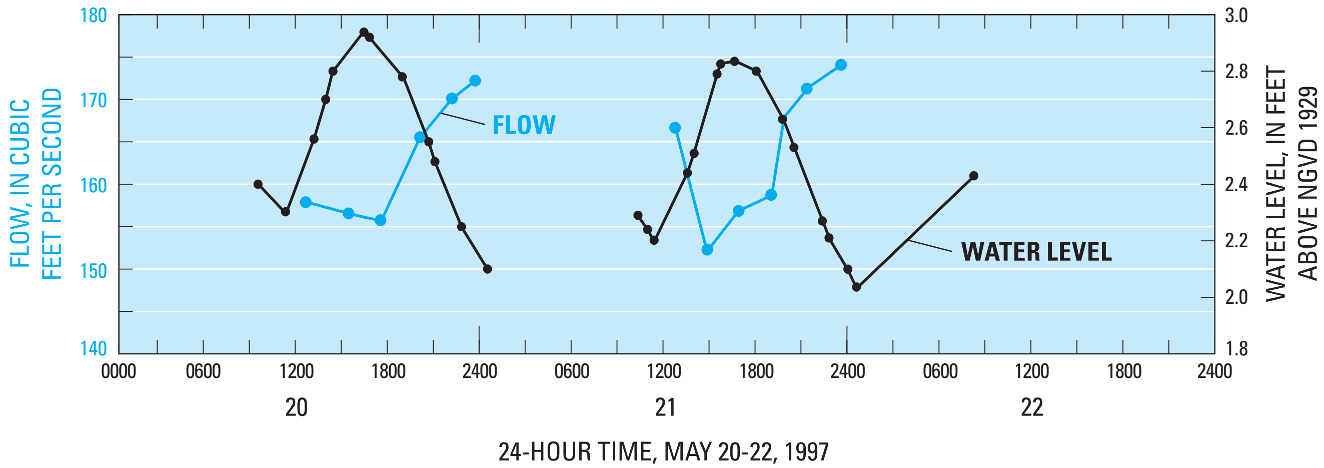


Figure 38. Tidally induced changes in spring pool elevation and spring flow in Manatee Springs during May 1997.

Temporal variations were evident over the course of tidal cycles. These variations were apparent in measurements taken at Manatee Springs during several spring-tide periods, when the difference between river levels at lower-low and higher-high tides was greatest. Results from one set of measurements taken during May 1997 indicate that the median flow was about 160 ft^3/s, and measured flow rates ranged from 152 to 175 ft^3/s (fig. 38). The flow of Manatee Springs (fig. 4) tended to increase during periods when the spring pool elevation was falling (in response to the ebbing stage in the Suwannee River) and to decrease when the pool elevation was rising. During this same period, the median flow from Fanning Springs (fig. 4) was about 80 ft^3/s, and ranged from about 70 to 90 ft^3/s. The range in spring-flow variations at both sites was about ±10 ft^3/s even though the range in spring pool elevation was about 0.25 ft higher at Manatee Springs, which indicates that smaller springs may have a wider range of flows during a tidal cycle, perhaps because there is less momentum to resist the variable stress imposed by the tidally induced fluctuations in spring-pool elevations.

Spatial Characteristics

Hydrologic data collected before and during the study period can be used to make inferences about how ground- and surface-water exchanges might vary along different reaches of the Lower Suwannee and Lower Santa Fe Rivers. Two sets of hydrologic data were used for this purpose: (1) locations of springs and estimates of their median flow rates, and (2) measured streamflows at the upstream and downstream limits of selected river reaches. These data were also used to evaluate what fraction of the total ground- and surface-water exchange might be accounted for by known springs.

More than 50 springs in the subregional study area (fig. 2) contribute to the flow of the Lower Suwannee and Lower Santa Fe Rivers (Rosenau and others, 1977; Hornsby and Ceryak, 1998; Scott and others, 2004; fig. 2). Five springs can be considered first-order springs (average discharge greater than 100 ft^3/s), and most of the remaining are second-order springs (average discharge greater than 10 ft^3/s). Nine of these springs and numerous unnamed springs contribute about 350 ft^3/s to the upper reaches of the Ichetucknee River, which drains the Santa Fe River at a point 7.1 mi upstream from the mouth of the Santa Fe River.

Spring-flow totals were computed for two river reaches and compared to the total average pickup estimated for these reaches. This calculation gives an estimate of how much of the total ground- and surface-water exchange is derived from known springs and how much is derived from unnamed springs or more diffuse exchanges. Two reaches were evaluated for this analysis (fig. 11): (1) reach 1 (as previously described) along the Santa Fe River downstream from the Fort White gaging station and along the Suwannee River from Branford to the Wilcox gaging station (this reach also includes the Ichetucknee River); and (2) reach 2 along the Santa Fe River from U.S. Highway 441 to the Fort White gaging station. For reach 1, the springs identified by Rosenau and others (1977) accounted for about 800 ft^3/s or about 55 percent of the estimated 1,500 ft^3/s net ground-water outflow during water years 1942-99. If the additional springs identified by Hornsby and Ceryak (1998) are included, then springs account for about 1,300 ft^3/s or about 85 percent of the long-term average net ground-water outflow in this reach. For reach 2, the springs identified by Rosenau and others (1977) accounted for about 600 ft^3/s or about 75 percent of the estimated 800 ft^3/s net ground-water outflow during water years 1942-71 (the period of concurrent data for the two gaging stations that define the limits of this reach). If the additional springs identified by Hornsby and Ceryak (1998) are included, then springs account for about 700 ft^3/s or about 90 percent of the long-term average net ground-water outflow in this reach. Data from reaches 1 and 2 indicate that springs are the dominant source of ground water to the Lower Suwannee and Lower Santa Fe Rivers.

Closer inspection of the spring-flow data reveals that spring flow is clustered along some reaches and more evenly distributed along other reaches. A substantial fraction of the ground-water outflow measured in reaches 1 and 2 is derived from clusters of springs. For example, the springs clustered around Ginnie Springs on the Santa Fe River just upstream from the gaging station near Fort White (fig. 2) account for about 40 percent of the long-term average net ground-water outflow to reach 2. Similarly, the Ichetucknee Springs group (fig. 2) accounts for about 25 percent of the long-term average net ground-water outflow to reach 1.

Another location of concentrated flow occurs in the reach of the Suwannee River between the gaging stations near Old Town and Wilcox (station indexes 8 and 7 at river miles 23.5 and 33.5, respectively, fig. 11). The average spring discharge along this reach was 330 ft^3/s (about 24 ft^3/s per mile of river), and is derived from three springs (fig. 2): Fanning Springs (median discharge equal to 110 ft^3/s), Little Fanning Springs (median discharge equal to 20 ft^3/s), and Manatee Springs (median discharge equal to 210 ft^3/s). The concentration of spring flow along this reach, although high, is smaller than that occurring in the previously described reaches of the Upper Ichetucknee where the median spring-flow contribution was about 100 ft^3/s per mile of river, and in reach 1 of the Santa Fe River where the median spring-flow contribution was about 95 ft^3/s per mile of river.

Spring discharge along other reaches of the Lower Suwannee and Lower Santa Fe Rivers was lower than the above river reaches. The spring-flow contributions to these remaining reaches ranged from a minimum of 3.6 ft^3/s per mile of river in the reach of the Suwannee River between Branford and the mouth of the Santa Fe River, to a maximum of 16 ft^3/s per mile of river in the reach of the Suwannee River between Hart Springs (near river mile 43) and the gaging station near Wilcox (fig. 2).

Data collected during several periods of relatively steady streamflow also indicated that the reach of the Lower Suwannee River between the Branford gaging station and the point just upstream from Rock Bluff Springs (near the Bell gaging station, fig. 2) received little net ground-water outflow (excluding the contributions from the mouth of the Santa Fe River). The streamflow pickup (excluding the Santa Fe River) along this 20-mi reach ranged from -166 to 90 ft^3/s during several periods (August 1996, August 1998, August 1999, and September 1999) of relatively steady streamflow. Streamflow pickup in the segments of this reach upstream and downstream from the Santa Fe River was also negligible or small during these periods. The measured change in streamflow in the reach of the Suwannee River between the Branford gaging station and the point just upstream from the mouth of the Santa Fe River generally ranged from -50 to 50 ft^3/s, and the pickup in the reach downstream from the mouth from the Santa Fe River to just upstream from Rock Bluff Springs ranged from 0 to 135 ft^3/s.

Evaluation of Ground-Water and Surface-Water Exchanges Using a Hydrologic Model

Interactions between ground and surface water were evaluated using a hydrologic model that simulates ground-water flows and levels in the Upper Floridan aquifer, streamflows and levels in the Lower Suwannee and Lower Santa Fe Rivers, and the exchange of water between the aquifer and two river reaches. The hydrologic model consists of separate aquifer and river submodels that have been linked through the river leakage terms (in the conservation of mass equations of the submodels) to form a coupled ground- and surface-water flow model. The hydrologic model gives additional insight into how the Upper Floridan aquifer and Lower Suwannee and Lower Santa Fe Rivers interact and how this interaction might be affected by changes within the system and at the system boundaries. The subsequent sections describe the: (1) model construction and calibration of the coupled model; (2) model evaluation of ground- and surface-water exchanges during water years 1998-99; (3) model evaluations of the effects of various hypothetical water-withdrawal scenarios on streamflow in the Lower Suwannee River; and (4) limitations of the model.

Model Construction

The hydrologic models of ground- and surface-water flow were developed in two phases. Each model was first constructed by translating a conceptual model of a particular aspect of the flow system into a mathematical model. Following construction, the models were calibrated by adjusting model parameters within the range of uncertainty until the best possible agreement was obtained between measured and simulated water levels and flows.

The subsequent sections describe the construction of the models of ground-water flow, surface-water flow, and ground- and surface-water interactions in the subregional study area. Construction of the models occurred in three phases: (1) construction of a ground-water flow model of the Upper Floridan aquifer; (2) construction of a surface-water flow model of the Lower Suwannee and Lower Santa Fe Rivers; and (3) linking the two models to construct a coupled ground-water and surface-water flow model.

Ground-Water Flow Model

An uncoupled, transient (unsteady-state), ground-water flow model was developed during this study. The model area is coincident with the subregional study area (fig. 1) and includes: (1) areas where the Upper Floridan aquifer contributes ground water to the Lower Suwannee and Lower Santa Fe Rivers; and (2) areas to the north and south of the Suwannee River where the aquifer contributes ground water to the Gulf of Mexico and to streams draining into the Gulf of Mexico (fig. 39). The model was used to simulate transient

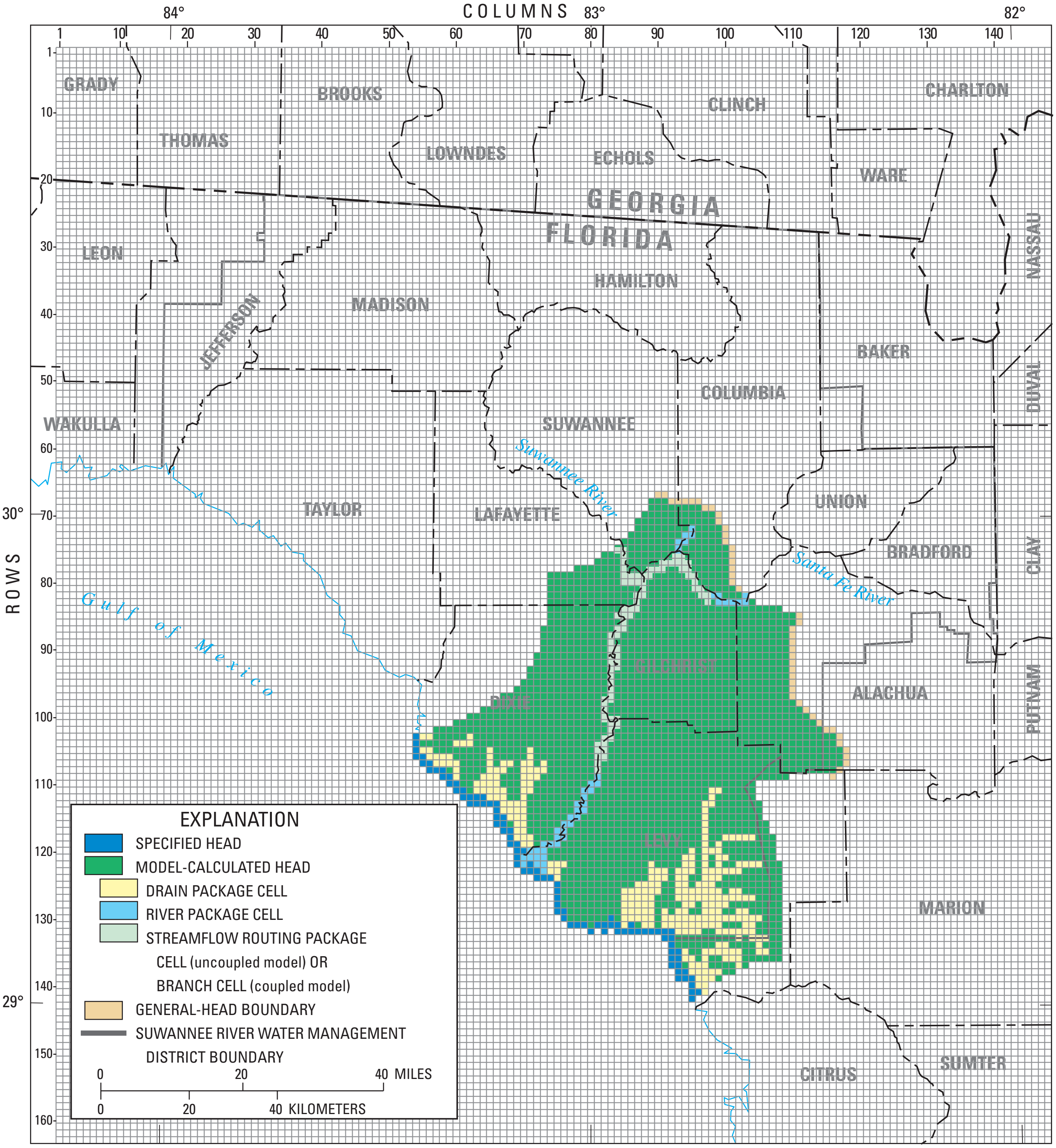


Figure 39. Model grid and lateral boundary conditions for the subregional ground-water flow model.

ground-water flows and levels for each day during water years 1998-99 and was developed to serve as the ground-water flow component (submodel) of the coupled model. The uncoupled model is identical to the ground-water flow submodel present in the coupled model, with the exception of the method used to represent the exchange of water between the rivers and the Upper Floridan aquifer. In the uncoupled version, water levels in the rivers were specified as input to the model. In the coupled version, water levels in the rivers were computed and output by the model.

Mathematical Basis and Approach

The MODFLOW computer program (McDonald and Harbaugh, 1988; Harbaugh and McDonald, 1996) was used to simulate ground-water flow within the Upper Floridan aquifer in the subregional study area. The program uses a finite-difference scheme to integrate the partial differential equation for three-dimensional, saturated ground-water flow under equilibrium (steady-state) or nonequilibrium (transient) conditions (McDonald and Harbaugh, 1988):

$$\frac{\partial}{\partial x}\left(K_{xx}\frac{\partial h}{\partial x}\right) + \frac{\partial}{\partial y}\left(K_{yy}\frac{\partial h}{\partial y}\right) + \frac{\partial}{\partial z}\left(K_{zz}\frac{\partial h}{\partial z}\right) - W = S_s\frac{\partial h}{\partial t}, \quad (6)$$

where

K_{xx}, K_{yy} and K_{zz} are hydraulic conductivity values along the x-, y-, and z-coordinate axes, in feet per day (which are assumed parallel to the major axes of hydraulic conductivity);

h is the hydraulic head at some point in the system, in feet;

W is a volumetric flux per unit volume of aquifer and represents sources or sinks of water (for example, wells, recharge, and leakage with rivers), in feet^3/(feet^3*day);

S_s is the specific storage of the aquifer at some point in the system, in feet^{-1}; and

t is time, in days.

In this study, vertical flow within the Upper Floridan aquifer is assumed to be negligible in comparison to horizontal flow. Accordingly, ground-water flow was simulated using a two-dimensional (vertically averaged) representation of ground-water flow in which the term $\frac{\partial}{\partial z}\left(K_{zz}\frac{\partial h}{\partial z}\right)$ is dropped from equation 6. The resulting system is also assumed to be isotropic ($K_{xx} = K_{yy}$) at a given point in the system) for lack of more detailed data indicating otherwise.

The exchange of water between rivers and the Upper Floridan aquifer is calculated as follows:

$$Q = C_{riv}\,(Z - h), \quad (7)$$

where

Q is the flow between the river and aquifer, in cubic feet per day;

C_{riv} is the conductance of the connection between the river and aquifer, in feet squared per day;

Z is the water-surface elevation (stage) of the river, in feet; and

h is the hydraulic head (ground-water level) in the aquifer beneath the river, in feet.

Note that if $Z > h$ then $Q > 0$ and the simulated flow direction will be from the river to the aquifer; if $Z < h$, then $Q < 0$ and the simulated flow direction will be from the aquifer to the river. For a given aquifer cell that is adjacent to a river, the value of Q calculated in equation 7 is added to all other sources and sinks (for example, well pumping or recharge) active in the cell, and this summation is used as the value of W in equation 6.

The ground-water flow model is constructed by: (1) using a grid to discretize the aquifer into many small units or cells; (2) formulating a ground-water flow equation for each cell with appropriate values of K, W, and S; and (3) specifying the appropriate values of flow, head, or head-dependent flows in cells along the model boundaries. These three steps result in a system of equations that is solved using the “preconditioned conjugate-gradient” method (Hill, 1990), which yields an estimate of hydraulic head and flow at every cell in the model.

Discretization and Model Boundaries

The ground-water flow model developed in this study employs a rectangular grid defined by 163 rows and 148 columns (fig. 39), with a uniform spacing of 5,000 ft for the width of each row and column. One model layer (McDonald and Harbaugh, 1988, p. 2-3) is used in the model, so the model simulates two-dimensional, areal (but not vertical) movement of ground water in the Upper Floridan aquifer.

The boundary of the ground-water model along the Gulf of Mexico is defined by model cells (referred to herein as “coastal boundary” cells) that are in the Gulf and contiguous with the Florida coastline in the subregional study area (fig. 39). The ground-water level of each coastal boundary cell is fixed at a value 0 ft (which was approximately equal to NGVD 1929 in this area during the study period) for each model time step. Discharge from the Upper Floridan aquifer to the Gulf of Mexico is represented by the simulated flow between the coastal boundary cells (in which the ground-water level is fixed) and adjacent cells on the landward side of the coastline (in which ground-water levels are calculated by the model). This type of boundary was chosen so that discharge from the Upper Floridan aquifer could occur only in the near-shore areas. This is consistent with the previous discussion, indicating that the westward limit of the freshwater-saltwater interface (which represents a no-flow boundary for the ground-water flow system) is probably within 1 mi of the Gulf shoreline.

The northern and western boundaries of the model represent no-flow boundary conditions. The northern no-flow boundaries of the ground-water flow model are formed by ground-water flow lines that terminate in the Suwannee River at Branford (figs. 2 and 23) and originate in upgradient areas to the west and east. The western part of this northern boundary originates along the axis of the potentiometric high west of the Suwannee River in Lafayette County. At this point, the western, no-flow model boundary is formed by a ground-water flow line that follows the gradient of the potentiometric surface of the Upper Floridan aquifer southwest through Dixie County to the coastline of the Gulf of Mexico.

The eastern part of the northern no-flow boundary of the model is drawn by beginning on the Suwannee River in Branford (figs. 2 and 23) and following the reverse (upgradient) path of the flow line to a point that approximately coincides with the limits of the unconfined part of the Upper Floridan aquifer. At this point, a head-dependent flux or general-head boundary (McDonald and Harbaugh, 1988, p. 11-1) is established that is approximately coincident with, or just west of, the boundary between parts of the Upper Floridan aquifer that are unconfined and parts that are overlain by the intermediate confining unit (fig. 14). General-head boundaries function by calculating flow across the model boundary that is proportional to the difference between the model-calculated ground-water level in the model cell adjacent to the boundary and a user-specified ground-water level at another point outside of the model boundary. These user-specified ground-water levels were calculated by interpolating (in space and time) from a series of seven potentiometric surface maps of the Upper Floridan aquifer that represent points in time that defined the general shape of the ground-water level time series along this boundary during water years 1998-99. The actual flux of water is calculated by MODFLOW, which: (1) simulates the ground-water level at a boundary cell; (2) calculates the water-level difference between the active boundary cell and the user-specified water level at the upgradient cell outside of the model boundary; and (3) multiplies the water-level difference by the conductance of the Upper Floridan aquifer along the flow path between the two points. These conductance values were set equal to the transmissivity along the flow path multiplied by 5,000 ft (the width of one face of a model cell) and divided by the length of the flow path (in this case, the width of one face of a model cell, also 5,000 ft).

The general-head boundary transitions to a no-flow boundary near the intersection of the boundaries between Alachua, Levy, and Marion Counties (fig. 39). The no-flow boundary extends south to the highest point on the potentiometric "high" west of Rainbow Springs. At this point, the no-flow boundary is located along the flow path directed downgradient from the potentiometric high, terminating at the boundary along the coastline of the Gulf of Mexico.

Representing Sources and Sinks of Water

The sources and sinks used in the ground-water flow model are identical to those previously described in the "Ground-Water Hydrology" section. Various MODFLOW packages (McDonald and Harbaugh, 1988, p. 1-1) were used to represent recharge, well pumpage, and exchanges between the Upper Floridan aquifer and rivers, springs, and wetlands.

Recharge was represented in the ground-water flow model using the MODFLOW Recharge Package (McDonald and Harbaugh, 1988, p. 7-1). Recharge values were estimated for each cell using rainfall and pan evaporation data and the daily SMA algorithm (E.P. Weeks, U.S. Geological Survey, written commun., 1993) as previously described.

Ground-water withdrawals from wells were represented using the Well Package (McDonald and Harbaugh, 1988, p. 8-1). Monthly ground-water withdrawals from October 1997 to September 1999 were estimated as previously described. Daily values for each well were then determined by dividing the monthly ground-water withdrawal for a given month by the number of days in that month.

The River Package (McDonald and Harbaugh, 1988, p. 6-1) and Streamflow-Routing Packages (Prudic, 1989) were used to represent exchanges between the Upper Floridan aquifer and the major rivers in the uncoupled ground-water flow model. Estimates of the water level of the river and riverbed conductance are required for each cell in the model grid as input for these packages. The water level at each grid cell was estimated by interpolating between the water levels at gaging stations for each day from October 1997 to September 1999. The Drain Package (McDonald and Harbaugh, 1988, p. 6-1) was also used to estimate discharge from selected streams and springs in the Waccasassa River network of streams, as well as the wetlands connected to streams draining coastal areas adjacent to the Gulf of Mexico.

