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Ground-Water and Surface-Water Flow and Estimated Water Budget for Lake Seminole, Southwestern Georgia and Northwestern Florida

By Melinda S. Dalton, Brent T. Aulenbach, and Lynn J. Torak

Prepared in cooperation with the
Georgia Department of Natural Resources
Environmental Protection Division
Georgia Geologic Survey

Scientific Investigations Report 2004-5073

U.S. Department of the Interior
U.S. Geological Survey

Cover photograph: Polk Lake, Chattahoochee, Florida, 1950

View looking east along the fixed-crest spillway during construction of Jim Woodruff Lock and Dam. During 2002, the U.S. Army Corps of Engineers and the U.S. Geological Survey participated in a dye-tracing study that found lake water moving through the ground-water system to the Apalachicola River via Polk Lake Spring.

Photograph by: U.S. Army Corps of Engineers, archives, Mobile District, Mobile, Alabama

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Ground-Water and Surface-Water Flow and Estimated Water Budget for Lake Seminole, Southwestern Georgia and Northwestern Florida

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Abstract

Lake Seminole is a 37,600-acre impoundment formed at the confluence of the Flint and Chattahoochee Rivers along the Georgia–Florida State line. Outflow from Lake Seminole through Jim Woodruff Lock and Dam provides headwater to the Apalachicola River, which is a major supply of freshwater, nutrients, and detritus to ecosystems downstream. These rivers, together with their tributaries, are hydraulically connected to karst limestone units that constitute most of the Upper Floridan aquifer and to a chemically weathered residuum of undifferentiated overburden.

The ground-water flow system near Lake Seminole consists of the Upper Floridan aquifer and undifferentiated overburden. The aquifer is confined below by low-permeability sediments of the Lisbon Formation and, generally, is semiconfined above by undifferentiated overburden. Ground-water flow within the Upper Floridan aquifer is unconfined or semiconfined and discharges at discrete points by springflow or diffuse leakage into streams and other surface-water bodies. The high degree of connectivity between the Upper Floridan aquifer and surface-water bodies is limited to the upper Eocene Ocala Limestone and younger units that are in contact with streams in the Lake Seminole area. The impoundment of Lake Seminole inundated natural stream channels and other low-lying areas near streams and raised the water-level altitude of the Upper Floridan aquifer near the lake to nearly that of the lake, about 77 feet.

Surface-water inflow from the Chattahoochee and Flint Rivers and Spring Creek and outflow to the Apalachicola River through Jim Woodruff Lock and Dam dominate the water budget for Lake Seminole. About 81 percent of the total water-budget inflow consists of surface water; about 18 percent is ground water, and the remaining 1 percent is lake precipitation. Similarly, lake outflow consists of about 89 percent surface water, as flow to the Apalachicola River through Jim Woodruff Lock and Dam, about 4 percent ground water, and about 2 percent lake evaporation. Measurement error and uncertainty in flux calculations cause a flow imbalance of about 4 percent between inflow and outflow water-budget components. Most of this error can be attributed to errors in estimating ground-water discharge from the lake, which was calculated using a ground-water model calibrated to October 1986 conditions for the entire Apalachicola–Chattahoochee–Flint River Basin and not just the area around Lake Seminole.

Evaporation rates were determined using the preferred, but mathematically complex, energy budget and five empirical equations: Priestley-Taylor, Penman, DeBruin-Keijman, Papadakis, and the Priestley-Taylor used by the Georgia Automated Environmental Monitoring Network. Empirical equations require a significant amount of data but are relatively easy to calculate and compare well to long-term average annual (April 2000–March 2001) pan evaporation, which is 65 inches. Calculated annual lake evaporation, for the study period, using the energy-budget method was 67.2 inches, which overestimated long-term average annual pan evaporation by 2.2 inches. The empirical equations did not compare well with the energy-budget method during the 18-month study period, with average differences in computed evaporation using each equation ranging from 8 to 26 percent. The empirical equations also compared poorly with long-term average annual pan evaporation, with average differences in evaporation ranging from 3 to 23 percent. Energy budget and long-term average annual pan evaporation estimates did compare well, with only a 3-percent difference between estimates. Monthly evaporation estimates using all methods ranged from 0.7 to 9.5 inches and were lowest during December 2000 and highest during May 2000. Although the energy budget is generally the preferred method, the dominance of surface water in the Lake Seminole water budget makes the method inaccurate and difficult to use, because surface water makes up more than two-thirds of the energy budget and errors in measured streamflow can be substantial.

Introduction

Lake Seminole is a 37,600-acre lake located in the lower Apalachicola–Chattahoochee–Flint (ACF) River Basin at the border between southwestern Georgia and northwestern Florida (fig. 1). The lake was created from the late 1940s to mid-1950s with the construction of Jim Woodruff Lock and Dam on the Apalachicola River, about 1,000 feet (ft) downstream of the confluence of the Flint and Chattahoochee Rivers (U.S. Army Corps of Engineers and others, 1948; James H. Sanders, Jr., U.S. Army Corps of Engineers, Mobile, Alabama, written commun., 2002). The lake was intended to aid navigation on the Chattahoochee and Flint Rivers, provide lift for hydroelectric-power generation, and serve recreational purposes. Lake Seminole is the source of headwater to the Apalachicola River and is the major supply of freshwater, nutrients, and detritus to ecosystems downstream, including Apalachicola Bay and its estuaries. Despite its size, Lake Seminole is a run-of-the-river impoundment, having less than 67,000 acre-feet (acre-ft) of useful storage. About 240 miles (mi) of shoreline encompass four impoundment arms: two major arms extend the lake from the dam to about 47 mi upstream along the natural courses of the Chattahoochee and Flint Rivers; two minor impoundment arms are present along Fishpond Drain and Spring Creek, both of which are tributaries to the Flint River arm of the lake (U.S. Army Corps of Engineers, 1980).

Recently, Lake Seminole and the water released from it have become focal points in water-allocation negotiations—resulting from the ACF River Basin Compact[1]—between Georgia, Florida, and Alabama. Increases in population, agriculture, and industry and the recent drought of 1998–2002 have made water supply and use in the lower ACF River Basin a major concern for water-resource managers in the region, as the three States place conflicting demands on the basin's limited water resources. These concerns led the three States to sign an interstate water compact in 1997, which intended to ensure the equitable use and availability of water resources in the region while protecting river ecology.

Essential to the State of Georgia's water-allocation plans was the necessity to undertake a technical study to develop a comprehensive water budget of the Lake Seminole area, to reasonably estimate the volume of water flowing into Florida before and after construction of the dam, and to monitor the effects of any sinkhole collapse within the lake (Harold F. Reheis, Director, Georgia Department of Natural Resources, Environmental Protection Division, written commun., 1997). The State of Georgia requested that the U.S. Geological Survey (USGS) conduct a technical study to address these issues; and the following four objectives were developed for a 3-year study, which began during 1999.

- Develop a water budget for Lake Seminole that will promote a reasonable understanding of the effect of the lake on the overall flow system in the lower ACF River Basin, and that can be used to guide water allocations between Alabama, Florida, and Georgia.

- Compare current and pre-Lake Seminole ground-water and surface-water flow regimes to determine whether the volume of water flowing out of Georgia has changed substantially after construction of Jim Woodruff Lock and Dam and the filling of the lake.

- Evaluate the possibility of a substantial amount of water entering the ground-water regime from Lake Seminole, flowing beneath Jim Woodruff Lock and Dam, and entering Florida downstream of the dam.

- Assess the likelihood of failure of dissolution features in the karst limestone of the lake bottom, such as sinkhole collapse, and the likelihood of sudden partial or complete draining of the lake. If such an occurrence is likely, then propose a data-collection system to monitor pertinent hydrologic features that could lead to a sudden draining of Lake Seminole, and could provide a warning of its occurrence.