Representation of Hydraulic Properties

The Block-Centered Flow Package (McDonald and Harbaugh, 1988, p. 5-1) was used to represent hydraulic properties in the ground-water flow model. A confined layer type was specified and was considered appropriate for the subregional study area because the magnitude of ground-water level changes is generally small relative to the thickness of the aquifer (transmissivity changes little as water levels fluctuate). An alternative model was tested in which transmissivity was set equal to hydraulic conductivity, and then multiplied by the saturated thickness of the freshwater section of the Upper Floridan aquifer and surficial sediments; however, this model failed to improve the quality of the ground-water flow models. Geographic variations in transmissivity were represented by assigning different values of transmissivity to groups of cells (transmissivity zones) in different areas. For example, the cells under the potentiometric surface high in Gilchrist County were treated as a single transmissivity zone with a smaller value of transmissivity than in the surrounding areas, which were treated as separate transmissivity zones. Specific yield estimates were required for the transient subregional model and were assigned a value of 0.2.

Surface-Water Flow Model

An uncoupled model of open-channel flow in the Lower Suwannee and Lower Santa Fe Rivers was constructed as a preliminary step to speed the development of the final coupled model of the subregional study area. The coupled model requires more extensive computer time than the uncoupled model, and therefore, much more testing was possible during calibration than would have otherwise been possible. This open-channel flow model was not coupled with a ground-water flow model, and thus, is not capable of calculating exchanges between the rivers and the Upper Floridan aquifer. Accordingly, exchanges with the ground-water system were estimated for selected periods and provided as input to the model. The subsequent sections describe the mathematical basis of the model, how the two rivers were represented in the mathematical model (schematic and discretization), and how channel and floodplain hydraulic characteristics were represented in the model.

Mathematical Basis and Approach

The BRANCH mathematical model (Schaffranek and others, 1981; Schaffranek, 1987) was used to simulate open-channel flow in selected reaches of the Lower Suwannee and Lower Santa Fe Rivers. The program uses a finite-difference scheme to integrate the de Saint Venant equations (Cunge and others, 1980), which are equations of conservation of mass (continuity) and conservation of momentum that describe unsteady, one-dimensional flow in open channels. The continuity equation is:

$$B\frac{\partial z}{\partial t}+\frac{\partial Q}{\partial x}+q=0\,; \qquad (8)$$

where

B is the channel top width, in feet;

z is the water-surface elevation in the channel, in feet;

t is the elapsed time since the beginning of the simulation, in seconds;

Q is the discharge in the channel, in cubic feet per second;

x is the distance measured along the channel centerline, in feet; and

q is the lateral outflow from, or inflow to, the channel, in cubic feet per second per foot of river channel.

The conservation of momentum equation is:

$$\frac{1}{gA}\frac{\partial Q}{\partial t}+\frac{2BQ}{gA^2}\frac{\partial Q}{\partial x}+\frac{BQ^2}{gA^3}\frac{\partial A}{\partial x}+\frac{\partial Z}{\partial x}+\frac{k}{A^2R^{4/3}}Q|Q|-\frac{\xi B}{gA}U_a^2\cos(\alpha)=0 \qquad (9)$$

where

g is the acceleration of gravity, in feet per second squared;

A is the submerged cross-sectional area of the channel, in feet squared;

β is the momentum or Boussinesq coefficient (dimensionless), which compensates for the nonuniform velocity distribution in the channel;

k is the flow-resistance coefficient of the channel, in seconds^2/feet$^{2/3}$;

R is the hydraulic radius of the channel, in feet;

ξ is a dimensionless wind-resistance coefficient; and

U_a is the wind velocity, in feet per second, making an angle, α, with the positive x-axis.

The open-channel flow model is developed by: (1) defining the channel centerline and discretizing it into smaller segments; (2) discretizing the time to be simulated by the model into smaller increments of time (time steps); (3) specifying the values of flow or water levels at upstream and downstream boundaries at the beginning and end of each of these time steps; and (4) formulating the continuity and momentum equations by specifying appropriate values for the coefficients and independent variables in the continuity and flow equations.

Step four may also include specifying the flow at one or more internal locations of the model for each time step to simulate streamflow gains or losses resulting from ground- and surface-water exchanges or contributions from tributaries. These four steps result in a system of equations that are solved using a weighted, four-point, implicit-solution technique (Schaffranek and others, 1981; Schaffranek, 1987), which yields an estimate of water level and discharge for each time step at the ends of each segment in the model.

Schematic and Discretization

The streamflow model developed for this study simulates 53 mi of the Suwannee River between the gaging stations at Branford at river mile 76.1 (map index no. 5, fig. 11) and near Old Town at river mile 23.5 (map index no. 8, fig. 11), and simulates about 18 mi of the Santa Fe River between the gaging station near Fort White at river mile 18.3 (map index no. 14, fig. 11) and the mouth of the Santa Fe River. The model could not be extended to include the Suwannee River downstream from the gaging station near Old Town because a continuous set of stage data for water years 1998-99 was not available at any of the downstream gaging stations.

Discretization of the simulated reaches of the Suwannee and Santa Fe Rivers was accomplished by first overlaying a map of the grid from the previously described ground-water flow model on a map of the centerline of the Suwannee and Santa Fe Rivers. The rivers are divided into segments at the points of intersection between the river centerlines and the ground-water model grid. This step was necessary because the software that links the ground-water flow and streamflow models requires that each river segment be assigned to only one aquifer model grid cell (Swain and Wexler, 1996). More segments were added in some instances to minimize the differences in length between the longest and shortest segments in the overall network of segments. Other segments were added so that points of interest (such as the confluence between the Suwannee and Santa Fe Rivers, gaging stations, and other flow measurement locations) within the network would coincide with the end of a segment. This approach ensures that simulated values of streamflow and water levels would be available for comparison with points where these values have been observed or estimated from measurements in the field (the ends of segments are locations where the model simulates river levels, flow, average velocity, channel top width, and submerged cross-sectional areas). This process of spatial discretization resulted in a network of 278 river segments that range from 480 to 1,900 ft in length. Implementing the BRANCH model also requires grouping contiguous segments into "branches" to improve the efficiency of calculations (Schaffranek and others, 1981). Accordingly, the 278 river segments were grouped into 71 branches, with each containing 2 to 4 segments.

Several time periods were simulated during development of the model. These time periods were generally about 1 month in duration for the uncoupled simulations that were used to refine the initial estimates of the channel roughness values and cross-sectional geometry of the segments in the model. After the calibration phase of the uncoupled model was completed, the BRANCH (river) model was coupled with the MODFLOW (aquifer) model and simulations were made from October 1, 1997, to September 30, 1999. For both the uncoupled and coupled models, the simulation periods were discretized into 15-minute time intervals. Boundary-value data were obtained from hourly streamflow estimates at the gaging stations on the Suwannee River at Branford and the Santa Fe River near Fort White, and from 15-minute interval stage measurements taken at the gaging station on the Suwannee River near Old Town. The necessary 15-minute interval data for the upstream flow boundaries were obtained by interpolating between the hourly flow data obtained from the Branford and Fort White gaging stations.

Hydraulic Characteristics of the River Channel and Floodplain

The final step in developing the uncoupled, open-channel flow model was defining the geometry and roughness of the river channel and floodplain. As previously mentioned, most of the river channel and floodplain geometry and roughness data for the present study were obtained from digital (computer) input files that were used to develop a water-surface profile model for an earlier flood study conducted by the U.S. Army Corps of Engineers (U.S. Army Corps of Engineers, 1989; John Good, Suwannee River Water Management District, written commun., 1996). Some differences probably exist between the digital input files obtained from the U.S. Army Corps of Engineers (USACE) for use in the present study and the input files that were used in the final version of the water-surface profile model from the earlier flood study by the USACE. For example, numerous fields in the input files obtained from the USACE have erroneous values of Manning's roughness coefficient (equal to 99), which indicates that these files were probably working "templates" for input files used to calibrate the water-surface profile model. For this reason, the digital files obtained from the USACE were used only as a starting point for calibrating the uncoupled, open-channel flow model in the present study. Other channel geometry data, including field measurements (taken during the present study) and discharge measurements notes (containing depth and water-level data), were also obtained for a few sites.

Both sources of data were reformatted for input to the FEQUTL computer program (Franz and Melching, 1997). This program uses the geometry and roughness data to generate hydraulic tables that specify values of the hydraulic properties (such as depth, area, top width, and conveyance) at different water-surface elevations at individual cross sections along the river system. A hydraulic table was generated for each cross section where channel and floodplain geometry data were available. Hydraulic tables for the cross sections at the ends of the river segments used in the BRANCH model were then generated by interpolating between the sections with geometry and roughness data using the FEQUTL XSINTERP command.

Coupled Ground-Water and Surface-Water Flow Model

The final phase of the model construction process was to construct the coupled model to simulate ground-water flow in the Upper Floridan aquifer, open-channel flow in selected reaches of the Lower Suwannee and Lower Santa Fe Rivers, and the exchange of ground water and surface water between the Upper Floridan aquifer and both rivers. Development of the coupled model makes it possible to simultaneously simulate changes in river stage and flow and ground-water levels and flow in response to changing stresses on the river-aquifer system.

The MODBRANCH computer program (Swain and Wexler, 1996) was used to simulate water levels and flows in the Lower Suwannee River, Lower Santa Fe River, and contiguous Upper Floridan aquifer. This program also simulates the dynamic exchange of water between these two rivers and the aquifer, using Darcy's law to compute the leakage (eq. 8) between the river channel and aquifer (Swain and Wexler, 1996, p. 4):

$$q = \frac{K'}{b'} B(Z - h) \; , \qquad (10)$$

where

$\frac{K'}{b'}$ is the leakage coefficient of the river-aquifer connection, in days^{-1} (the hydraulic conductivity, K', in feet per day, divided by the flow path length, b' , in feet);

B is the river topwidth, in feet;

Z is the river stage, in feet; and

h is the hydraulic head (ground-water level) in the aquifer adjacent to the river, in feet.

MODBRANCH consists of modified versions of MODFLOW (McDonald and Harbaugh, 1988; Harbaugh and McDonald, 1996) and BRANCH (Schaffranek and others, 1981). These modifications: (1) contain mechanisms for using MODFLOW-calculated ground-water heads in equation 10; (2) incorporate results of equation 10 into the continuity equation (eq. 8) in BRANCH and the ground-water flow equation (eq. 6) in MODFLOW; and (3) account for the different time-step lengths that typically are used when applying MODFLOW and BRANCH, especially with regard to maintaining consistency in the mass balances calculated by each of the two submodels (Swain and Wexler, 1996, p. 2). Accommodating different time-step lengths was necessary because changes in streamflow typically occur over minutes and hours, whereas changes in aquifer flow typically occur over minutes, days, or months (Swain and Wexler, 1996). Therefore, MODBRANCH allows multiple BRANCH time steps to occur during a given MODFLOW time step or stress period (Swain and Wexler, 1996). This is accomplished by: (1) running MODFLOW for one stress period; (2) interpolating the ground-water heads for the starting and ending times of each BRANCH time step; (3) formulating and solving the BRANCH model equations for each time step; (4) averaging the leakage rates calculated for all river segments in each MODFLOW aquifer cell over the MODFLOW stress period; and (5) passing these average leakage rates back to MODFLOW. Consistent mass-balance results are obtained by iterating between MODFLOW and BRANCH (repeating steps 1 to 5) until the simulated water levels in the river and aquifer have converged to stable values.

Note that equation 10 is analogous to equation 7 used in the uncoupled model. The primary difference is that river stage, Z, is simulated (calculated) by the model in equation 10, unlike equation 7, which assumes that the stage is known. Therefore, the coupled model makes it possible to evaluate changes in river stage that result from changes in recharge rates, ground-water pumping, or rates of upstream runoff into the simulated river system.

Model Calibration

Calibration of the simulation models was performed in two steps. In the first step, the uncoupled subregional ground-water flow model and the uncoupled surface-water flow model were calibrated independently. In the second step, the uncoupled ground- and surface-water flow models that were calibrated in step one were linked to form a coupled model of ground water and surface water.

Uncoupled Ground-Water Flow Model

Calibration of the uncoupled ground-water flow model was accomplished primarily by adjusting the conductive properties used in the model in an effort to optimize the correspondence of measured and simulated ground-water levels and streamflows at locations where measured data were available (figs. 4 and 27). All of these conductive properties depend on the hydraulic conductivity and thickness of different parts of the aquifer. In the uncoupled ground-water flow model constructed for the present study, the conductive properties include aquifer transmissivity, riverbed conductance, and general-head boundary conductance (McDonald and Harbaugh, 1988; Harbaugh and McDonald, 1996).

Most of the ground-water model calibration effort was devoted to determining the values of aquifer transmissivity parameters, which are areas (zones) with unique values of transmissivity. Early versions of the uncoupled ground-water flow model attempted to simulate the ground-water flow system using only two parameter values: (1) a low transmissivity value used in two zones corresponding to the potentiometric high areas under the Waccasassa Flats (in Gilchrist County) and in southern Levy County; and (2) a higher transmissivity value in the rest of the subregional study area. These versions yielded promising model fits in the low transmissivity zones but poor fits in the remaining area. Subsequent efforts attempted to improve the model fit by subdividing this latter area into a few additional zones, using the conceptual model and discrepancies between simulated and measured ground-water levels as a guide for delineating the boundaries of the new zones. For example, zones were added in areas along the Gulf Coast, in areas paralleling the rivers, and areas east of the Waccasassa Flats, where data indicated that the transmissivity values were probably higher than in adjacent areas.

A map of the transmissivity zones and the values used in the final version of the uncoupled ground-water flow model is shown in figure 40. Values for transmissivity and other parameters are given in tables 4 and 5. The values were obtained using a least-squares regression approach (Hill, 1998; Doherty, 2000) which was implemented using the parameter estimation (PEST) optimization software (Doherty, 2000). No attempt was made to constrain the value that could be obtained from the optimization software. Observation weights were calculated using the methods described by Hill (1992; 1998).

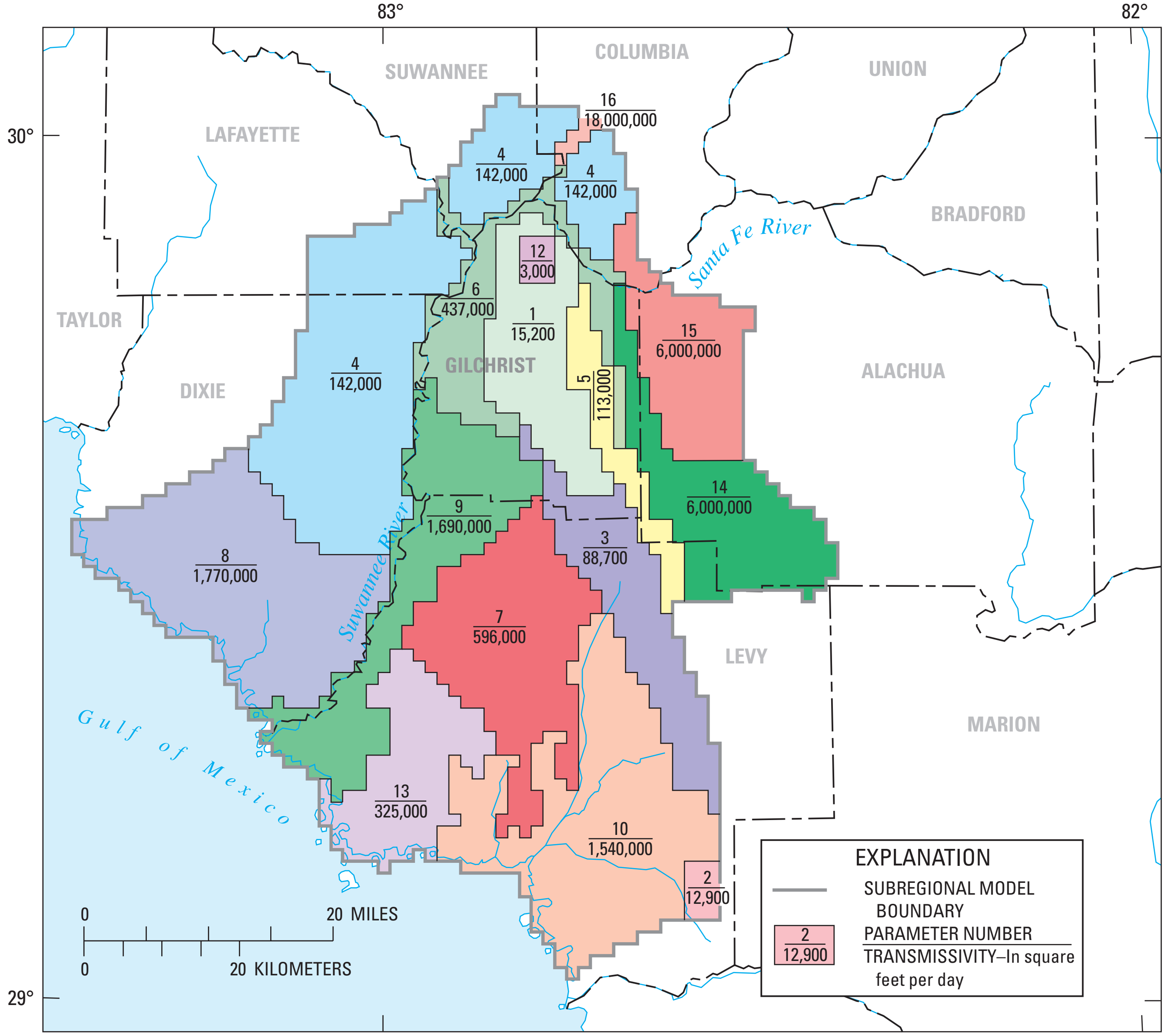


Figure 40. Calibrated transmissivity distribution from the uncoupled ground-water flow model.

For transmissivity parameters 1 to 10, 95-percent confidence intervals are given in table 4 and composite and composite-scaled sensitivities are given in table 6. Transmissivity parameters 12 to 16 do not have confidence intervals or sensitivities because their values were fixed during the latter stages of model calibration when the model became insensitive to these parameters and the regression began to have problems converging. The values used for these "fixed parameters" are consistent with the values obtained in previous calibrations when the parameters were estimated using regression analysis. The parameter value and confidence interval for parameter 11 (the multiplier used to estimate riverbed conductance values as part of the river and streamflow routing packages) are given in table 4, and composite sensitivities for this parameter are given in table 6.

Confidence intervals for each of the model parameters were relatively tight, indicating that the estimated values were well resolved (table 4). This inference, however, should be taken with some caution because confidence intervals were estimated using a linearity assumption that may not be valid for the full width of the confidence limits (Doherty, 2000, p. 5-21). Moreover, the parameter values depend on the delineation of the parameter zone boundaries, which are also subject to uncertainty. Because the parameter confidence intervals were relatively narrow, correlations between parameters did not seem to be a problem. The absolute values of the correlation coefficients were less than 0.95 for all of the possible combinations of parameters. The highest correlations were -0.85 between transmissivity parameters 2 and 10, -0.90 between parameters 4 and 8, -0.84 between parameters 5 ad 6, and -0.82 between parameters 3 and 10.

Table 4. Estimated parameter values and 95-percent confidence limits from the uncoupled ground-water flow model.

[Parameter numbers for transmissivity parameters correspond to the map index numbers for the transmissivity zones shown in fig. 41. Parameter label refers to the parameter names that are used in the computer input and output files for the Parameter Estimation (PEST) program (Doherty, 2000). Transmissivity parameter values and confidence limits are reported in units of feet squared per day, and the riverbed conductance multiplier parameter value and confidence limits are reported in units of feet per day]

Parameter type	Parameter number	Parameter label	Estimated value	95-percent confidence limits	
				Lower limit	Upper limit
Transmissivity	1	t2e4w	15,206.9	14,866.9	15,554.6
Transmissivity	2	t2e4t	12,885	12,763.8	13,007.4
Transmissivity	3	t6e4w	88,709.6	87,082.8	90,366.8
Transmissivity	4	t6e4i	142,058	139,122	145,055
Transmissivity	5	t1e5w	112,513	109,412	115,700
Transmissivity	6	t5e5	436,512	366,522	519,867
Transmissivity	7	t5e5c	596,010	575,485	617,268
Transmissivity	8	t5e5wc	1.767649×10^6	1.657172×10^6	1.885491×10^6
Transmissivity	9	t1e6r	1.692817×10^6	1.645372×10^6	1.741630×10^6
Transmissivity	10	t2e6w	1.544231×10^6	1.487747×10^6	1.602860×10^6
Riverbed conductance multiplier	11	srpc1	807.971	644.570	1,012.79

Table 5. Specified parameter values from the uncoupled ground-water flow model.

[Parameter numbers for transmissivity parameters correspond to the map index numbers for the transmissivity zones shown in fig. 41. Parameter label refers to the parameter names that are used in the computer input and output files for the Parameter Estimation (PEST) program (Doherty, 2000). Transmissivity parameter values are reported in units of feet squared per day. General-head boundary and riverbed conductance values are reported in units of feet squared per day. Specific yield values are dimensionless]

Parameter type	Parameter number	Parameter label	Fixed value
Transmissivity	12	t5e3	3,000
Transmissivity	13	t1e5c	325,000
Transmissivity	14	t2e6	6.0×10^6
Transmissivity	15	t5e6	6.0×10^6
Transmissivity (of Ichetucknee Trace, upgradient from Ichetucknee River Headsprings)	16	tiche	1.8×10^6
Conductance of general-head boundary	17	ghb2	6.0×10^6
Conductance of general-head boundary	18	ghb3	6.0×10^6
Riverbed conductance	19	riv1	2.0×10^6
Riverbed conductance (upper reach of the Ichetucknee River that contains the Ichetucknee Springs Group)	20	rivich	8.5×10^6
Specific yield	21	sy1	0.2

Table 6. Values of composite and composite-scaled parameter sensitivities for the uncoupled ground-water flow model.

[Sensitivities measure the amount of change in the model input for a given change in the value of a parameter (see fig. 27 for well locations). Parameter numbers for transmissivity parameters correspond to the map index numbers for the transmissivity zones shown in fig. 40. Parameter label refers to the parameter names that are used in the computer input and output files for the Parameter Estimation (PEST) program (Doherty, 2000). Composite-scaled sensitivities indicate the total amount of information provided by the observations for a given parameter (Hill, 1998). Composite sensitivities are equal to the composite-scaled sensitivity divided by the $\log_{10}$-transformed parameter value. Transmissivity parameter values are reported in units of feet squared per day. The riverbed conductance multiplier value is reported in units of feet per day. Composite sensitivity values are reported in units of $\log_{10}$ (feet squared per day) for transmissivity parameters, and in units of $\log_{10}$ (feet squared per day) for the riverbed conductance parameter. All composite-scaled sensitivity values are dimensionless]

Parameter type	Parameter number	Parameter label	Value	Composite sensitivity	Composite-scaled sensitivity
Transmissivity	1	t2e4w	15,206.9	6.4	27
Transmissivity	2	t2e4t	12,885	24	99
Transmissivity	3	t6e4w	88,709.6	12.	59
Transmissivity	4	t6e4i	142,058	15	79
Transmissivity	5	t1e5w	112,513	8.2	41
Transmissivity	6	t5e5	436,512	1.6	9.1
Transmissivity	7	t5e5c	596,010	4.8	28
Transmissivity	8	t5e5wc	1.767649×10^6	4.4	27
Transmissivity	9	t1e6r	1.692817×10^6	5.9	37
Transmissivity	10	t2e6w	1.544231×10^6	8.8	55
Riverbed conductance multiplier	11	srpc1	807.971	0.9	2.5

The model was most sensitive to transmissivity parameters 2 (composite sensitivity of 24) and 4 (composite sensitivity of 15), and least sensitive to the riverbed conductance parameter 11 (composite sensitivity of 0.9) and transmissivity parameter 6 (composite sensitivity of 1.6) as noted in table 6. Generally, the sensitivity of the model to a given parameter was a function of the number of observations in that zone and the proximity of a head-dependent flux or constant-head boundaries. For example, the most sensitive transmissivity parameters contained either numerous observation wells or a few wells with a high frequency of data collection. The least sensitive transmissivity parameters, such as parameter 6, contained only a few wells with limited data and were close to the rivers or the eastern general-head boundary. Higher sensitivities result in more accurate parameter estimates, and lower sensitivities result in less accurate parameter estimates; however, even the least sensitive parameters were sensitive enough to have well-resolved values.