The 3-year study investigated features of the hydrologic system near Lake Seminole that contribute directly to the surface- and ground-water flow regime of the lake. The study focused on only those elements of the hydrologic cycle, surface-water features, and hydrogeologic units that are in hydraulic connection with the lake. A multidiscipline investigative approach was used that involved acquisition of chemical, limnologic, hydrogeologic, and meteorologic information, followed by analysis, interpretation, and reporting of the resulting data and corresponding uncertainty.

Purpose and Scope

This report is one of a series of reports documenting a study to evaluate the effects of impoundment of Lake Seminole on water resources in the lower ACF River Basin. The first of the four study objectives listed previously is addressed herein, namely, to develop a water budget for Lake Seminole that will promote a reasonable understanding of the effects of the lake on the overall flow system in the lower ACF River Basin, and that can be used to guide water allocations between Alabama, Florida, and Georgia. The following technical tasks were performed to achieve this objective and are discussed in this report:

[1]As adopted by: the Alabama Legislature on February 18, 1997, and signed by the Governor of Alabama on February 25, 1997, as Alabama Acts 97-67, Alabama Code, Title 33-19-1 *et seq*.; the Florida Legislature on April 14, 1997, and signed by the Governor of Florida on April 24, 1997, as Chapter 97-25, Laws of Florida, Section 373.71, Florida Statutes (1997); the Georgia Legislature on February 11, 1997, as Georgia Acts No. 7, and signed by the Governor of Georgia on February 25, 1997, as Georgia Code Annual Section 12-10-100 *et seq*., and passed by the United States Congress on November 7, 1997, and signed by the President of the United States on November 20, 1997, as Public Law Number 105-104, 111 Statute 2219