Analysis of the model residuals indicates relatively good correspondence between the simulated and measured ground-water levels and streamflows. The average ground-water level residual was 0.2 ft, and the 10th, 25th, 50th, 75th, and 90th percentiles of the residuals were -3.0, -1.5, 0.1, 1.8, and 3.6, respectively. The minimum ground-water level residual was -24 ft and the maximum residual was 11 ft. The estimation of this minimum residual could be improved by increasing the transmissivity of part of the northern Waccasassa Flats potentiometric high surrounding well s291 (figs. 18 and 27). This was not done, however, because only one head observation was available at well s291.

Time series of measured and simulated ground-water levels are shown for selected wells near the Lower Suwannee River (fig. 41). The time series of measured and simulated ground-water levels in other wells are essentially the same as those in the coupled model, which is discussed later.

The spatial distribution of ground-water level residuals during high and low water-level periods are shown in figures 42 and 43, respectively. Spatial trends are largely absent from the data, with the possible exception of an area in central Levy County during a high water-level period (fig. 42). During this period, the simulated ground-water levels in a cluster of wells in central Levy County are 2 to 9 ft higher than measured ground-water levels. The higher levels in this area seem to be an artifact of the initial conditions used in the uncoupled model, and are not present in the final coupled model.

Time series of measured and simulated daily mean streamflows in the Suwannee River near Wilcox are shown in figure 44. Streamflow was simulated relatively well, reproducing the pattern

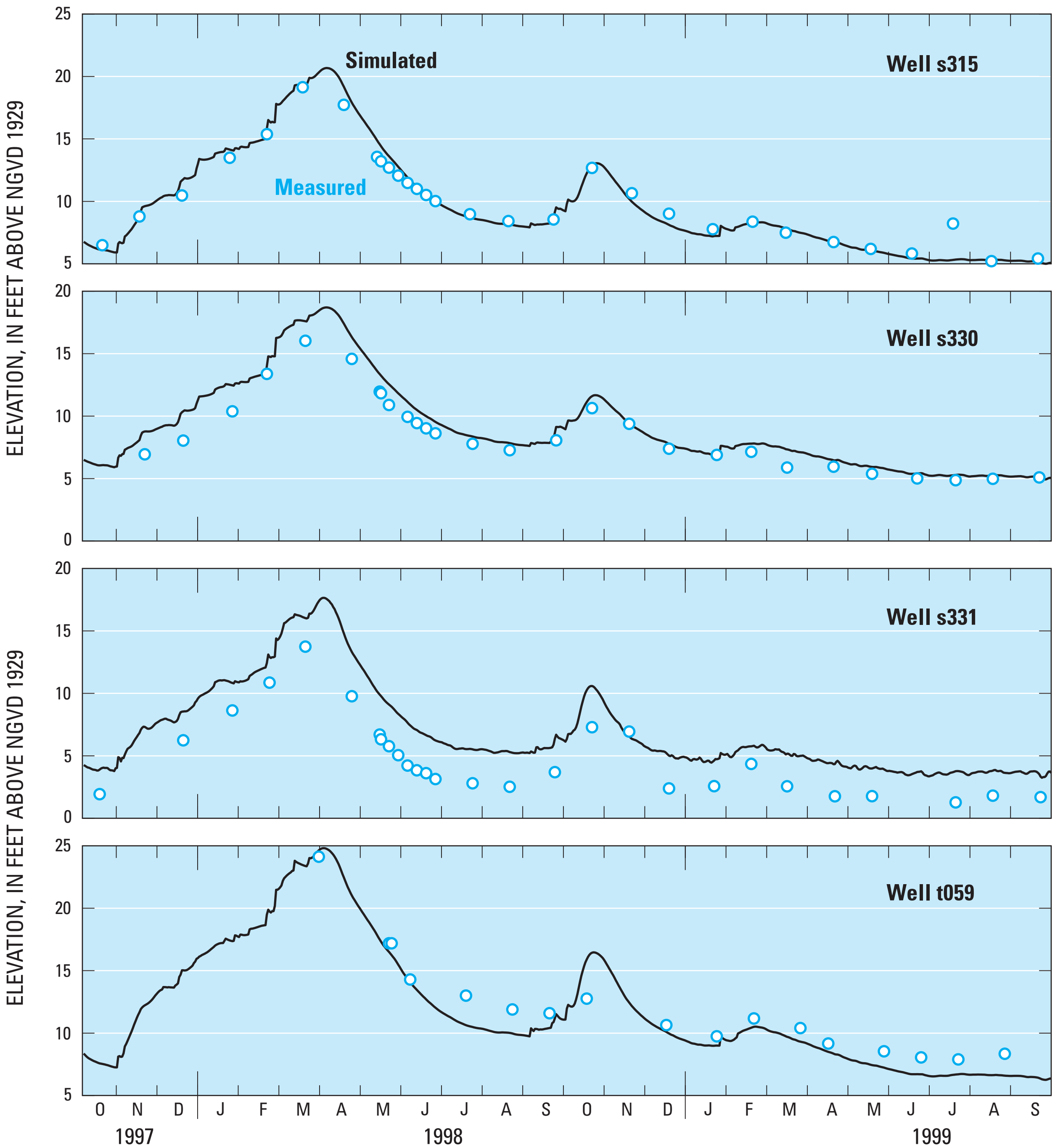


Figure 41. Time series of measured ground-water levels and ground-water levels that were simulated with the ground-water flow model, October 1997 to September 1999 (location of wells s315, s330, s331, and t059 are shown in fig. 27).

of fluctuations in measured streamflow in this reach. The mean residual flow in this reach was about 300 ft^3/s, which represents about 3 percent of the average annual flow and about 20 percent of the average annual ground-water discharge to this reach. The mean residual of the simulated flows in the Santa Fe River upstream from the gaging station near Fort White (map index no. 14, fig. 11) was 150 ft^3/s, which represents about 20 percent of the mean ground-water discharge to this reach.

Uncoupled Surface-Water Flow Model

Calibration of the uncoupled model of surface-water flow in the Lower Suwannee and Lower Santa Fe Rivers primarily consisted of adjusting the channel and floodplain roughness parameter. This parameter was adjusted until the simulated river water levels and flows matched those that were measured in May and August 1997, August 1998, and September 1999.

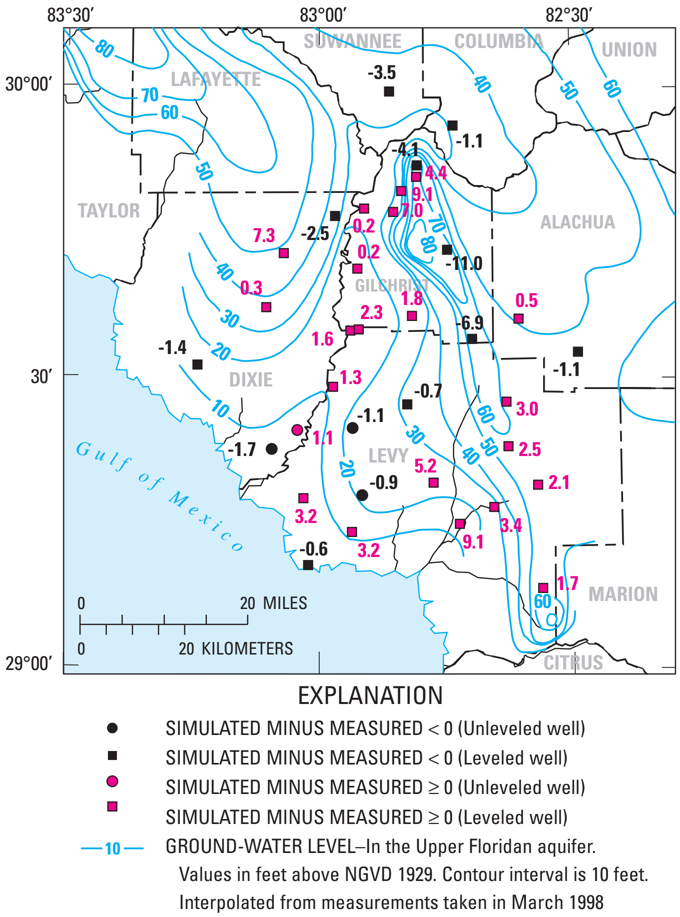


Figure 42. Spatial distribution of ground-water level residuals from the ground-water flow model during a high water-level period.

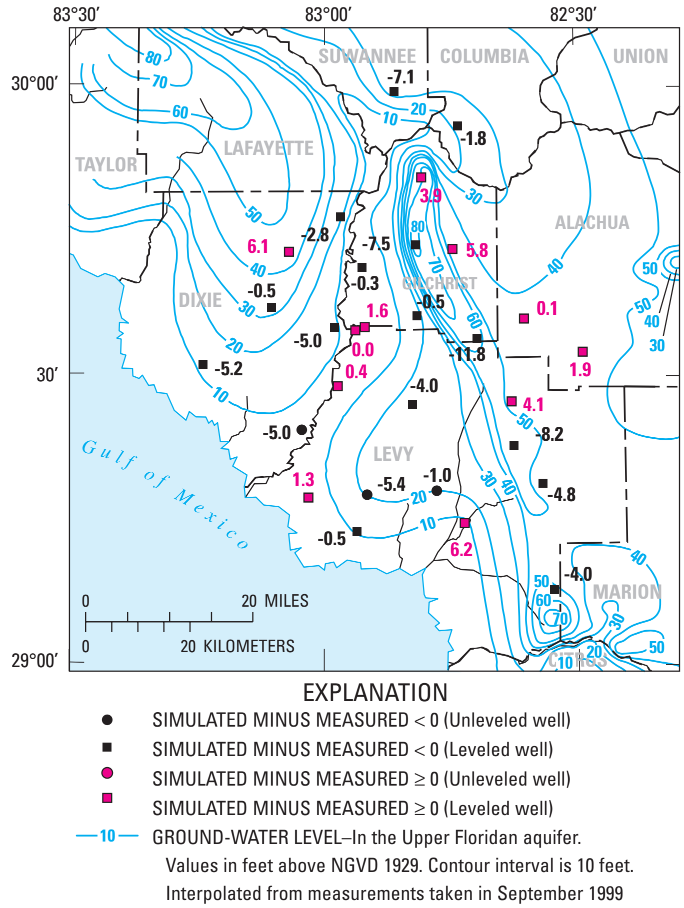


Figure 43. Spatial distribution of ground-water level residuals from the ground-water flow model during a low water-level period.

Flow measurements during these periods were obtained using acoustic Doppler current profilers (ADCPs) that were operated nearly continuously for 10 to more than 48 hours at many of the measured sites. Data from these measurements were used to estimate ground-water inflow along the Lower Suwannee and Lower Santa Fe Rivers. These data were then used to represent ground-water inflow to both rivers in the uncoupled surface-water flow model, by specifying them as inflows to selected cross sections in the model.

Calibrated roughness values for the Suwannee River were similar to those in the model input files obtained from an earlier study by the U.S. Army Corps of Engineers (1989). The within-channel values of the roughness parameter, represented as Manning's *n* coefficient (Chow and others, 1988), varied from 0.027 to 0.0385 in the uncoupled model developed for the present study, compared with values of 0.030 to 0.035 in the USACE model input files. Overbank (floodplain) roughness values were identical for the two models, except for the reach of the Suwannee River upstream from river mile 60.31 (fig. 11), where the uncoupled model in this study had a value of 0.15, compared with the value of 0.20 in the USACE input files.

Differences between the models were slightly greater for the Lower Santa Fe River. Downstream from river mile 19 (fig. 11), the within-channel roughness value was 0.025 for the uncoupled model in this study compared with the value of 0.045 in the USACE model input files. Upstream from this reach, the within-channel roughness value was identical (0.035) in both models. The overbank roughness value was 0.15 (downstream and upstream) for the model used in this study, compared with values of 0.28 (downstream) and 0.20 (upstream) in the USACE model input files.

Some changes in the channel geometry data in the USAC model input files were needed during calibration of the uncoupled surface-water flow model to more accurately reflect field conditions and improve simulated river water levels and flows. For the Santa Fe River, most of the channel geometry changes were made in the reach between the gaging stations near Hildreth (map index no. 16, fig. 4) and at the mouth of the Ichetucknee River. These changes were made to reflect observations near the Hildreth gaging station, where the channel elevation was about 7 ft deeper than that indicated in the USACE model input files, and to improve the simulation of stage in this reach. Changes were also made: (1) upstream near river mile 13 (fig. 11) to reduce the influence of one

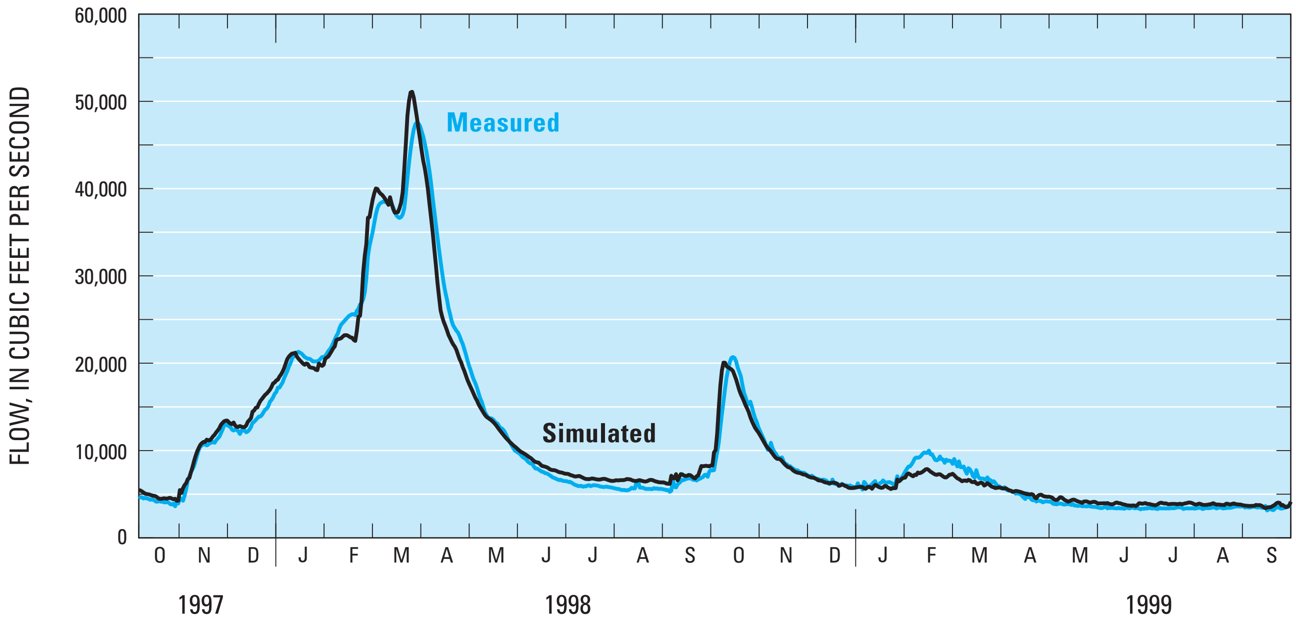


Figure 44. Time series of measured and simulated streamflows in the Suwannee River near Wilcox, Florida (map index no. 7, fig. 11). Simulated flow values are from the uncoupled ground-water flow model.

section that may not have represented the cross sections in this reach; and (2) near the low-water shoals just downstream from the Fort White gaging station (map index no. 14, fig. 11) based on depth measurements taken on the shoals during this study. For the Suwannee River, the channel geometry used in the uncoupled surface-water flow model was identical to that of the USACE model input files, except for a short reach near river mile 62 (fig. 11) where the deepest parts of the cross sections were lowered to reflect the deeper elevations of sections just upstream and downstream from this reach.

To simplify calibration of the uncoupled surface-water flow model, the Suwannee and Santa Fe Rivers were calibrated separately. In both instances, streamflow was specified as the upstream boundary condition and stage was specified as the downstream boundary condition. Additionally, the Suwannee River was calibrated in a “downstream” fashion by sequentially calibrating reaches between the Branford and Bell gaging stations, Branford and Wilcox gaging stations, and Branford and Old Town gaging stations (fig. 11 and table 1). Use of the Gopher River gaging station (map index no. 10, fig. 11) as a downstream boundary condition was not possible because the water-level dataset for this station contained long periods of missing data during the study.

Results from the calibration of the Santa Fe River segment of the model are presented in table 7. Simulated water-surface elevations (stages) were about 1.1 and 0.8 ft lower than measured stages at the Fort White gaging station (map index no. 14, fig. 11) at 08:00 local time on August 14, 1999, and September 27, 1999, respectively. Farther downstream at the gaging station near the mouth of the Ichetucknee River (map index no. 15, fig. 4), simulated stages were 0.2 ft lower and 0.6 ft higher than measured stages for the same time periods, respectively, mentioned above. The simulated stage was 0.2 ft higher than the measured stage at the gaging station near Hildreth (map index no. 16, fig. 4) during the September time period (measured data were unavailable at this location for the August time period). Flow measurements were not used to calibrate the model of the Santa Fe River because flow conditions were steady during the calibration periods, and gains and losses along the river were supplied as inputs to the model.

Table 7. Simulated and measured water-surface elevations in the Lower Santa Fe River for selected time periods.

[Locations are shown in fig. 4. Simulated values are from the uncoupled surface-water flow model; elevations in feet above NGVD 1929]

Location	Water-surface elevation			
	August 14, 1998		September 27, 1999	
	Simulated	Measured	Simulated	Measured
Santa Fe River near Fort White (station 02322500)	21.54	22.67	21.22	21.99
Santa Fe River at the mouth of the Ichetucknee River	10.11	10.30	8.51	7.92
Santa Fe River near Hildreth (station 02322570)	9.72	Missing	7.54	7.34

For the final uncoupled model of the Suwannee River between the Branford and Old Town gaging stations, the difference between simulated and measured stages ranged from less than 0.1 to 3.8 ft for the range of conditions used in the calibration effort. Under low-water conditions on September 27, 1999, simulated stages were generally within 0.1 ft of measured stages at the Branford gaging station, about 1 ft lower than measured stages at the gaging station near Bell, and within 0.1 ft at the gaging station near Wilcox; some of the lowest measured stages were erroneous (too low) because of clogged intakes at the gaging station (fig. 45). Under average streamflow conditions during August 18-19, 1997, simulated stages were generally within 0.1 ft of measured stages at the Branford gaging station, about 1.7 ft lower than measured stages at the gaging station near Bell, and about 0.2 to 0.3 ft higher at the gaging station near Wilcox (fig. 46). Under high

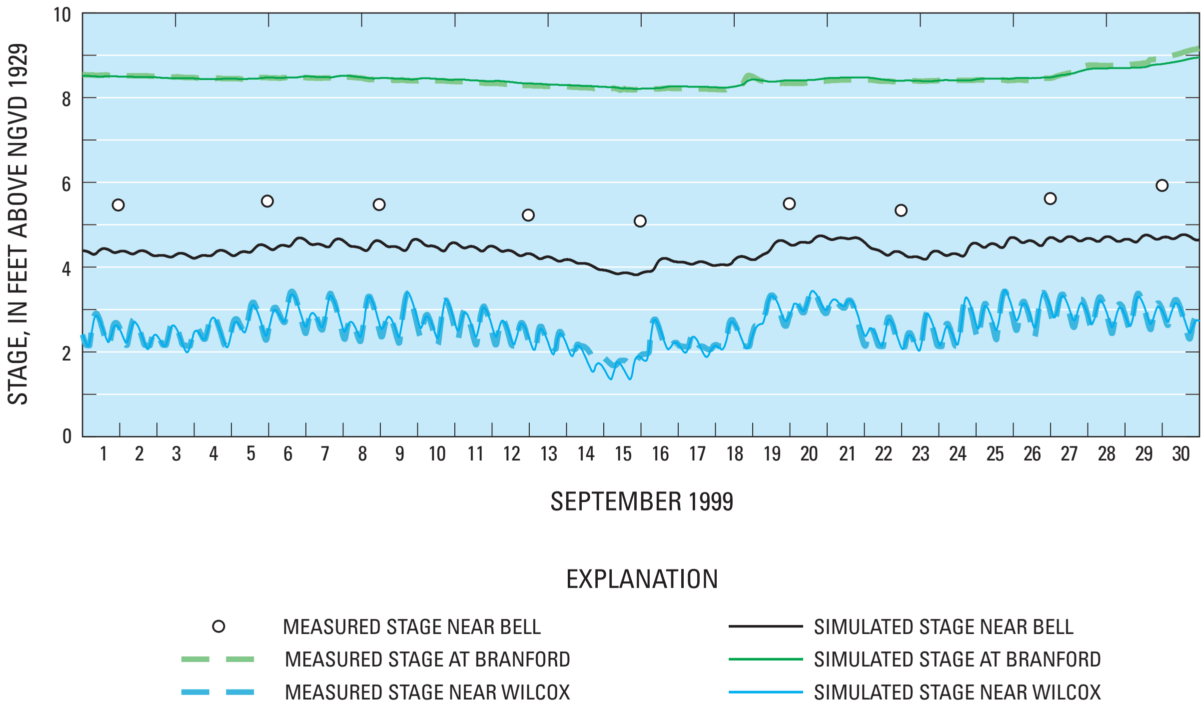


Figure 45. Comparison of simulated and measured water-surface elevations in the Lower Suwannee River during September 1999. Locations are shown in fig. 11.

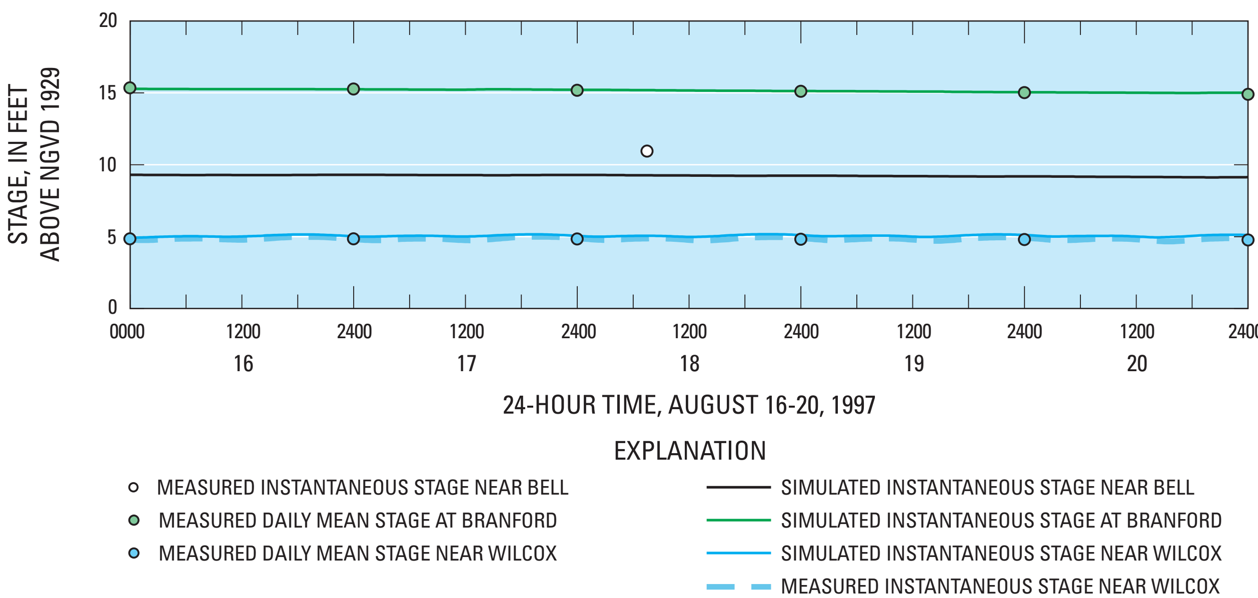


Figure 46. Comparison of simulated and measured water-surface elevations in the Lower Suwannee River during August 19997. Locations are shown in fig. 11.

streamflow conditions in March 1998, simulated stages were 3 to 3.5 ft lower than measured values at the Branford gaging station, 3.1 to 3.8 ft lower at the Bell gaging station, and 1 to 1.5 ft lower at the Wilcox gaging station.

Patterns of simulated and measured streamflows were similar for the Lower Suwannee River gaging stations near Bell and near Wilcox, and at the cross section upstream from Manatee Springs (fig. 4) during September 28-30, 1999 (fig. 47). The simulated tidal minimum and maximum flows, however, were more extreme than the measured flows at the gaging station near Wilcox. Similar results were obtained in the simulation of the average streamflow condition during August 18-20, 1997 (fig. 48), although simulations of some of the flows were affected by problems with the downstream gaging station near Old Town (fig. 6) during low tide. More importantly in terms of the study objectives, comparisons indicated that simulated daily streamflows are probably within 5 to 10 percent of measured streamflows (based on interpolating and extrapolating from the instantaneous measurements with the ADCPs).

Coupled Ground-Water and Surface-Water Flow Model

Calibration of the coupled ground- and surface-water flow model consisted of verifying that the parameter values obtained in the calibrations of the uncoupled ground- and surface-water flow models yielded acceptable results when used in the coupled model. The parameter values used in the coupled model were identical to those used in the uncoupled models, with the exception of riverbed leakance used by the river submodel, BRANCH', in the coupled model. The BRANCH' submodel calculates the conductance between the river and aquifer in each river segment by multiplying the simulated river topwidth by user-specified values of segment length and riverbed leakance (hydraulic conductivity divided by flow path length). In the uncoupled ground-water flow model, the riverbed conductance is specified by the user directly. An attempt was made to use riverbed leakance values in the coupled model that were equivalent to those used in the uncoupled ground-water flow model (by dividing the riverbed

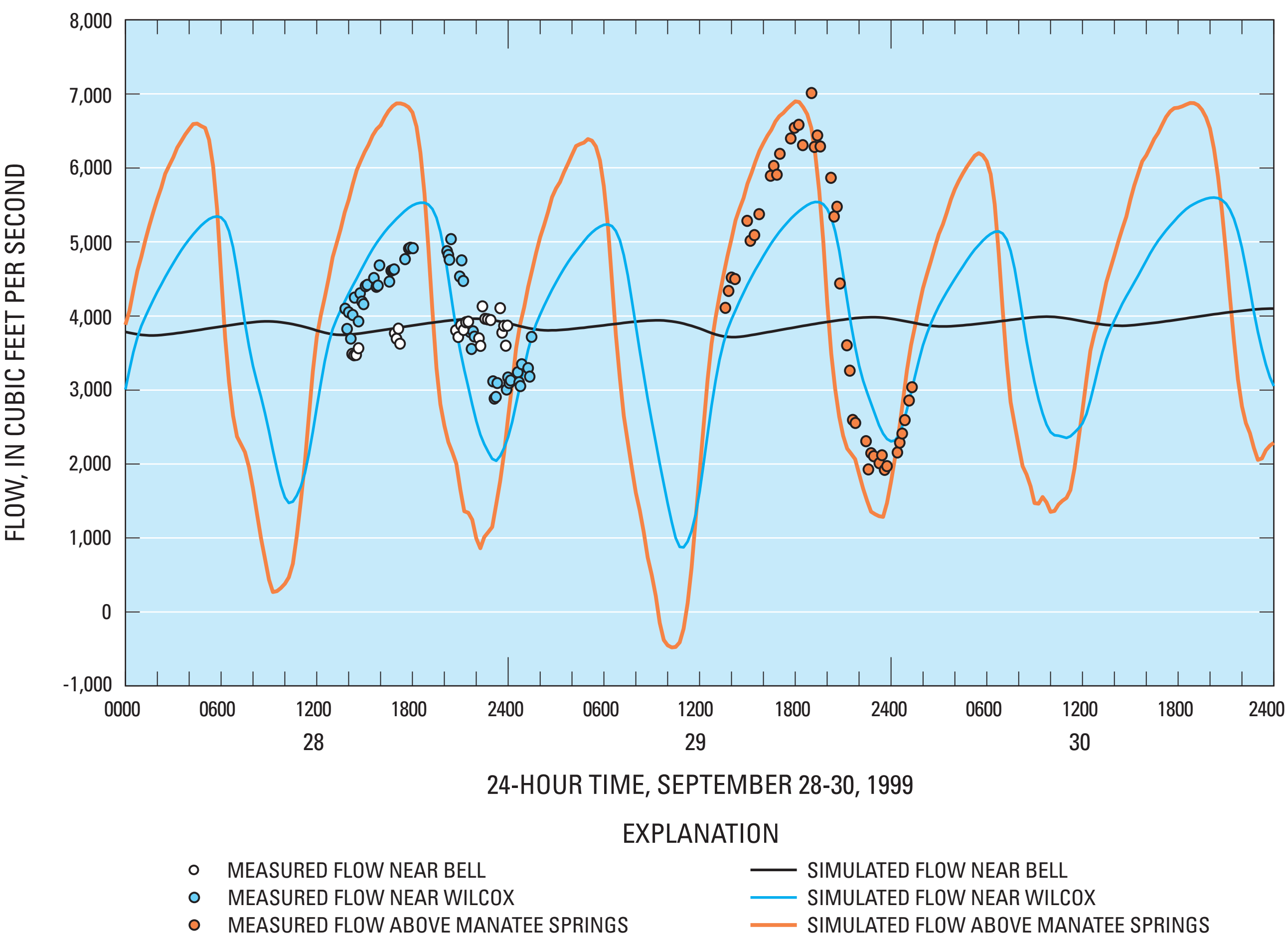


Figure 47. Comparison of simulated and measured flows in the Lower Suwannee River during September 1999. Locations are shown in fig. 4.

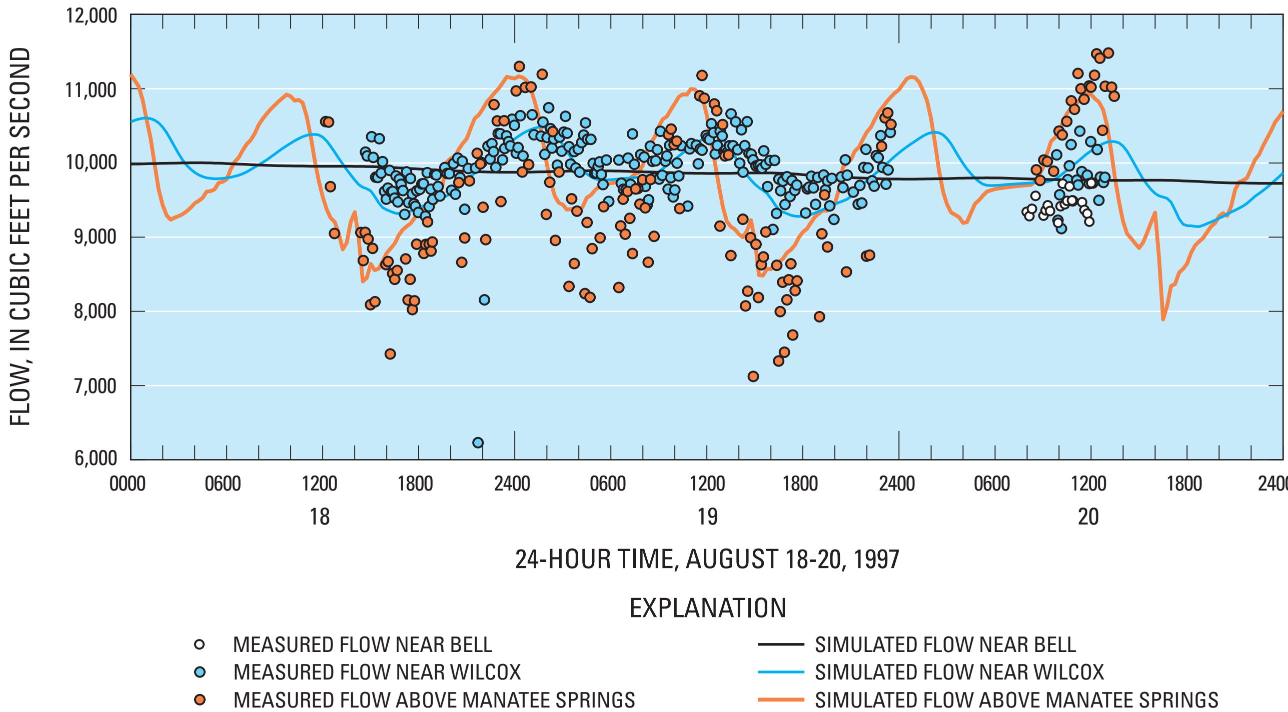


Figure 48. Comparison of simulated and measured flows in the Lower Suwannee River during August 1997. Locations are shown in fig. 4.

conductance values used in the uncoupled model by average river topwidth values). However, these equivalent values proved to be too high for stable convergence of the coupled model, even after adjusting some of the parameters that were used to solve the coupled model equations. Accordingly, the leakance values were decreased incrementally until, at a riverbed leakance value of 0.09 sec^{-1} (1.0 × 10-6 $days^{-1}$), the coupled model would converge to a stable solution.

Generally, simulated ground-water levels in the coupled model were similar to those in the uncoupled ground-water flow model. The ground-water level residuals were reasonably unbiased, with the mean and median values of all residuals equal to -0.5 and -0.7 ft, respectively, and with 50 percent of the residuals ranging from -3.0 to 2.2 ft, and 90 percent ranging from -6.7 to 5.1 ft. The minimum and maximum residuals were similar in magnitude to those obtained from the uncoupled ground-water flow model.

Comparisons of measured and simulated ground-water levels in selected wells during water years 1998-99 are shown in figures 49 to 51. The similarity between measured and simulated ground-water levels generally did not depend on geographic location (figs. 52 and 53), but there were some exceptions. For example, some of the better matches were evident in three of the four wells north of the Santa Fe River (for example, well s245 in fig. 49). Additionally, simulated ground-water levels in wells s315, s330, s331, s346, and t059 (fig. 50) along the Suwannee River upstream from river mile 25 generally compared fairly well with measured water levels, although the matches were not as good as those achieved in the uncoupled ground-water flow model because of the low riverbed leakance values that were necessary for model stability. Good matches were also generally obtained in wells in the Chiefland Limestone Plain east of the Suwannee River and in western Alachua County (for example, wells s364 and s337, fig. 49).

Conversely, the matches for three of the four wells along the Suwannee River downstream from river mile 25 were not as good as in upstream locations (for example, well s360, fig. 51). The poor match in this area may be due to difficulty in accurately simulating recharge rates in this area during the summer periods of 1998 and 1999 and the fall of 1997. Poor matches also occurred at some of the wells located near karst features. At well s385 near Wekiva Springs, for example, simulated water levels were much more variable than the relatively constant measured water levels (fig. 51). The poor simulation at this site probably is because of the lack of data available to determine changes in transmissivity that are present in this area, as evidenced by the difficulty in improving the match at this site while maintaining the quality of the matches of wells to the east. As a result, simulated water levels in well s385 were somewhat low (by an average of about 4 ft) and fell below the elevation of the spring (by an average of about 3 ft), causing the spring discharge from this site not to be simulated by the MODFLOW Drain package. Simulated water levels in wells were also poor near some sinkholes that may occasionally receive storm runoff (for example, well s319, fig. 51); however, some simulations near other such features were much better (for example, well s333, fig. 49).

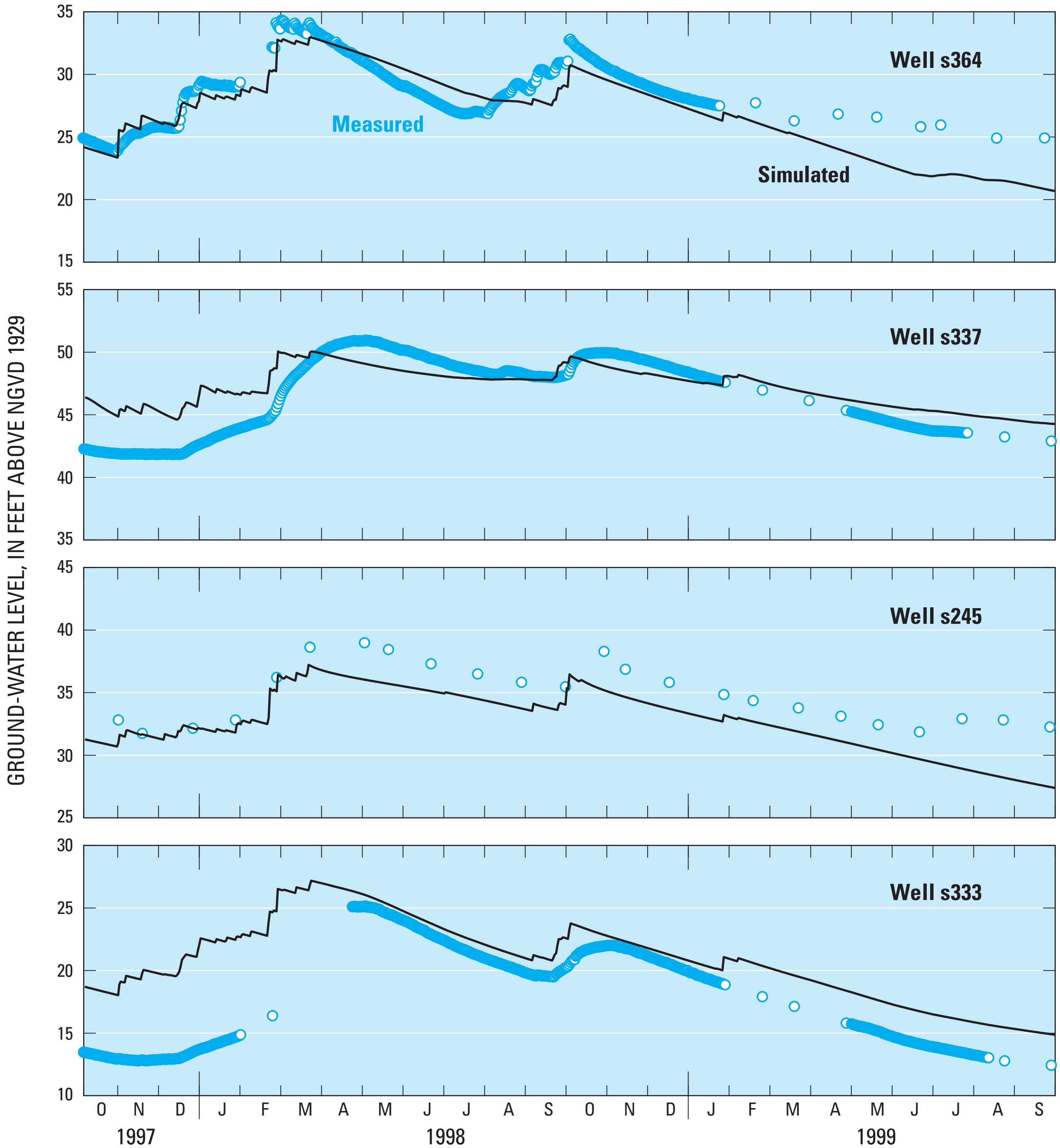


Figure 49. Comparisons of measured and simulated (coupled model) ground-water levels from October 1997 to September 1999 for wells s364, s337, s245, and s333. Well locations are shown in fig. 27.

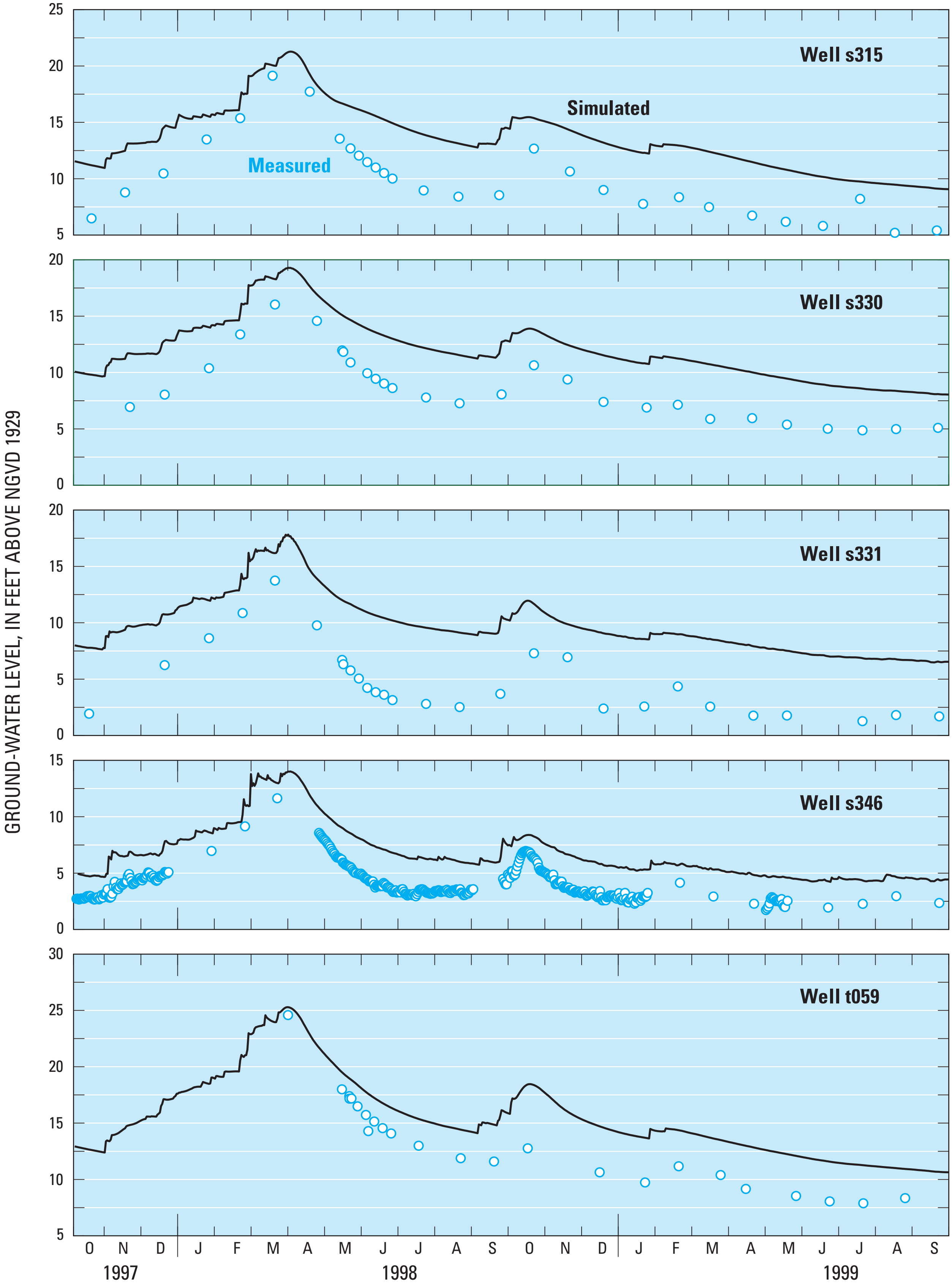


Figure 50. Comparisons of measured and simulated (coupled model) ground-water levels from October 1997 to September 1999 for wells s315, s330, s331, s346, and t059 . Well locations are shown in fig. 27.

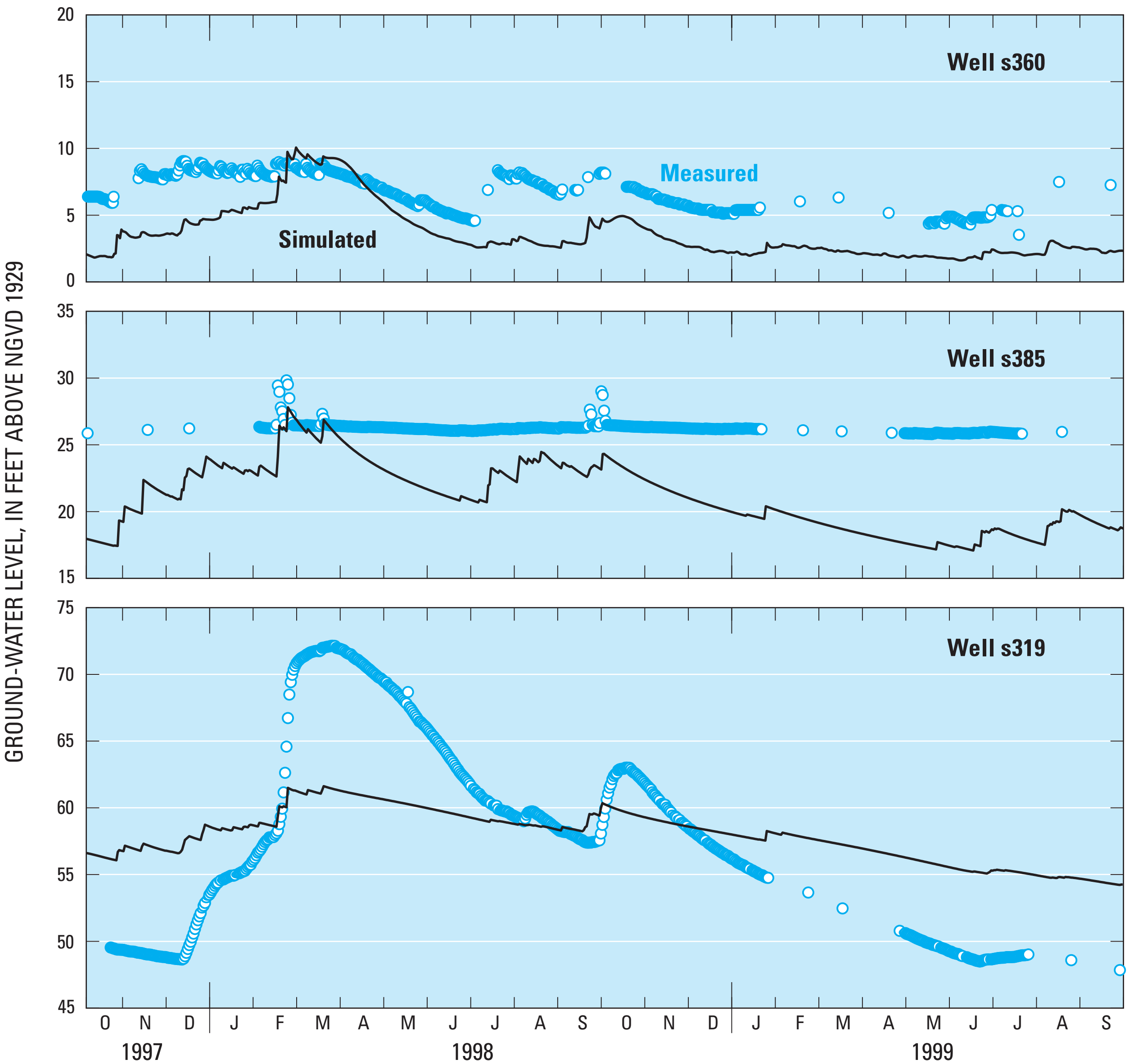


Figure 51. Comparisons of measured and simulated (coupled model) ground-water levels from October 1997 to September 1999 for wells s360, s385, and s319. Wells locations are shown in fig. 27.

The simulated and measured streamflows and stages were generally in good agreement during water years 1998-99 (figs. 54-59). Streamflow residuals (differences between simulated and measured mean daily streamflows) were comparable to the accuracy of "measured" flows (estimated from gaging-station data). For example, in the Suwannee River near Wilcox, 95 percent of the simulated daily mean flows were within 17 to 18 percent of the "rated" daily mean flows during water years 1998-99. These "rated" daily mean flows were estimates based on a relation (rating) between flow, stage, and water-surface slope. This rating is created, checked, and periodically updated using concurrent measurements of flow, stage, and water-surface slope that are taken, on average, about once every 8 weeks. Once this relation is created, hourly estimates of flow are made from hourly measurements of stage and water-surface slope. The hourly estimates for a given day are then averaged to compute a "rated" daily mean flow for that day. For the gaging station on the Suwannee River near Wilcox, the accuracy of this rating was described as "fair" for water year 1998 (95 percent of the daily mean streamflow values were estimated to be within 15 percent of the unknown true daily mean streamflow values) and "poor" for water year 1999 (less than 95 percent of the daily mean streamflow values were estimated to be within 15 percent of the unknown true values). Thus, the simulated daily mean flows at the Wilcox gaging station generally fell within (or close to) the confidence limits of the estimates generated from the rating at this location.

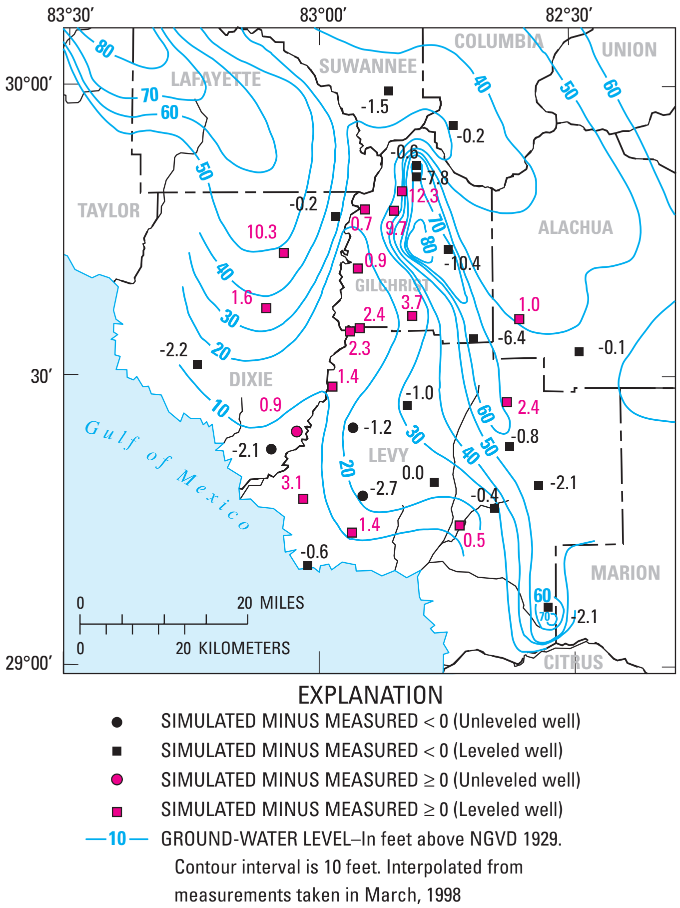


Figure 52. Potentiometric surface of the Upper Floridan aquifer and differences between measured and simulated ground-water levels in March 1998.

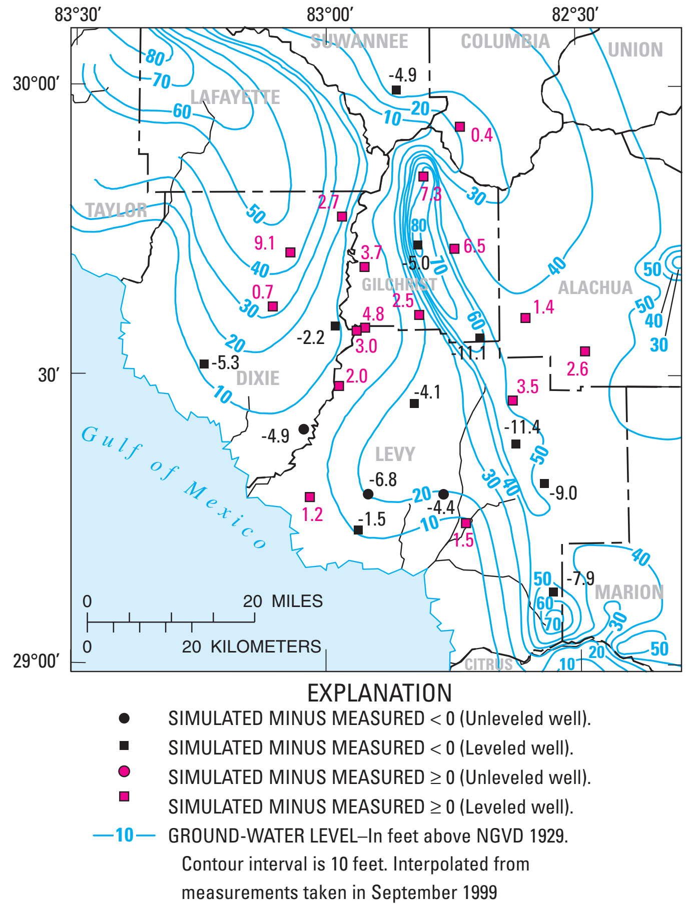


Figure 53. Potentiometric surface of the Upper Floridan aquifer and differences between measured and simulated ground-water levels in September 1999.

The actual accuracy of the simulated flows may be better than indicated in the previous comparison because adjustments (rating shifts) that are applied to the rating are often imprecise because they are based on a limited set of data. This became apparent after analysis of the data collected during this study indicated that individual measurements could be randomly scattered around the rated or simulated flows by several hundred cubic feet per second. For example, the rating shift that was in effect during the summer of 1998 was calculated based on the best fit of data from eight measurements taken at the Wilcox gaging station from October 1997 to January 1999 (only one of which was made during the summer of 1998). Subsequent analysis of the 76 flow measurements taken on August 12, 1998, however, indicated that the flow estimates, computed using this shifted rating, were about 1,000 ft^3/s lower than the measured values during this period. Conversely, simulated flows were generally within 300 ft^3/s of the ADCP-measured flows during this period. If the rating shifts were changed during the summer period to reflect the 76 measurements, then there probably would be an even better agreement between the rated and simulated flows.

The coupled model also simulated streamflow reasonably well at other sites during different periods. Comparisons were made between simulated and ADCP-measured flows during field trips in the summers of 1998 and 1999, when flows were measured nearly continuously for 10 to 13 hours at most sites (figs. 57-59). The simulated instantaneous flows were generally within 10 percent of the measured flows at most of the sites. The simulated mean daily flows during these intensive measurements were probably even more accurate based on comparisons of flows averaged over the 10- to 13-hour measurement periods. More limited ADCP data also were available from routine field trips for the Suwannee River near Wilcox site. These data indicated that flows were simulated by the coupled model during these periods with an accuracy that was comparable to the slope-discharge rating. The poorest simulations probably occurred on June 9, 1998, when 10 ADCP streamflow measurements were taken during a “trough” in the tidal flow hydrograph. The ADCP and slope rating data indicated that the simulated daily mean flow on that day was probably within 10 to 20 percent of the true daily mean flow.

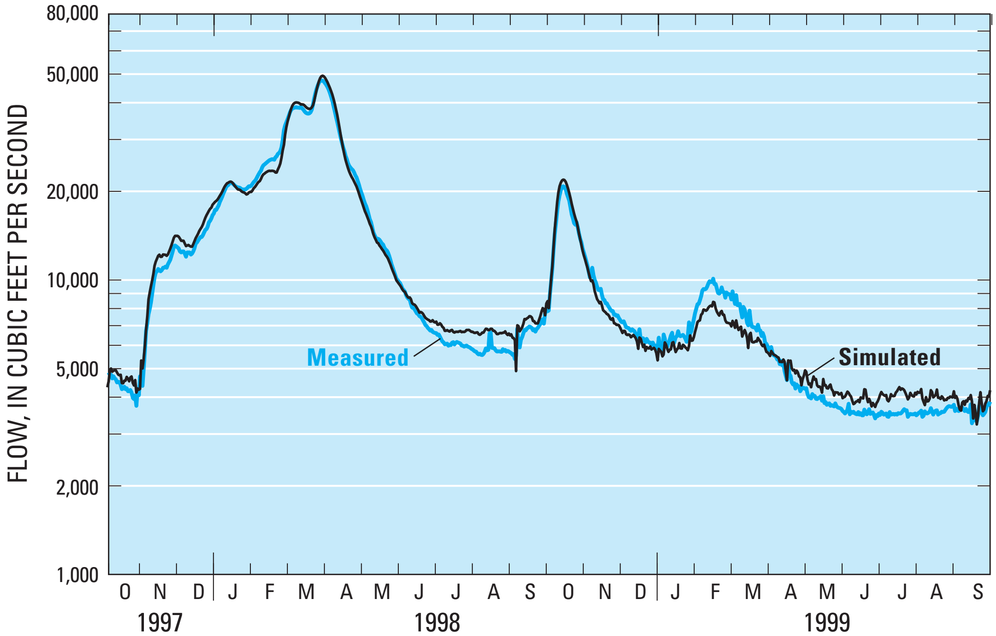


Figure 54. Comparison of measured and simulated (coupled model) flows in the Suwannee River near Wilcox, Florida, October 1997 to September 1999. Site location (map index no. 7) is shown in fig. 4.

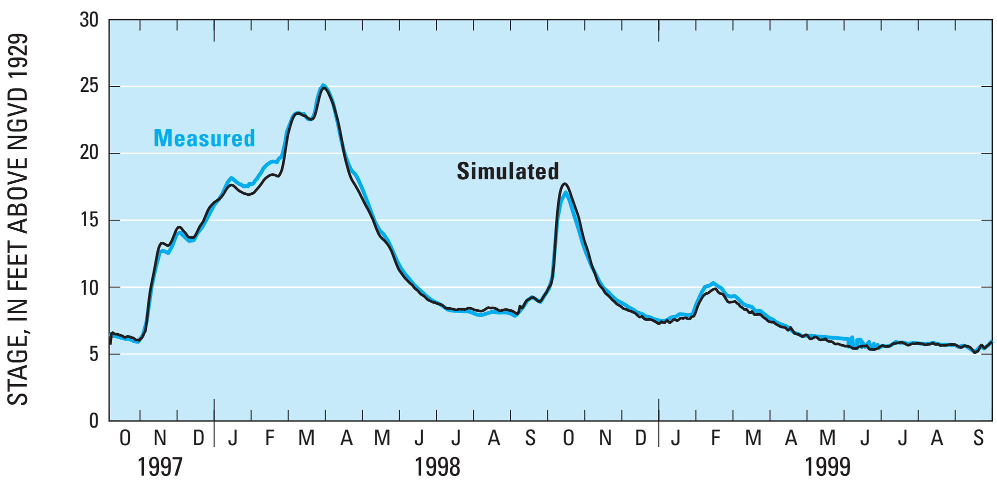


Figure 55. Comparison of measured and simulated (coupled model) stage in the Suwannee River near Bell, Florida, October 1997 to September 1999. Site location (map index no. 6) is shown in fig. 4.

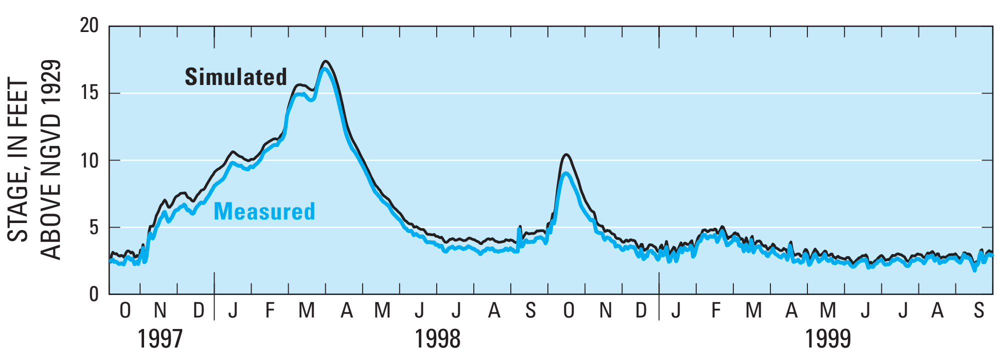


Figure 56. Comparison of measured and simulated (coupled model) stage in the Suwannee River near Wilcox, Florida, October 1997 to September 1999. Site location (map index no. 7) is shown in fig. 4.

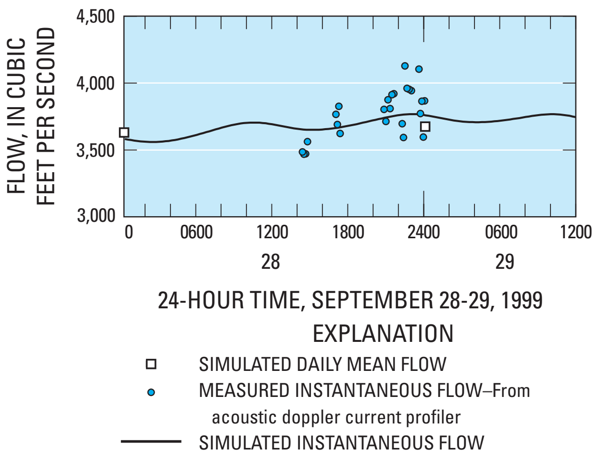


Figure 57. Comparison of measured and simulated (coupled model) flows in the Suwannee River near Bell, Florida, September 28-29, 1999. Site location (map index no. 6) is shown in fig. 4.

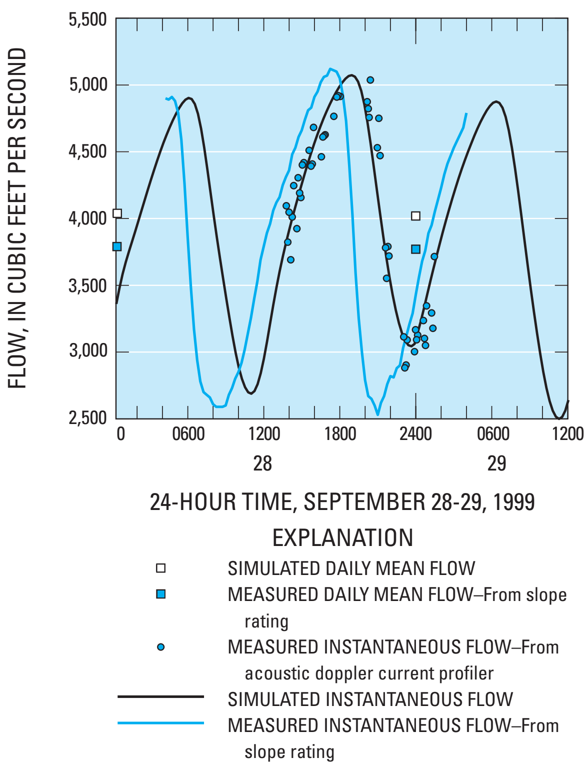


Figure 58. Comparison of measured and simulated (coupled model) flows in the Suwannee River near Wilcox, Florida, September 28, 1999. Site location (map index no 7) is shown in fig. 4.

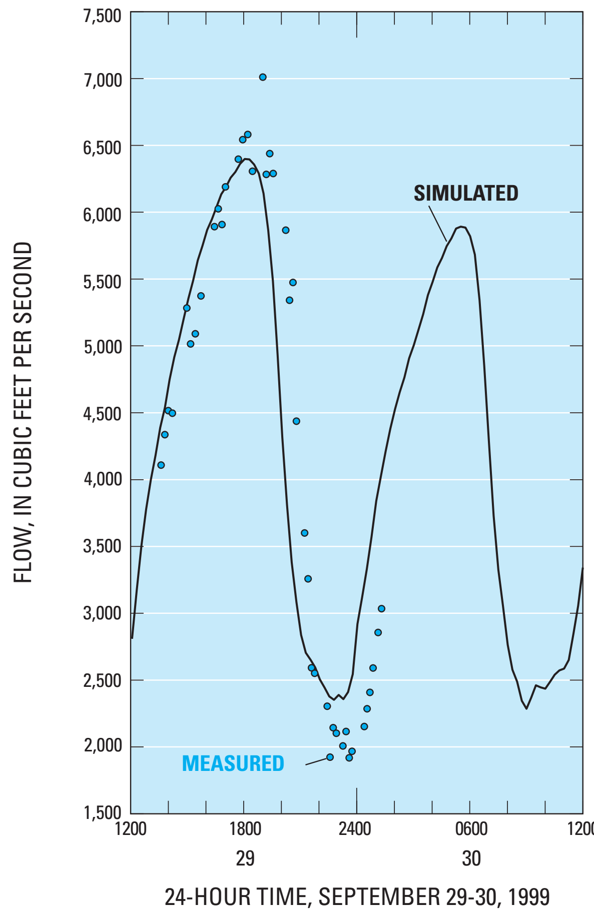


Figure 59. Comparison of measured and simulated (coupled model) flows in the Suwannee River above Manatee Springs near Chiefland, Florida, September 29-30, 1999. Site location (map index no. 19) is shown in fig. 4.

The simulated water budget for the subregional study area compared favorably with the water budget that was estimated independently from the model. The following budget terms were used in this comparison: (1) change in the amount of water stored in the ground-water system from the beginning to the end of the simulation period (water years 1998-99); (2) exchange between the ground-water system and the Lower Suwannee, Lower Santa Fe, and Ichetucknee Rivers; and (3) discharge from the ground-water system to the Gulf of Mexico and to streams and wetlands draining coastal areas along the Gulf of Mexico. A comparison was not made between the simulated value of the flux of ground water across the general-head boundaries of the model (396 ft^3/s) and the model-independent estimate of the flux because of the large uncertainty in computing an estimate of this flux. Additionally, comparisons were not made for the budget terms of recharge and well withdrawals because the model-independent estimates of these terms (4,331 and 40 ft^3/s, respectively) were specified as inputs to the model.

The model-independent estimate of the change in the amount of water stored in the ground-water system within the subregional study area during water years 1998-99 was computed by multiplying the average change in ground-water levels (-0.84 ft) by the range in probable values of specific yield (0.1 to 0.3). This calculation resulted in an estimated reduction in ground-water storage ranging between 5.1×10^9 ft^3 and 1.5×10^{10} ft^3 from the beginning to the end of the study period (or a reduction of 81 to 243 ft^3/s when expressed as a rate of change in storage). The actual range of uncertainty associated with this estimate is larger because of uncertainties associated with the estimated value of the average change in ground-water levels. The corresponding simulated reduction in ground-water storage was 1.2×10^{10} ft^3 (or 195 ft^3/s when expressed as a rate of change), which fell within the range of the model-independent estimate.

The model-independent, average river-aquifer exchange along the Lower Suwannee, Lower Santa Fe, and Ichetucknee Rivers was estimated to range from 2,175 to 3,200 ft^3/s during water years 1998-99. The lower limit of this range was computed by adding the estimated combined average discharge (225 ft^3/s) from Manatee Springs and Fanning Springs (fig. 4) to the measured average streamflow pickup (1,950 ft^3/s) between the: (1) upstream gaging stations on the Suwannee River at Branford and the Santa Fe River at U.S. Highway 441 near High Springs; and (2) downstream gaging station near Wilcox. The upper limit was computed by adding the expected recharge to the estimated ground-water contributing area (to the Lower Suwannee River downstream from the gaging station near Wilcox) to the above-measured streamflow pickup of 1,950 ft^3/s. The corresponding simulated value of the average river-aquifer exchange along the Lower Suwannee, Lower Santa Fe, and Ichetucknee Rivers was 2,760 ft^3/s (within the range of the model-independent estimate).

The model-independent estimate of ground-water discharge to the Gulf of Mexico and to streams and wetlands draining coastal areas along the Gulf of Mexico was 2,125 ft^3/s. This estimate was computed by multiplying the estimated size of the contributing areas to the Gulf of Mexico by the average recharge rates to these areas during water years 1998-99. This model-independent estimate was nearly identical to the simulated value of 2,129 ft^3/s.

Model Evaluation

This section describes two applications of the coupled model of ground- and surface-water flow. The first application is an evaluation of the temporal and spatial characteristics of simulated water exchanges between the Lower Suwannee and Lower Santa Fe Rivers and the Upper Floridan aquifer. The second application is an evaluation of the potential effects of future ground- and surface-water withdrawals using several hypothetical water withdrawal scenarios. The section concludes with a discussion of some of the limitations of the coupled model.

Temporal and Spatial Characteristics of Ground- and Surface-Water Exchanges

The coupled model was used to evaluate the temporal and spatial characteristics of exchanges of water between the Lower Suwannee and Lower Santa Fe Rivers and the Upper Floridan aquifer during water years 1998-99. This evaluation focused primarily on the Suwannee River downstream from the mouth of the Santa Fe River and upstream from river mile 7.5, referred to as reach 3 (fig. 11), because of concerns about possible consequences of increased water use on flow in this reach. Another major group of river reaches, referred to collectively as reach 1 (fig. 11), is bounded by long-term gaging stations. As previously discussed, reach 1 consists of the Suwannee River between the gaging stations at Branford and near Wilcox (river miles 76.1 and 33.5, respectively), the Santa Fe River downstream from the gaging station near Fort White (18.3 miles upstream from the mouth of the Santa Fe River), and the Ichetucknee River. Simulation results from reach 1 were compared with estimates obtained from the previously described analysis of changes in streamflow and channel storage along this reach (referred to as the corrected pickup analysis).

Over the 2-year study period, simulated (coupled model) net ground-water inflow to reach 1 averaged about 1,400 ft^3/s. Although this simulated value was higher than the estimate (1,050 ft^3/s) obtained from the corrected pickup analysis for this reach over the same time period, the latter value has limited significance (with a 95-percent confidence interval of -1,500 to 3,600 ft^3/s). As previously described, many of the estimates from the corrected pickup analysis have limited significance because of the uncertainties in the flow estimates from gaging stations at the upstream and downstream ends of the reach, and because of uncertainties in the estimates of channel storage changes during periods of rapidly rising or falling river levels. Therefore, the discrepancy between the simulated and corrected pickup estimates of the average net ground-water inflow to reach 1 during water years 1998-99 is not significant.

The simulated (coupled model) average ground-water inflows to reach 3 were about 1,600 and 1,300 ft^3/s, during water years 1998 and 1999, respectively. The difference in these two values is probably attributable to: (1) high rates of ground-water inflow to reach 3 that occurred just before the onset, and just after the peak, of the El Niño-associated flooding during the winter-early spring of water year 1998; and (2) lower ground-water inflow rates to reach 3 caused by the extended La Niña-associated drought that occurred during most of water year 1999.

Shorter than annual time-scale averages support this inference. The simulated monthly ground-water inflows to reach 3 ranged from 1,000 to 2,400 ft^3/s, with the maximum value occurring in April 1998, following the winter flood peak of late March 1998. minimum values occurred in September 1999 and the preceding spring and summer months of 1999 when extended drought conditions occurred in association with La Niña, and during the winter months of late 1997 and early 1998 when river levels were rising faster than ground-water levels. The simulated daily ground-water inflow data show a similar pattern (fig. 60). The minimum daily ground-water inflow (about 500 ft^3/s) occurred on March 24, 1998, which is coincident with the peak of the winter-early spring flood period during water year 1998. The maximum values (2,800 to 2,900 ft^3/s) occurred on and around: (1) February 23, 1998, during a period of rapidly rising ground-water levels; and (2) April 13, 1998, about 3 weeks after the winter flood peak when river levels were falling faster than ground-water levels.

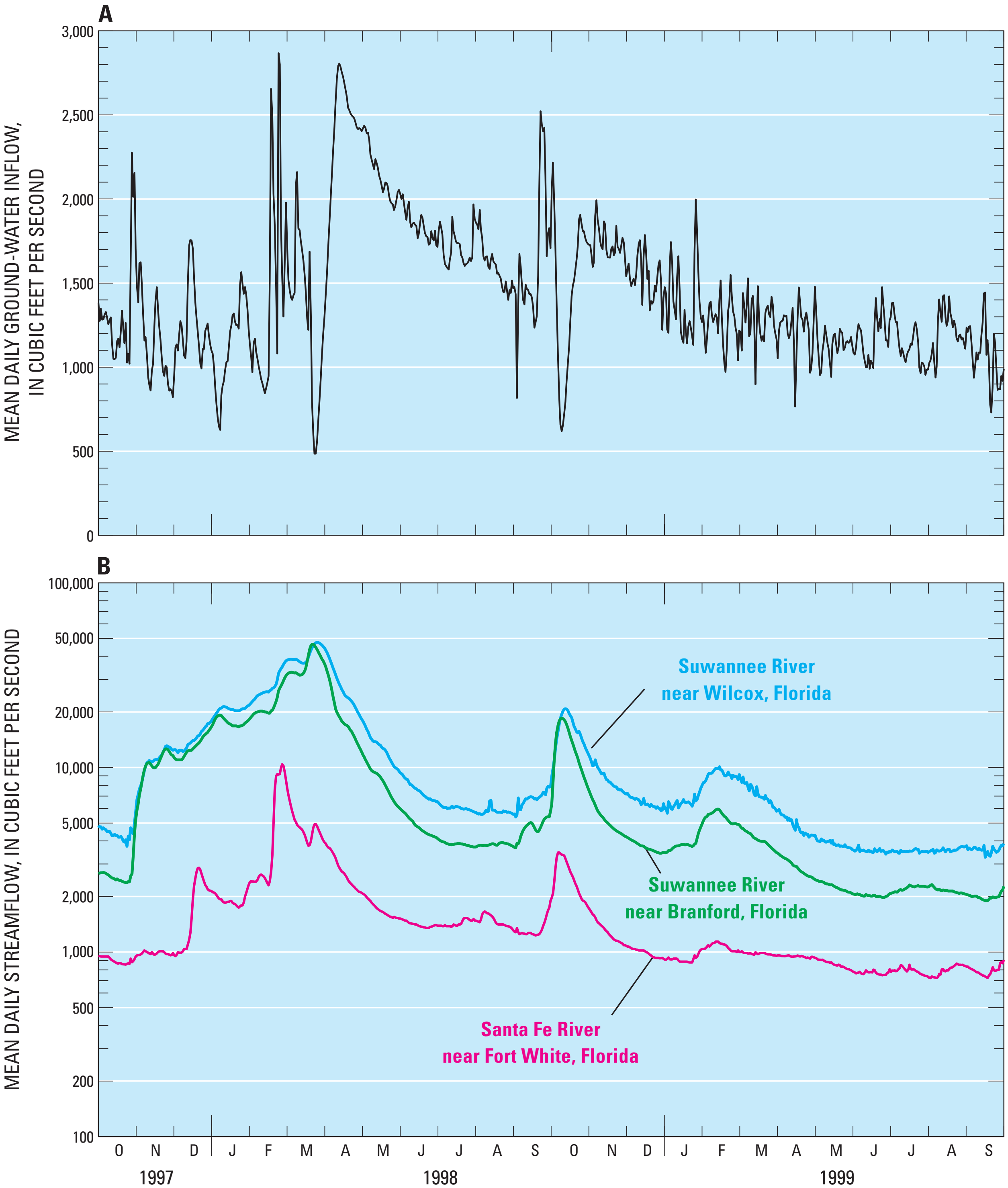


Figure 60. (A) Simulated daily ground-water inflow to reach 3, and (B) measured daily streamflows in the Suwannee River at Bradford, Suwannee River near Wilcox, and Santa Fe River near Fort White, Florida.

Collectively, the annual, monthly, and daily data show a consistent pattern in which low rates of ground-water inflow occur during periods of peak flood levels, when the hydraulic gradient (aquifer level minus river level) toward the river is typically at its lowest point because of the high river levels. Low rates of ground-water inflow also occur after extended droughts, when ground-water storage has been depleted from reduced recharge. Conversely, the highest rates of ground-water inflow typically occur during periods of receding water levels that closely follow peak river levels. During these conditions, the hydraulic gradient from the aquifer to the river is typically greatest because river levels have fallen and ground-water levels are higher. The ground-water levels are higher because of: (1) high recharge rates associated with the wet conditions that lead up to the flood peak, and (2) the accumulation of ground water during the period of peak flooding when the high river levels impeded ground-water outflow. These patterns in the simulated data are also consistent with the patterns inferred from the corrected pickup analysis.

The simulation results from the coupled model reproduced the essential spatial patterns that were evident from the historic and study period streamflow measurements. The reaches with the highest average ground-water inflow rates during water years 1998-99 were the Ichetucknee River above U.S. Highway 27 (about 130 ft^3/s per mile of river) and the Santa Fe River upstream from the gaging station near Fort White to the upstream model boundary near U.S. Highway 27 (about 80 ft^3/s per mile of river). As previously discussed, these reaches have the highest concentration of spring discharge in the subregional study area and also the greatest change in measured pickup under average and low-flow conditions. Additionally, the simulated ground-water inflows to the Suwannee River between the gaging station at Branford and the mouth of the Suwannee River (about 9 ft^3/s per mile of river) and the Santa Fe River downstream from the Fort White gaging station (about 7 ft^3/s per mile of river) had the lowest average ground-water inflow rates. The reach of the Ichetucknee River upstream from the gaging station near Hildreth at U.S. Highway 27 had the greatest variability in the amount of ground-water-to-river inflow; the standard deviation of the daily ground-water inflows was about 30 ft^3/s per river mile (fig. 61). In contrast, the reach of the Santa Fe River downstream from the Fort White gaging station and the reach of the Suwannee River between Branford and the mouth of the Santa Fe River had the least variability in the amount of ground-water-to-river inflow; the standard deviation of the daily ground-water inflows was 3 and 4 ft^3/s per mile in these reaches, respectively.

Simulated Effects of Water Withdrawals on Streamflow

The effect of increased water withdrawals on the flow of the Lower Suwannee River is a major concern to water managers and others. Because of the closely linked ground- and surface-water systems in the Lower Suwannee Basin, ground-water withdrawals will, in many instances, result in decreased ground-water discharge to, and subsequently reduced flow in, the Lower Suwannee River. The coupled model of ground- and surface-water flow was developed, in part, as a tool to estimate these effects.

To explore this important potential use of the coupled model, 11 hypothetical water-withdrawal scenarios were simulated (table 8), and their effects on a set of statistical descriptions of streamflow (streamflow characteristics) were evaluated at the gaging stations on the Suwannee River near Bell, near Wilcox, and above the Gopher River (fig. 4 and table 1). Several of these scenarios (scenario group 4) correspond to those used by Light and others (2002), who evaluated potential changes in floodplain forest composition and the degree of inundation and saturation of floodplain soils along the Lower Suwannee River. Eight streamflow characteristics were evaluated for each scenario-gaging station combination. Three of these characteristics were the lowest daily flow that is not exceeded in a given year, on average, once every 2, 5, or 10 years (1Q2, 1Q5, and 1Q10, respectively). The other five characteristics were the daily flows that are not exceeded 10, 20, 50, 80, and 90 percent of the time (p10, p20, p50, p80, and p90, respectively).

The approach used to estimate values of each of the eight statistics under the various scenarios consists of three steps as illustrated for scenario 4 (withdrawal = 500 ft^3/s) in the Suwannee River near Wilcox (figs. 62 and 63). In the first step, a given hypothetical scenario is simulated with the coupled model for water years 1998-99 by modifying the boundary conditions of the calibrated model (referred to as scenario 1 or the baseline scenario) to accommodate the hypothetical scenario. For example, a hypothetical scenario might be created by increasing the 1998-99 monthly ground-water withdrawal rates that were used in the calibrated model (which averaged, 35 ft^3/s), or by reducing the flow at the upstream river boundaries of the calibrated model.

In the second step, the daily mean flows at each of the three gaging station locations are extracted from the model output for each day of the simulation period (water years 1998-99). These simulated daily mean flows are then matched with concurrent daily mean flows that were simulated for the baseline conditions during water years 1998-99 using the results of the calibration of the coupled model.

In the third step, a linear relation is fit to the dataset generated in the second step, using the line-of-organic correlation method of regression analysis (Helsel and Hirsch, 1992, p. 276). Three linear relations (one for each gaging station) are produced in this step, with each relation then used to estimate the values of individual streamflow characteristics for a given scenario gaging station/combination. For example, the 1Q10 streamflow characteristic for the Suwannee River near Wilcox under scenario 4 is estimated by substituting the values of 1Q10 under baseline conditions into the best-fit relation between flows in the Suwannee River near Wilcox under scenario 4 and baseline conditions. This step produces a set of eight streamflow characteristics for each gaging station/scenario combination. Using this approach for

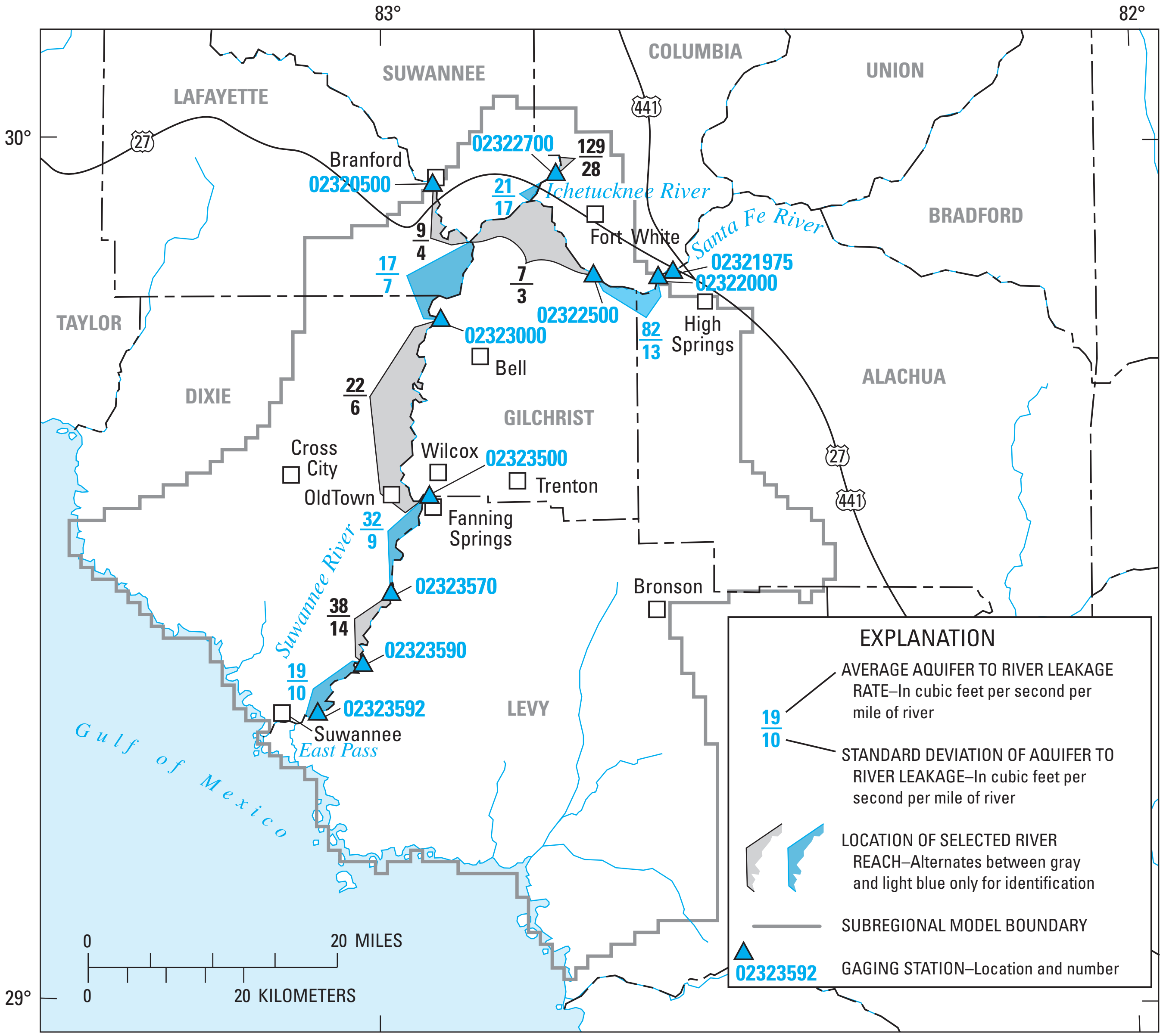


Figure 61. Simulated average and standard deviation of ground-water leakage to selected river reaches.

estimating changes in streamflow characteristics is analogous to the approach used for estimating streamflow characteristics at sites with limited data using relations with long-term sites (Riggs, 1972; Helsel and Hirsch, 1992, p. 265).

Results of the scenario simulations for the three gaging stations are given in tables 9 to 13. The streamflow characteristics estimated for each scenario at the gaging stations near Bell and Wilcox, respectively, are presented in tables 9 and 10. These streamflow characteristics could not be estimated for the gaging station on the Suwannee River above the Gopher River because this site did not have enough data to estimate baseline streamflow characteristics. The values of the slope and intercept parameters that define the linear relation between the hypothetical and baseline scenario conditions are presented for the gaging station on the Suwannee River above the Gopher River in table 11. The values can be used to estimate the streamflow characteristics when sufficient data become available at the gaging station above the Gopher River to estimate the baseline streamflow characteristics for this site. These values also can be used to estimate various other streamflow characteristics that were not considered in the examples used in this report. For this reason, slope and intercept parameters are presented for the gaging stations near Bell and Wilcox in tables 12 and 13, respectively.

Table 8. Description of baseline and hypothetical withdrawal scenarios that were simulated with the coupled model of ground-water and surface-water flow.

[ft^3/s, cubic feet per second]

Scenario number	Scenario description
1 (baseline conditions)	Model boundary conditions from calibrated model of water years 1998 and 1999 (total ground-water withdrawals were approximately 35 ft^3/s).
2	Same as baseline conditions scenario except for the addition of a 540 ft^3/s withdrawal from the Suwannee River just upstream from the gage near Wilcox.
3	Same as baseline conditions scenario except ground-water withdrawals are increased by 67 percent (to approximately 58 ft^3/s).
4 (100 ft^3/s)	Same as baseline conditions scenario except the river inflow at the upstream boundaries (Suwannee River near Branford and Santa Fe River near Fort White) is reduced by 100 ft^3/s. Eighty percent of the reduction is assigned to the Suwannee boundary and 20 percent to the Santa Fe boundary.
4 (300 ft^3/s)	Same as scenario 4 (100 ft^3/s) except a 300 ft^3/s reduction is imposed.
4 (500 ft^3/s)	Same as scenario 4 (100 ft^3/s) except a 500 ft^3/s reduction is imposed.
4 (1,000 ft^3/s)	Same as scenario 4 (100 ft^3/s) except a 1,000 ft^3/s reduction is imposed.
4 (2,000 ft^3/s)	Same as scenario 4 (100 ft^3/s) except a 2,000 ft^3/s reduction is imposed.
5 (300 ft^3/s)	Combination of scenario 3 and scenario 4 (300 ft^3/s).
5 (500 ft^3/s)	Combination of scenario 3 and scenario 4 (500 ft^3/s).
5 (1,000 ft^3/s)	Combination of scenario 3 and scenario 4 (1,000 ft^3/s).
6 (no pumping)	Same as baseline conditions scenario except ground-water withdrawals are not simulated (to approximate predevelopment conditions).

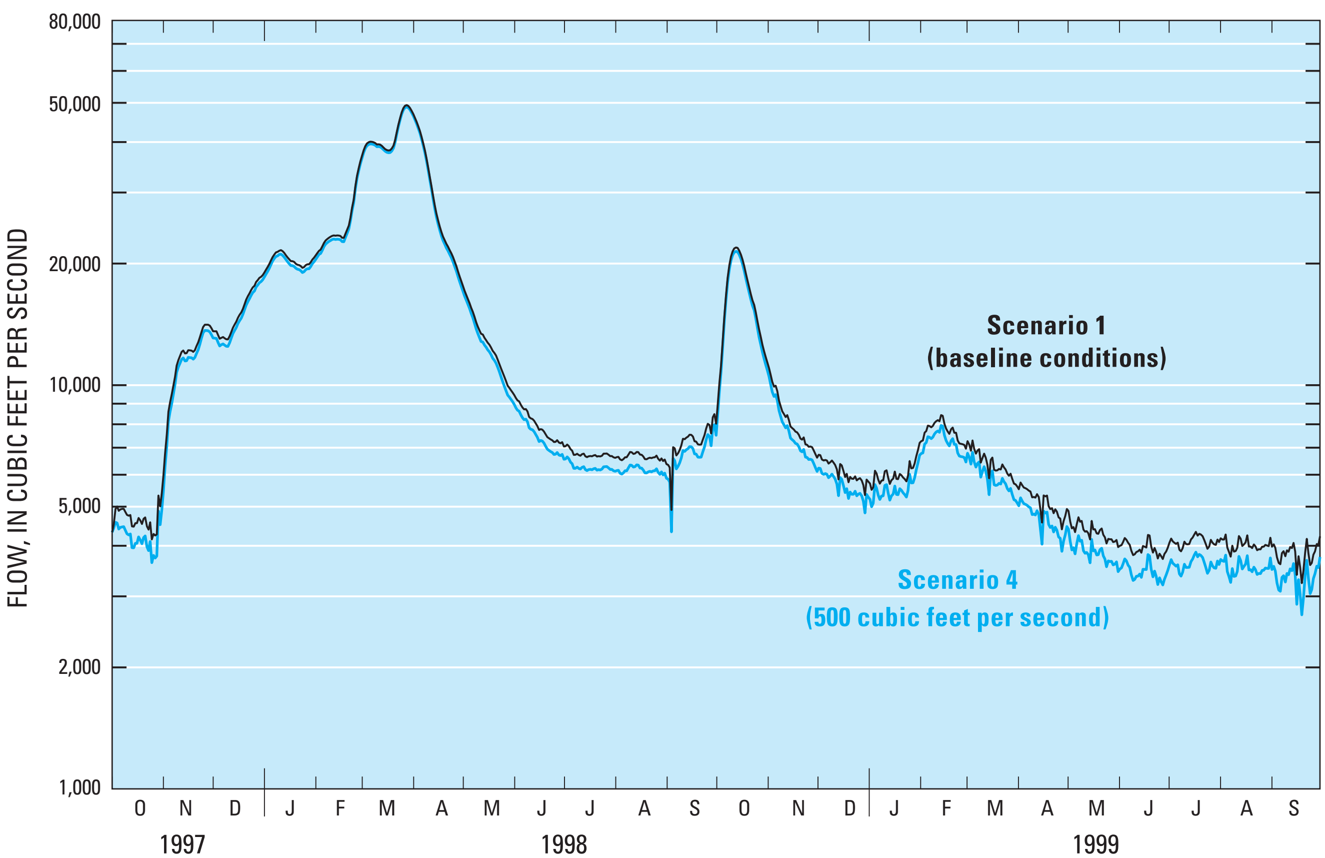


Figure 62. Mean daily flows during water years 1998-99, simulated for scenarios 1 and 4 in the Suwannee River near Wilcox, Florida.

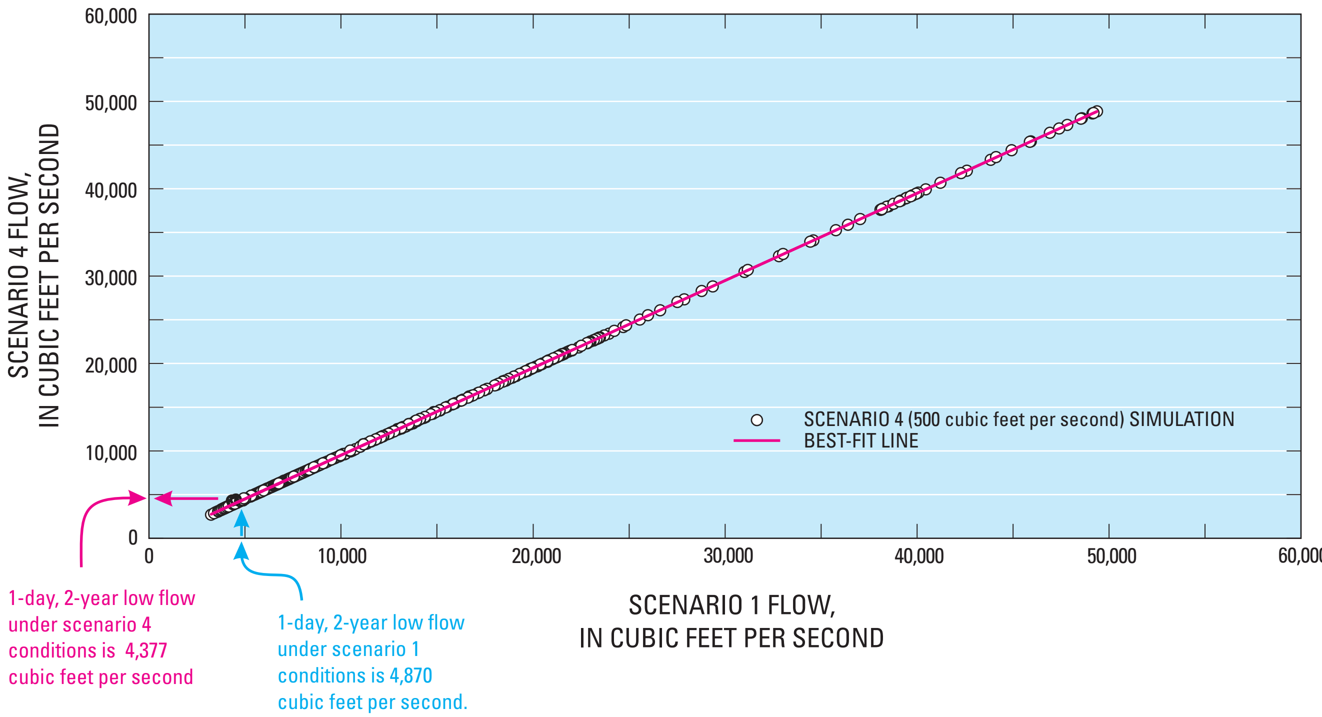


Figure 63. Relation between daily flows during water years 1998-99, simulated for scenarios 1 and 4. The procedure for estimating the 1-day, 2-year low-flow frequency characteristic of the Suwannee River near Wilcox with this relation is illustrated.

Results of the scenario simulations for the three gaging stations indicated that instream withdrawals produced relatively straightforward changes in streamflow characteristics. For example in scenario 2, the hypothetical point withdrawal of 540 ft^3/s just upstream from the Wilcox gaging station resulted in negligible changes in streamflow at the gaging station near Bell (which was upstream from the point withdrawal), and changes that were nearly equal to the instream withdrawal rate at the gaging stations near Wilcox and above Gopher River (which were both downstream from the point withdrawal). Similarly, for scenario group 4 (flow reductions of 100, 300, 500, 1,000, and 1,000 ft^3/s), reductions in flows at the gaging stations near Bell, near Wilcox, and above the Gopher River were similar to the hypothetical combined flow reductions at the upstream gaging stations on the Suwannee River at Branford and the Santa Fe River near Fort White. For example, the version of scenario 4 with the hypothetical upstream flow reduction of 1,000 ft^3/s resulted in downstream flow reductions of 988, 989, and 989 ft^3/s at the gaging stations near Bell, near Wilcox, and above the Gopher River, respectively (tables 9-13). These reductions were slightly smaller than the upstream withdrawal rate because the flow reductions decreased the river stages, which increased the hydraulic-head gradient and ultimately increased the flux of ground water to the river.

Scenarios 3 and 6 represented changes in ground-water withdrawals from the system. Changes in streamflows (relative to baseline conditions) were a function of the size and location of the ground-water contributing area to a given gaging station and the geographic distribution of wells. For example, changes in streamflows were smallest for the gaging station near Bell because its contributing ground-water area was the smallest of the three gaging stations. Under scenario 3, streamflows were reduced by about 2, 7, and 14 ft^3/s on average at the gaging stations near Bell, near Wilcox, and above the Gopher River, respectively (tables 9-13). Reductions in flows were less than they might have otherwise been because the increased ground-water withdrawals were offset, in part, by increased movement of ground water into the model area across the general-head boundaries. Under scenario 6, streamflows increased by about 3, 11, and 20 ft^3/s, on average, at the gaging stations near Bell, near Wilcox, and above the Gopher River, respectively (tables 9-13). Increases in flows were less than they might have otherwise been because the decreased ground-water withdrawals resulted in decreasing rates of ground-water movement into the model area across the general-head boundaries.

Scenario group 5 (flows of 300, 500, and 1,000 ft^3/s) represents a combination of instream and ground-water withdrawals. The simulation results resembled the addition or "superposition" of the results from scenarios 3 and 4 (flows of 300, 500, and 1,000 ft^3/s). For example, in scenario 5 (300 ft^3/s), streamflows decreased by 299 ft^3/s at the Bell gaging station, which was equal to the sum of the river-flow reductions under scenarios 3 (2 ft^3/s) and 4 (300 ft^3/s) at 297 ft^3/s. Similar results were obtained at the other gaging stations (tables 9-13).

Table 9. Estimates of streamflow characteristics of the Suwannee River near Bell, Florida, under baseline and hypothetical conditions.

[The annual low flow statistics reported in columns two through four represent the magnitude of the lowest 1-day average flow occurring during a given year at the specified recurrence interval. For example, for scenario 1, the lowest daily average flow is not expected to exceed 2,890 ft^3/s (cubic feet per second) in 1 year out of 10 (10-year recurrence interval), on average]

Scenario number for indicated recurrence interval, in years	Annual lowest 1-day flow, in ft^3/s, for indicated recurrence interval, in years			Daily mean flow that would be exceeded for indicated percent of time, in ft^3/s				
	2-year	5-year	10-year	10 percent	20 percent	50 percent	80 percent	90 percent
1 (baseline conditions)	4,050	3,220	2,890	15,439	11,352	6,664	4,411	3,711
2	4,050	3,220	2,890	15,439	11,352	6,664	4,411	3,711
3	4,048	3,218	2,888	15,437	11,350	6,662	4,409	3,709
4 (100 ft^3/s)	3,951	3,121	2,791	15,339	11,253	6,565	4,312	3,612
4 (300 ft^3/s)	3,753	2,923	2,593	15,140	11,054	6,367	4,114	3,414
4 (500 ft^3/s)	3,555	2,726	2,396	14,941	10,855	6,169	3,916	3,217
4 (1,000 ft^3/s)	3,061	2,231	1,902	14,442	10,358	5,673	3,422	2,722
4 (2,000 ft^3/s)	2,071	1,242	913	13,445	9,363	4,682	2,432	1,732
5 (300 ft^3/s)	3,751	2,921	2,591	15,138	11,052	6,365	4,112	3,412
5 (500 ft^3/s)	3,553	2,724	2,394	14,939	10,853	6,166	3,914	3,214
5 (1,000 ft^3/s)	3,059	2,229	1,899	14,440	10,356	5,671	3,419	2,720
6 (no pumping)	4,053	3,223	2,893	15,442	11,355	6,667	4,414	3,714

Table 10. Estimates of streamflow characteristics of the Suwannee River near Wilcox, Florida, under baseline and hypothetical conditions.

[The annual low flow statistics reported in columns two through four represent the magnitude of the lowest 1-day average flow occurring during a given year at the specified recurrence interval. For example, for scenario 1, the lowest daily average flow is not expected to exceed 3,680 ft^3/s (cubic feet per second) in 1 year out of 10 (10-year recurrence interval), on average]

Scenario number for indicated recurrence interval, in years	Annual lowest 1-day flow, in ft^3/s, for indicated recurrence interval, in years			Daily mean flow that would be exceeded for indicated percent of time, in ft^3/s				
	2-year	5-year	10-year	10 percent	20 percent	50 percent	80 percent	90 percent
1 (baseline conditions)	4,870	4,050	3,680	18,835	14,631	8,426	5,666	4,817
2	4,329	3,509	3,139	18,293	14,089	7,885	5,125	4,276
3	4,862	4,043	3,673	18,829	14,624	8,419	5,659	4,810
4 (100 ft^3/s)	4,771	3,951	3,582	18,735	14,532	8,327	5,567	4,718
4 (300 ft^3/s)	4,574	3,754	3,384	18,535	14,333	8,129	5,370	4,521
4 (500 ft^3/s)	4,377	3,557	3,187	18,336	14,134	7,931	5,173	4,324
4 (1,000 ft^3/s)	3,884	3,064	2,695	17,836	13,636	7,437	4,679	3,831
4 (2,000 ft^3/s)	2,896	2,078	1,709	16,837	12,640	6,446	3,691	2,844
5 (300 ft^3/s)	4,567	3,747	3,377	18,529	14,326	8,122	5,363	4,514
5 (500 ft^3/s)	4,370	3,550	3,180	18,329	14,127	7,924	5,165	4,317
5 (1,000 ft^3/s)	3,877	3,057	2,688	17,830	13,629	7,430	4,672	3,824
6 (no pumping)	4,881	4,061	3,691	18,845	14,641	8,436	5,677	4,828

Table 11. Estimates of the parameters necessary for estimating changes in streamflow characteristics of the Suwannee River above the Gopher River under baseline and hypothetical conditions.

[Estimates of streamflow characteristics are calculated using a linear relation defined by the slope and intercept parameters shown below. For example, if the baseline 1-day, 2-year low flow (1Q2) were estimated to be 5,000 ft^3/s (cubic feet per second), then the estimate of the 1Q2 under scenario 2 conditions would be equal to -541 + 0.99997*5,000 = 4,459 ft^3/s]

Scenario number	Parameters defining the linear relation between the flow occurring under the indicated scenario and the flow under baseline conditions	
	Intercept	Slope
2	-541	0.99997
3	-15	1.00016
4 (100 ft^3/s)	-98	0.99991
4 (300 ft^3/s)	-295	0.99972
4 (500 ft^3/s)	-491	0.99953
4 (1,000 ft^3/s)	-982	0.99908
4 (2,000 ft^3/s)	-1,965	0.99816
5 (300 ft^3/s)	-309	0.99988
5 (500 ft^3/s)	-506	0.99969
5 (1,000 ft^3/s)	-997	0.99923
6 (no pumping)	22	0.99976

Table 12. Estimates of the parameters necessary for estimating changes in streamflow characteristics of the Suwannee River near Bell, Florida, under baseline and hypothetical conditions.

[Estimates of streamflow characteristics are calculated using a linear relation defined by the slope and intercept parameters shown below. For example, if the baseline 1-day, 2-year low flow (1Q2) were estimated to be 5,000 ft^3/s (cubic feet per second), then the estimate of the 1Q2 under scenario 3 conditions would be equal to -2 + 1.00001*5,000 = 4,998 ft^3/s]

Scenario number	Parameters defining the linear relation between the flow occurring under the indicated scenario and the flow under baseline conditions	
	Intercept	Slope
2	0	0.99999
3	-2	1.00001
4 (100 ft^3/s)	-99	0.99994
4 (300 ft^3/s)	-296	0.99981
4 (500 ft^3/s)	-493	0.99968
4 (1,000 ft^3/s)	-987	0.99935
4 (2,000 ft^3/s)	-1,974	0.99869
5 (300 ft^3/s)	-296	0.99981
5 (500 ft^3/s)	-496	0.99969
5 (1,000 ft^3/s)	-989	0.99937
6 (no pumping)	3	0.99998

Table 13. Estimates of the parameters necessary for estimating changes in streamflow characteristics of the Suwannee River near Wilcox, Florida, under baseline and hypothetical conditions.

[Estimates of streamflow characteristics are calculated using a linear relation defined by the slope and intercept parameters shown below. For example, if the baseline 1-day, 2-year low flow (1Q2) were estimated to be 5,000 ft^3/s (cubic feet per second), then the estimate of the 1Q2 under scenario 2 conditions would be equal to -541 + 0.99998*5,000 = 4,459 ft^3/s]

Scenario number	Parameters defining the linear relation between the flow occurring under the indicated scenario and the flow under baseline conditions	
	Intercept	Slope
2	-541	0.99998
3	-7	1.000057
4 (100 ft^3/s)	-98	0.999912
4 (300 ft^3/s)	-295	0.999734
4 (500 ft^3/s)	-491	0.999555
4 (1,000 ft^3/s)	-982	0.999109
4 (2,000 ft^3/s)	-1,964	0.998204
5 (300 ft^3/s)	-302	0.999789
5 (500 ft^3/s)	-498	0.999609
5 (1,000 ft^3/s)	-989	0.999160
6 (no pumping)	11	0.999915

The accuracy of these scenario predictions was estimated by calculating the standard deviation of a predicted flow for a given day during water years 1998-99 that was of the same magnitude as the baseline streamflow characteristic for a given combination of scenario, site, and streamflow characteristic.

This standard deviation was calculated as follows (Hill, 1992, p. 28):

$$s_{\hat{y}_m} = \left[\sum_{i=1}^{NP} \sum_{j=1}^{NP} \frac{\partial \hat{y}_m}{\partial b_i} V(b)_{ij} \frac{\partial \hat{y}_m}{\partial b_j} \right]^{\frac{1}{2}} \tag{11}$$

where

$s_{\hat{y}m}$ is the standard deviation of the estimated value of prediction $\hat{y}_m$,

$\frac{\partial \hat{y}_m}{\partial b_i}$ and $\frac{\partial \hat{y}_m}{\partial b_j}$ are the partial derivatives (sensitivities) of $\hat{y}_m$, with respect to model parameters, b_i and b_j, evaluated at the set of optimal model parameter values, b;

$V(b)_{ij}$ is the element in row, i, column, j, of the variance covariance matrix on the parameters (Hill, 1994, p. 57); and

i and j are indexes denoting a particular model parameter.

For example, the streamflow characteristic, 1Q2, under baseline conditions in the Suwannee River near Wilcox is equal to 4,870 ft^3/s, and flow close to this magnitude was simulated at the Wilcox gaging station on April 14, 1999. The standard deviation of the predicted value of 1Q2 under a given scenario was calculated by substituting the sensitivities of the predicted value on this day into equation 11, yielding an approximation of the standard deviation for the predicted value of 1Q2 at the Wilcox gaging station under the given scenario. This process is repeated to obtain estimates of the standard deviation of other combinations of streamflow characteristics, scenarios, and gaging station locations.

Results of these calculations were consistent for all of the scenarios. In each of the possible combinations of scenarios and gaging station location, the standard deviation of the predicted value of a streamflow characteristic ranged from about 8 to 54 ft^3/s. Smaller standard deviation values in this range corresponded to the statistics with the lower flows (1Q2, 1Q5, 1Q10, p10, and p20), and larger values corresponded to the statistics with the larger flows. As a result, values of the coefficient of variation of the predictions (standard deviation of the prediction divided by the simulated value of the prediction) were relatively consistent and generally less than 0.5 percent of the predicted value of each streamflow characteristic under the various scenarios. Results of these calculations for scenario 4 (500 ft^3/s) are given in table 14. As previously stated, results of the other scenarios were similar, so additional tables for the other scenarios are not presented.

Table 14. Standard deviation and coefficient of variation of predicted values of selected streamflow statistics for scenario 4 (500 cubic feet per second) for locations on the Suwannee River corresponding to the gaging stations near Bell and Wilcox, Florida, and above the Gopher River.

[Site locations are shown in fig. 4]

Station name	Streamflow statistic	Standard deviation, in cubic feet per second	Coefficient of variation, in percent
Suwannee River near Bell	1Q2	8.5	0.2
	1Q5	7.8	0.2
	1Q10	7.8	0.3
	p10	15.5	0.1
	p20	33.9	0.3
	p50	15.7	0.2
	p80	9.1	0.2
	p90	8.4	0.2
Suwannee River near Wilcox	1Q2	9.4	0.2
	1Q5	12.0	0.3
	1Q10	13.6	0.4
	p10	24.3	0.1
	p20	53.5	0.4
	p50	18.6	0.2
	p80	9.9	0.2
	p90	8.9	0.2
Suwannee River above the Gopher River	1Q2	8.7	0.2
	1Q5	22.7	0.6
	1Q10	27.6	0.8
	p10	28.7	0.2
	p20	72.5	0.5
	p50	23.9	0.3
	p80	7.8	0.1
	p90	9.3	0.2

Model Limitations

Use of the model for predictive purposes should be limited to estimating how flows in the Lower Suwannee River might change as a result of changes in patterns of ground-water or surface-water withdrawals. The model also should not be used to simulate a stress (for example, a large ground-water withdrawal) that would cause large ground-water level declines along the model boundaries, unless the simulation results are qualified. Possible exceptions include drawdowns that occur near the specified-head boundary condition along the Gulf of Mexico coastline (as long as the simulated ground-water levels along the coast do not fall below NGVD 1929), or drawdowns that occur near the head-dependent flux boundaries at the northeastern limits of the model area. Simulation of this latter case might require the iterative adjustment of the specified heads along the head-dependent flux boundaries until simulated drawdowns reach an equilibrium solution. Another approach would be to: (1) simulate the stress using a larger model (encompassing the area modeled in this report); and (2) use the simulated heads from the larger model to estimate ground-water levels along the head-dependent flux boundaries of the model described in this report. Alternatively, simulation results could be qualified by stating that the results are based on the assumption that locations of flow boundaries and heads at head-dependent flux boundaries are not changed appreciably.

Several other important limitations exist regarding use of the model. One limitation is that the model should not be used (or used with extreme caution) to estimate drawdowns in specific locations in response to hypothetical uses. This limitation arises in large part because sufficient data are not currently available to accurately map the conductive properties of the Upper Floridan aquifer at a detailed level in the subregional study area. Indeed, the work done in this and earlier studies, including the work by Bush and Johnston (1988) and other investigators conducting aquifer tests in the study area, represent the first steps in this mapping process. Because of these limitations, the transmissivity zones in the models are, by necessity, generalized and the locations of the boundaries are subject to considerable uncertainty. For this reason, site-specific drawdown estimates are best made using site-specific aquifer tests at this time.

Another important limitation is that the model has not been tested for use in estimating the time of travel of water or solutes in the Upper Floridan aquifer. Estimates of time of travel and solute transport would require a more detailed understanding of spatial variations in conductive properties and effective porosity of the aquifer. This problem is especially severe in the vicinity of large springs where hydraulic conductivity and porosity are highly variable. Even when the large conduits to springs have been mapped, transport can be strongly influenced by the presence of moderate scale conduits (about 10^{-2} to 10^{-1} ft in diameter), which are almost impossible to map at present. Site-specific testing (using tracer techniques) again provides an interim solution, as well as data necessary for ultimately developing subsurface transport models for the subregional study area. Finally, users of the model should make note of parameter confidence intervals, model sensitivities, and geographic areas where model performance is strongest or weakest to assess the accuracy of any predictions made with the model.

Summary

The Lower Suwannee River and the Upper Floridan aquifer are closely linked hydrologic systems, and predicting the effects of increased water use on ground- and surface-water flows and levels requires a quantitative understanding of the hydraulic behavior of these two systems and their linkages. Historic and newly collected data were evaluated to better understand the variability of water flows and levels in the rivers and aquifers, timing and magnitude of ground- and surface-water exchanges, and physical properties and processes that influence these water flows, levels, and exchanges. A key element of this evaluation was the development of a computer model capable of simulating the ground- and surface-water flows, levels, and exchanges as well as estimating the effects of hypothetical ground- and surface-water withdrawal scenarios on streamflows.

Flow in the Lower Suwannee River is governed by inflows from the Upper Suwannee River, Santa Fe River (its major tributary), and Upper Floridan aquifer. The long-term average flow, as measured at the gaging station near Wilcox, is about 10,000 ft^3/s. Substantial departures from this average flow rate occur over time scales that range from decadal (and longer) to hourly. During the study, El Niño-La Niña climatic conditions produced a wide range of hydrologic conditions in the Lower Suwannee River Basin. High rainfall totals associated with strong El Niño conditions during the winter and early spring of 1998 resulted in a peak daily flow in late March that approached 50,000 ft^3/s (a flow that typically occurs only once every 20 to 25 years). Conversely, a La Niña-associated drought during the rest of water year 1998 and much of water year 1999 resulted in daily flows as low as 3,400 ft^3/s during the late summer-early fall of this year (conditions that also typically occur once every 20 to 25 years). Over the past 60 years, the highest flows and stages in the Lower Suwannee River typically occurred during March and April, and the lowest flows and stages typically occurred from November to January and in June and July. These seasonal flow patterns are strongly influenced by the seasonal climate patterns in the Upper Suwannee River basin and, to a lesser extent, on the climate patterns in the Santa Fe River Basin. Substantial fluctuations in water flows and levels also occur along much of the Lower Suwannee River because of tidal fluctuations near the mouth of the river and the low gradient of the river bottom, which falls (on average) about 0.10 to 0.15 ft/mi of river. The magnitude of these tidally induced fluctuations diminishes with increasing (mean) flow in the river and with increasing upstream distance from the Gulf of Mexico.

With the exception of the Lower Suwannee and Santa Fe Rivers and some of the coastal areas, channelized surface drainage is largely absent in the subregional study area. Drainage generally occurs through the Upper Floridan aquifer, which is a thick sequence of carbonate rocks subject to a considerable amount of dissolution. This dissolution accounts for the presence of karst features, such as numerous springs and closed depressions, and a generally high transmissivity in the subregional study area. Long-term average recharge to the aquifer is probably between 14 and 24 in/yr, although estimated recharge rates were two to three times this amount during water year 1998 because of wetter-than-normal conditions. Total ground-water withdrawals over the subregional study area were estimated to average about 35 ft^3/s during water years 1998 and 1999. Estimates of minimum and maximum monthly withdrawal rates during this period were 5 and 155 ft^3/s, respectively, with the minimum rates occurring in January and the maximum rates (associated with peak levels of agricultural irrigation) occurring in May.

Data collected before and during the study were evaluated to estimate the magnitude and timing of exchanges of water between the Lower Suwannee and Lower Santa Fe Rivers and the Upper Floridan aquifer. Long-term net-ground-water discharge to much of these two rivers was estimated by calculating the change in discharge between three gaging stations (Suwannee River at Branford, Suwannee River near Wilcox, and Lower Santa Fe River near Fort White), using historic streamflow data collected during climate years 1942-2000. These data indicate that net ground-water discharge to this reach of the Lower Suwannee and Santa Fe Rivers (designated as reach 1 in this study) averages about 1,500 ft^3/s, with about half the years during this period having estimated ground-water discharge rates between 1,200 and 1,900 ft^3/s. The annual ground-water discharge estimates for reach 1 exhibited a substantial amount of interannual correlation, which is expected given the longer response time (relative to the river) of the adjacent ground-water system.

An attempt to estimate ground- and surface-water exchanges in reach 1 over short time scales was made by correcting the change in discharge calculations for channel and floodplain storage. The method yielded estimates of annual ground-water discharge of 1,000 and 1,100 ft^3/s for water years 1998 and 1999, respectively. Estimates were also computed at seasonal, monthly, and daily time scales. In all instances (estimates from annual to daily time scales), however, the estimated values had wide confidence intervals, because the uncertainty of the basic daily discharge data from the three gaging stations were typically similar in magnitude to the estimated ground-water discharge. Stated differently, in most instances, it was not possible to determine if the day-to-day (or to a lesser extent, year to year) variations in ground-water discharge were real or due to uncertainty associated with estimating the daily discharge values at the gaging stations. For these reasons, estimating ground- and surface-water exchanges by analyzing changes in discharge between gaging stations was considered suitable only for estimating the long-term characteristics of these exchanges (for example, the long-term, annual mean ground-water discharge to reach 1).

Analyses of historic spring-flow data and streamflow measurements taken during the study indicated that some river reaches received more ground-water inflow than others. The most concentrated areas of ground-water discharge coincided with large magnitude springs, which were commonly clustered. These concentrated areas of spring flow are located in the upper reaches of the Ichetucknee River (a tributary to the Santa Fe River), the Santa Fe River just upstream from the gaging station near Fort White, and the Suwannee River near Fanning Springs and Manatee Springs. Springs account for a large percentage of ground-water discharge to the Lower Suwannee and Lower Santa Fe Rivers (for example, about 55 to 85 percent of the discharge in reach 1). Discharge measurements taken during the study indicated that the ground-water discharge to the Suwannee River between Branford and Rock Bluff Springs (between river miles 76 and 56) was small or negligible. Tidally induced variations in spring flow in Manatee Springs and Fanning Springs (as much as ±6 and ±12 percent of the mean spring flow, respectively) during the tidal cycle were also observed.

A coupled model of river and aquifer flows and levels was developed to help evaluate ground- and surface-water exchanges and to determine how streamflows might change in response to changes in surface-water or ground-water use. The model consisted of two separate submodels: (1) ground-water flow in the Upper Floridan was simulated using the MODFLOW model, and (2) water flows and levels in the Lower Suwannee and Lower Santa Fe Rivers were simulated using the BRANCH model. Coupling of the two models was implemented using the MODBRANCH code. The model was calibrated to the wide range of hydrologic conditions that occurred from October 1997 to September 1999.

The simulated values of net ground-water inflow to reach 1 averaged about 1,350 ft^3/s during water years 1998 and 1999. In the Suwannee River downstream from the mouth of the Santa Fe River (reach 3), simulated average ground-water inflows were about 2,000 ft^3/s during the same time period. Simulated monthly ground-water inflows to this reach of the Suwannee River ranged from 1,500 to 3,200 ft^3/s, with the lower value occurring in November 1997 (during the early part of the winter-early spring flood period) and the higher value occurring during April 1998 (following the winter flood peak). Lower rates of simulated ground-water inflows also occurred during September 1999 after an extended period of drought. Abrupt changes in ground-water discharge were simulated in the vicinity of hydrograph peaks. For example, the simulated daily ground-water inflow to the Suwannee River downstream from the Santa Fe River changed from about 750 ft^3/s near the time of peak flooding on the Lower Suwannee River during March 1998 to 3,800 ft^3/s at about 20 days after the peak.

These patterns in the timing of ground-water inflow variations also were evident in the simulated daily values of ground-water discharge to the Lower Suwannee River. Low rates of simulated ground-water discharge occurred during periods when the hydraulic gradient between the aquifer and river was low, either because river levels had risen faster than ground-water levels or because of extended droughts, which reduced aquifer storage. Conversely, high rates of ground-water discharge generally occurred shortly after flood peaks when river levels fell and ground-water levels (aquifer storage) increased as a result of higher recharge rates and the restriction of ground-water discharge (because of high river levels) in the preceding wet period.

Several hypothetical water-withdrawal scenarios were simulated with the model to evaluate the effects of water-use changes on long-term streamflow statistics. The characteristics for the scenarios that were evaluated include: (1) increased ground-water withdrawals, (2) reduced upstream river inflows to the Lower Suwannee and Lower Santa Fe Rivers, (3) a point withdrawal from the Suwannee River, and (4) combinations of these types of withdrawals. Changes in the streamflow statistics were calculated by developing linear relations between the daily flows that were simulated for a given scenario and the daily flows simulated during water years 1998-99. Sensitivity analyses indicated that the uncertainties of the predictions resulting from uncertainties in the transmissivity and riverbed conductance parameters used in the model were relatively small.

Use of the model for predictive purposes should generally be limited to estimating how flows in the Lower Suwannee River might change in response to changing patterns of water withdrawals. Simulations that produce large drawdowns near lateral no-flow boundaries should be avoided (or conducted with caution) and presented with appropriate qualifiers. The locations of the boundaries of the transmissivity zones used in the model are subject to considerable uncertainty and should be interpreted as representing only generalized patterns of transmissivity variations. For this reason, the model should not be used to make site-specific estimates of pumping-induced ground-water declines. Instead, these questions are best addressed using site-specific aquifer pumping tests. Despite these limitations, the model should provide useful information for evaluating future water withdrawals in the subregional study area, as well as providing a framework for: (1) evaluating changes in data-collection programs that integrate new water-level, flow, and aquifer property data; and (2) building enhanced models capable of answering new questions.

Selected References

Burnson, Terry, Shoemyen, J.L., Cameron, J.R., Webster, K.B., Oxford, L.C., Jr., Ceryak, Ron, Copeland, R.E., Leadon, C.J., and Batchelder, Pat, 1984, Suwannee River Water Management District, *in* Fernald, E.A., and Patton, D.J., eds.,Water Resources Atlas of Florida: Tallahassee, Florida State University, Institute of Science and Public Affairs, 291 p.

Bush, P.W., and Johnston, R.H., 1988, Ground-water hydraulics, regional flow, and ground-water development of the Floridan aquifer system in Florida and in parts of Georgia, South Carolina, and Alabama: U.S. Geological Survey Professional Paper 1403-C, 80 p.

Butt, P.L., Hayes, A.W., Morris, T.L., and Skiles, W.C., 2000, Results of the Rose Creek Swallet to Ichetucknee Springs dye trace study August-September 1997: Abstract from Florida Springs Conference: Natural Gems—Troubled Waters, Gainesville, Florida.

Chow, V.T., Maidment, D.R., and Mays, L.W., 1988, Applied hydrology: New York, McGraw-Hill, 572 p.

Clark, G.A., and Stanley, C.D., 1994, Using reference evapotranspiration data: Gainesville, University of Florida Institute of Food and Agricultural Sciences, Fact Sheet AE-251.

Clarke, W.E., 1965, Relation of ground-water inflow and of bank and channel storage to streamflow pickup in the Santa Fe River, Florida, *in* U.S. Geological Survey Professional Paper 525-D, p. D211-D113.

Col, Nolan, Rupert, Frank, Enright, Meryl, and Horvath, Glen, 1997, Reappraisal of the geology and hydrogeology of Gilchrist County, Florida, with emphasis on the Waccasassa Flats: Tallahassee, Florida Geological Survey Report of Investigations 99, 76 p.

Cooper, H.H., 1959, A hypothesis concerning the dynamic balance of fresh water and salt water in a coastal aquifer: Journal of Geophysical Research, v. 64, no. 4, p. 461-467.

Countryman, R.A., II, and Stewart, M.T., 1997, Geophysical delineation of the position of the saltwater interface in the Lower Suwannee River Basin: Final contract report for Suwannee River Water Management District and U.S. Geological Survey, 538 p.

Crandall, C.A., Katz, B.G., and Hirten, J.J., 1999, Hydrochemical evidence for mixing of river water and groundwater during high-flow conditions, Lower Suwannee River Basin, Florida, USA: Hydrogeology Journal, v. 7, p. 454-467.

Crane, J.J., 1986, An investigation of the geology, hydrogeology, and hydrochemistry of the Lower Suwannee River Basin: Tallahassee, Florida Geological Survey Report of Investigations 96, 205 p.

Cunge, J.A., Holly, F.M., Jr., and Verwey, A., 1980, Practical aspects of computational river hydraulics: London, Pittman Publishing, 420 p.

Darst, M.R., Light, H.M., Lewis, L.J., 2002, Ground-cover vegetation in wetland forests of the Lower Suwannee River Floodplain, Florida, and potential impacts of flow reductions: U.S. Geological Survey Water-Resources Investigations Report 02-4027, 46 p. Accessible online at *http://fl.water.usgs.gov/Abstracts/wri02_4027_darst.html*

Davis, J.H., 1996, Hydrogeologic investigation and simulation of ground-water flow in the Upper Floridan aquifer of north-central Florida and southwestern Georgia and delineation of contributing areas for selected City of Tallahassee, Florida, water-supply wells: U.S. Geological Survey Water-Resources Investigations Report 95-4296, 56 p. Accessible online at *http://fl.water.usgs.gov/Abstracts/wri95_4296_davis.html*

Doherty, John, 2000, PEST model-independent parameter estimation users guide (4th ed.): Watermark Numerical Computing, 230 p.

Ellins, K.K., Kincaid, T.R., Hisert, R.A., Johnson, N.A., Davison, C.A., and Wanninkohf, R.H., 1991, Using ^{222}Rn and SF6 to determine groundwater gains and stream low losses in the Santa Fe River: Hydrogeology of the western Santa Fe River basin: Southeastern Geological Society, Field Trip Guidebook 32.

Faulkner, G.L., 1973, Geohydrology of the Cross-Florida Barge Canal area, with special reference to the Ocala vicinity: U.S. Geological Survey Water-Resources Investigations 73-1, 117 p.

Fenneman, N.M., 1938, Physiography of eastern United States: New York, McGraw Hill, 691 p.

Fetter, C.W., 1988, Applied hydrology: Columbus, Ohio, Merrill Publishing Co., 592 p.

Florida Geological Survey, 2004, Florida Geological Survey lithologic database, data and interfacing program: Tallahassee: Accessible online at *http://www.dep.state.fl.us/geology/gisdatamaps/litholog.htm*

Franz, D.D., and Melching, C.S., 1997, Full Equations Utilities (FEQUTL) model for the approximation of hydraulic characteristics of open channels and control structures during unsteady flow: U.S. Geological Survey Water-Resources Investigations Report 97-4037, 205 p. Accessible online at *http://il.water.usgs.gov/proj/feq/fequtl/fequtl.toc_1.html*

Grubbs, J.W., 1997, Recharge rates to the Upper Floridan aquifer in the Suwannee River Water Management District, Florida: U.S. Geological Survey Water-Resources Investigations Report 97-4283, 30 p. Accessible online at *http://pubs.er.usgs.gov/usgspubs/wri/wri974283*

Harbaugh, A.W., and McDonald, M.G., 1996, User's documentation for MODFLOW-96, an update to the U.S. Geological Survey Modular Finite-Difference Ground-Water Flow Model: U.S. Geological Survey Open-File Report 96-485, 56 p. Accessible online at *http://pubs.er.usgs.gov/usgspubs/ofr/ofr96485*

Helsel, D.R., and R.M. Hirsch, 1992, Statistical methods in water resources: Studies in Environmental Science 49, Elsevier, 522 p.

Hill, M.C., 1990, Preconditioned conjugate-gradient 2 (PCG2), a computer program for solving ground-water flow equations: U.S. Geological Survey Water-Resources Investigations Report 90-4048, 43 p. Accessible online at *http://pubs.usgs.gov/wri/wrir_90-4048/*

Hill, M.C., 1992, A computer program (MODFLOWP) for estimating parameters of a transient, three-dimensional, ground-water flow model using nonlinear regression: U.S. Geological Survey Open-File Report 91-484, 358 p. Accessible online at *http://pubs.er.usgs.gov/usgspubs/ofr/ofr91484*

Hill, M.C., 1994, Five computer programs for testing weighted residuals and calculating linear confidence and prediction intervals on results from the ground-water parameter-estimation computer program MODFLOWP: U.S. Geological Survey Open-File Report 93-481, 81 p. Accessible online at *http://pubs.er.usgs.gov/usgspubs/ofr/ofr93481*

Hill, M.C., 1998, Methods and guidelines for effective model calibration: U.S. Geological Survey Water-Resources Investigations Report 98-4005, 90 p. Accessible online at *http://water.usgs.gov/nrp/gwsoftware/modflow2000/modflow2000.html*

Hisert, R.A., 1994, A multiple tracer approach to determine the ground water and surface water relationships in the western Santa Fe River, Columbia County, Florida: Unpublished Ph.D. Dissertation, Department of Geology, University of Florida, Gainesville.

Hornsby, David, and Ceryak, Ron, 1998, Springs of the Suwannee River Basin in Florida: Live Oak, Fla., Suwannee River Water Management District Water Resources Publication 99-02, 178 p.

Hull, R.W., Dysart, J.E., and Mann, IV, W.B., 1981, Quality of the surface water in the Suwannee River Basin, Florida, August 1968 through December 1977: U.S. Geological Survey Water-Resources Investigations 80-110, 97 p.

Hunn, J.D., and Slack, L.J., 1983, Water resources of the Santa Fe River Basin, Florida: U.S. Geological Survey Water-Resources Investigations Report 93-4075, 105 p. Accessible online at *http://pubs.er.usgs.gov/usgspubs/wri/wri934075*

Hydrologic Engineering Center, 1991, HEC-2, Water surface profiles, users's manual: Davis, Calif., U.S. Army Corps of Engineers.

Jones, J.W., Allen, L.H., Shih, S.F., Rogers, J.S., Hammond, L.C., Smajstrala, A.G., and Martsolf, J.D., 1984, Estimated and measured evapotranspiration for Florida climate, crops, and soils: Gainesville, University of Florida Institute of Food and Agricultural Sciences, Bulletin 840, 65 p.

Katz, B.G., 1998, Using delta 18O and delta D to quantify ground-water/surface-water interactions in karst systems of Florida, *in* Monitoring: Critical Foundations to Protect Our Waters, Proceedings of National Water Quality Monitoring Council National Conference: Washington, D.C., U.S. Environmental Protection Agency, p. 195-207.

Katz, B.G., and Catches, J.S., 1996, The Little River Basin study area—Interactions between ground water and surface water, *in* Winkler, W., and Davis, K., compilers, Surface water and groundwater interaction along the Cody Scarp transition region of the Suwannee River Basin near Live Oak, Florida: Southeastern Geological Society, Field Trip Guidebook 36, p. 22-28.

Katz, B.G., Catches, J.S., Bullen, T.D., and Michel, R.L., 1998, Changes in the isotopic and chemical composition of ground water resulting from a recharge pulse from a sinking stream: Journal of Hydrology, v. 211, p. 178-207.

Katz, B.G., DeHan, R.S., Hirten, J.J., and Catches, J.S., 1997, Interactions between ground water and surface water in the Suwannee River Basin, Florida: Journal of the American Water Resources Association, v. 33, no. 6, p. 1237-1254.

Kincaid, T.R., 1998, River water intrusion to the unconfined Floridan aquifer: Environmental and Engineering Geoscience, v. IV, no. 3, p. 361-374.

Knowles, Leel, Jr., 1996, Estimation of evapotranspiration in the Rainbow Springs and Silver Springs basins in north-central Florida: U.S. Geological Survey Water-Resources Investigations Report 96-4024, 37 p. Accessible online at *http://pubs.er.usgs.gov/usgspubs/wri/wri964024*

Krause, R.E., and Randolph, R.B., 1989, Hydrology of the Floridan aquifer system in southeast Georgia and adjacent parts of Florida and South Carolina: U.S. Geological Survey Professional Paper 1403-D, 65 p. Accessible online at *http://pubs.er.usgs.gov/usgspubs/pp/pp1403D*

Lee, Paul, and Passehl, John, 1995, Delineation of ground and surface water areas potentially impacted by an industrial discharge to the Fenholloway River of Taylor County, Florida: Tallahassee, Florida Department of Environmental Protection, Bureau of Drinking Water and Ground Water Resources, 140 p.

Lee, T.M., 1996, Hydrogeologic controls on the groundwater interactions with an acidic lake in karst terrain, Lake Barco, Florida: Water Resources Research, v. 32, no. 4, p. 831-844.

Light, H.M., Darst, M.R., and Lewis, L.J., 2002, Hydrology, vegetation, and soils of riverine and tidal floodplain forests of the Lower Suwannee River, Florida, and potential impacts of flow reductions: U.S. Geological Survey Professional Paper 1656-A, 124 p. Accessible online at *http://fl.water.usgs.gov/Abstracts/pp1656A_light.html*

Miller, J.A., 1986, Hydrogeological framework of the Floridan aquifer system in Florida, and in parts of Georgia, Alabama, and South Carolina: U.S. Geological Survey Professional Paper 1403-B, 91 p. Accessible online at *http://pubs.er.usgs.gov/usgspubs/pp/pp1403B*

Miller, J.A., Hughes, G.H., Hull, R.W., Vecchioli, John, and Seaber, P.R., 1978, Impact of potential phosphate mining on the hydrology of Osceola National Forest, Florida: U.S. Geological Survey Water-Resources Investigations Report 78-6, 159 p. Accessible online at *http://pubs.er.usgs.gov/usgspubs/wri/wri786*

McDonald, M.G., and Harbaugh, A.W., 1988, A modular three-dimensional finite-difference ground-water flow model: U.S. Geological Survey Techniques of Water-Resources Investigations, book 6, chap. A1, 586 p.

National Oceanographic and Atmospheric Administration, 2002, Climatography of the United States No. 85: Division normals and standard deviations of temperature, precipitation, and heating and cooling degree days 1971-2000 (and previous normals periods), National Climatic Data Center/National Environmental Satellite, Data, and Information Service/National Oceanographic and Atmospheric Administration: Asheville, N.C. Accessible from *http://www.ncdc.noaa.gov*

National Oceanographic and Atmospheric Administration, 2004, Center for Operational Oceanographic Products and Services, Verified / Historical Water Level Data, U.S. and Global Coastal Stations, accessed from *http://co-ops.nos.noaa.gov/data_retrieve.shtml?input_code=100111111vwl* (September 28, 2004).

National Oceanographic and Atmospheric Administration, 2005, The ENSO cycle: National Weather Service, Climate Prediction Center, accessed from *http//www.cpc.ncep.noaa.gov/products/analysis_monitoring/ensocycle/enso_cycle.shtml* (modified December 19, 2005).

Nigam, Sumant, Barlow, Mathew, and Berbery, E.H., 1999, Analysis links Pacific decadal variability to drought and streamflow in United States: EOS, v. 80, no. 61, p. 621-625.

Pittman, J.R., Hatzell, H.H., and Oaksford, E.T., 1997, Spring contributions to water quantity and nitrate loads in the Suwannee River during base flow in July 1995: U.S. Geological Survey Water-Resources Investigations Report 97-4152, 12 p. Accessible online at *http://fl.water.usgs.gov/Abstracts/wri97_4152_pittman.html*

Prudic, D.E., 1989, Documentation of a computer program to simulate stream-aquifer relations using a modular, finite-difference, ground-water flow model: U.S. Geological Survey Open-File Report 88-729, 113 p. Accessible online at *http://pubs.er.usgs.gov/usgspubs/ofr/ofr88729*

Puri, H.S., and Vernon, R.O., 1964, Summary of the geology of Florida and a guidebook to the classic exposures: Tallahassee, Florida Geological Survey Special Publication 5, 312 p.

Puri, H.S., Yon, J.W., Jr., and Oglesby, W.R., 1967, Geology of Dixie and Gilchrist Counties, Florida: Tallahassee, Florida Geological Survey Geological Bulletin 49, 155 p.

Raulston, Marvin, Johnson, Cindy, Webster, Kirk, Purdy, Carolyn, and Ceryak, Ron, 1998, Suwannee River Water Management District, *in* Fernald, E.A., and Purdum, E.D., eds., Water Resources Atlas of Florida: Tallahassee, Florida State University, Institute of Science and Public Affairs, chap. 10, p. 194-213.

Riggs, H.C., 1972, Low-flow investigations: U.S. Geological Survey Techniques of Water-Resources Investigations, book 4, chap. B1, 18 p.

Rosenau, J.C., Faulkner, G.L., Hendry, C.W., and Hull, R.W., 1977, Springs of Florida: Tallahassee, Florida Bureau of Geology Bulletin 31, 460 p.

Rupert, F.R., 1988a, Geology and geomorphology of Gilchrist County, Florida: Tallahassee, Florida Geological Survey Open-File Report 18, 16 p.

Rupert, F.R., 1988b, Geology and geomorphology of Levy County, Florida: Tallahassee, Florida Geological Survey Open-File Report 19, 18 p.

Ryder, P.D., 1985, Hydrology of the Floridan aquifer system in west-central Florida: U.S. Geological Survey Professional Paper 1403-F, 63 p. Accessible online at *http://pubs.er.usgs.gov/usgspubs/pp/pp1403F*

Schaffranek, R.W, 1987, Flow model for open-channel reach or network: U.S. Geological Survey Professional Paper 1384, 11 p. Accessible online at *http://pubs.er.usgs.gov/usgspubs/pp/pp1384*

Schaffranek, R.W., Baltzer, R.A., and Goldberg, D.E., 1981, A model for simulation of flow in singular and interconnected channels: U.S. Geological Survey Techniques of Water-Resources Investigations, book 7, chap. C3, 110 p.

Scott, T.M., 1992, A geologic overview of Florida: Tallahassee, Florida Geological Survey Open-File Report 50, 78 p.

Scott, T.M., Means, G.H., Meegan, R.P., Means, R.C., Upchurch, S.B., Copeland, R.E., Jones, James, Roberts, Tina, and Willet, Alan, 2004, Springs of Florida: Tallahassee, Florida Geological Survey Bulletin 66, 377 p.

Sepúlveda, Nicasio, 2002, Simulation of ground-water flow in the intermediate and Floridan aquifer systems in peninsular Florida: U.S. Geological Survey Water-Resources Investigations Report 02-4009, 130 p. Accessible online at *http://fl.water.usgs.gov/Abstracts/wri02_4009_sepulveda.html*

Southeastern Geological Society Ad Hoc Committee on Florida Hydrostratigraphic Unit Definition, 1986, Hydrogeological units of Florida: Tallahassee, Florida Geological Survey Special Publication 28, 8 p.

Southwest Florida Water Management District, 2000, Aquifer characteristics within the Southwest Florida Water Management District: Brooksville, Southwest Florida Water Management District Technical Report 99-1, 122 p.

Sumner, D.M., 2001, Evapotranspiration from a cypress and pine forest subjected to natural fires in Volusia County, Florida, 1998-1999: U.S. Geological Survey Water-Resources Investigations Report 01-4245, 56 p. Accessible online at *http://fl.water.usgs.gov/Abstracts/wri01_4245_sumner.html*

Swain, E.D., and Wexler, E.J., 1996, A coupled surface-water and ground-water flow model (MODBRANCH) for simulation of stream-aquifer interaction: U.S. Geological Survey Techniques of Water-Resources Investigations, book 6, chap. A6, 125 p.

Tibbals, C.H., 1990, Hydrology of the Floridan aquifer system in east-central Florida: U.S. Geological Survey Professional Paper 1403-E, 98 p. Accessible online at *http://pubs.er.usgs.gov/usgspubs/pp/pp1403E*

U.S. Army Corps of Engineers, 1989, Suwannee River, Georgia and Florida survey-review report: Jacksonville, Fla. District.

U.S. Geological Survey, 1998, National Water Information System (NWIS): U.S. Geological Survey Fact Sheet FS-027-98 Accessible from *http://water.usgs.gov/pubs/fs/FS-027-98*

Veri, George, DuChene, Harvey, Crawford, N.C., Groves, D.G., Huppert, G.N., Kastning, E.H., Olson, Rick, and Wheeler, B.J., 2001, Living with karst: A fragile foundation: American Geological Institute, ISBN 0-922152-58-6, 64 p.

Vernon, R.O., 1951, Geology of Citrus and Levy Counties, Florida: Tallahassee, Florida Geological Survey Bulletin 33, 256 p.

Vorhis, R.C., 1961, A hydrogeologic reconnaissance of reservoir possibilities in northern Lowndes County, Georgia: Georgia Mineral Newsletter, v. 14, no. 4, p. 123-129.

White, W.A., 1958, Some geomorphic features of central peninsular Florida: Tallahassee, Florida Geological Survey Geological Bulletin 41, 92 p.

White, W.A., 1970, The geomorphology of the Florida Peninsula: Tallahassee, Florida Geological Survey Geological Bulletin 51, 164 p.

Wilson, W.L. and Skiles, W.C., 1988, Aquifer characterization by quantitative dye tracing at Ginnie Spring, northern Florida, *in* Proceedings of the Second Conference on Environmental Problems in Karst Terranes and their Solutions: Nashville, Tennessee, Association of Groundwater Scientists and Engineers: Dublin, Ohio, National Water Well Association, p. 121-141.

Yon, J.W., Jr., and Puri, H.S., 1962, Geology of Waccasassa Flats, Gilchrist County, Florida: American Association of Petroleum Geologists Bulletin, v. 46, no. 5, p. 674-684.