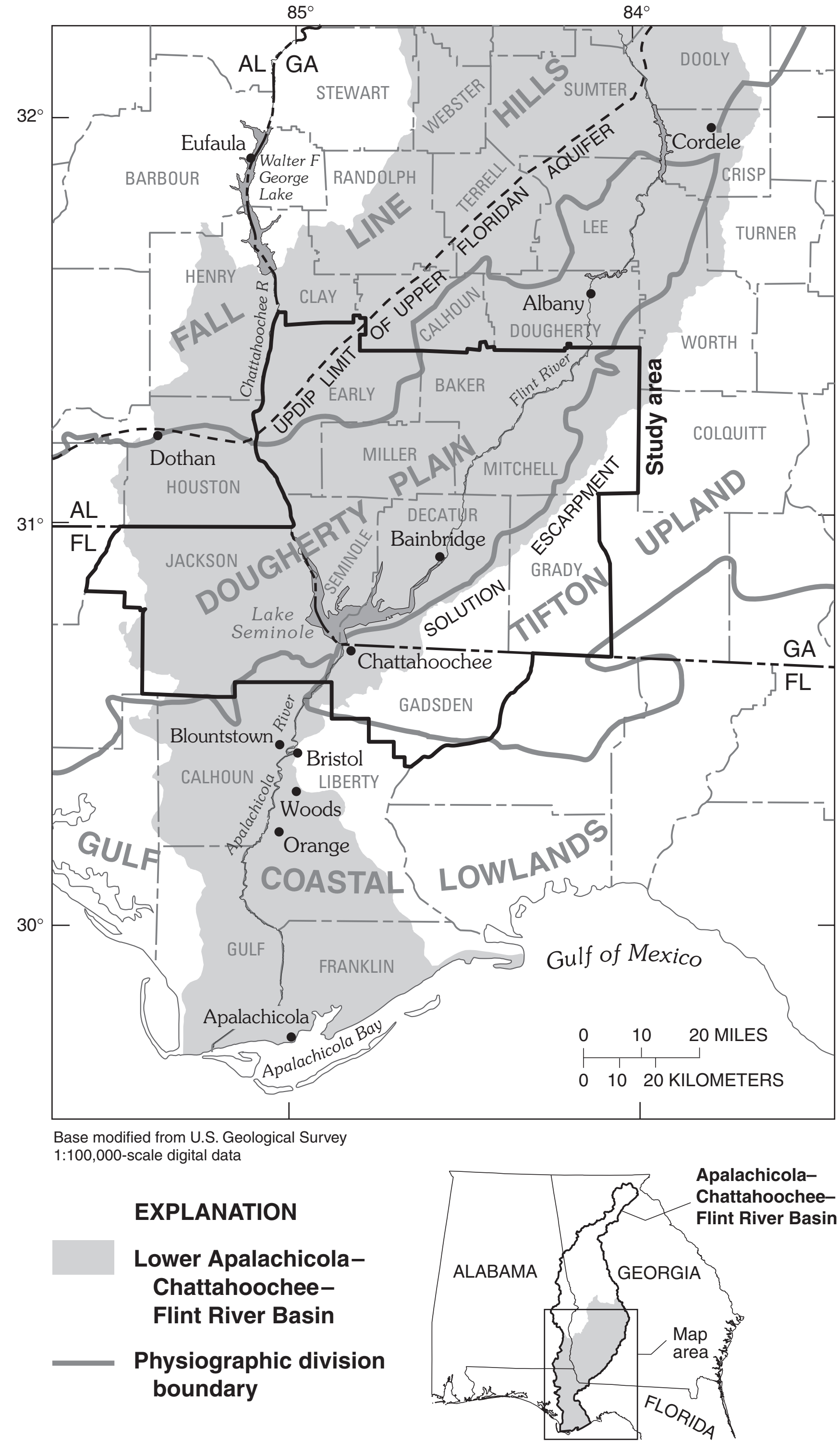


Figure 1. Location of study area, boundaries of the lower Apalachicola–Chattahoochee–Flint River Basin, and physiographic divisions of the Coastal Plain Province in southeastern Alabama, northwestern Florida, and southwestern Georgia (modified from Torak and others, 1996).

- Describe the ground-water and surface-water flow systems of Lake Seminole and vicinity as a framework for identifying water-budget components;
- Develop a water budget for Lake Seminole that accounts for inflows to and outflows from the lake on a month-to-month basis;
- Evaluate techniques used for estimating evaporation; and
- Assess the importance of each water-budget component through sensitivity analyses.

Ground-water and surface-water flow systems described in this report are based on results of numerous hydrogeologic investigations that have been performed in the study area since 1965. From an initial investigation on the geology and water resources of the Lake Seminole area by Sever (1965) to an evaluation of stream-aquifer relations and the potentiometric surface of the Upper Floridan aquifer in the lower ACF River Basin by Mosner (2002), this report combines results of previous hydrogeologic investigations with recent and historical data to provide pertinent information for developing concepts of ground- and surface-water flow in the reservoir-stream-aquifer flow system. The ground-water conceptualization contains descriptions of the hydrogeologic setting of Lake Seminole and the Upper Floridan aquifer, a discussion of hydraulic characteristics of the Upper Floridan aquifer system, and a discussion of seasonal and long-term variations in ground-water levels.

The surface-water conceptualization contains descriptions of the drainage network, streamflow, dams, and navigational improvements for the main streams of the lower ACF River Basin. These descriptions are intended to provide necessary hydrologic information from which reservoir-stream-aquifer interaction can be evaluated for development of the Lake Seminole water budget.

The calculation of fluxes for water-budget components utilized limnologic and climatologic information obtained from instrumentation installed in and over Lake Seminole for this study and results of digital modeling of ground-water flow having surface-water interaction. Two overwater climate stations supplied values for climatic variables pertinent to the empirical formulas used to calculate lake evaporation. A network of lake-temperature monitoring and surface-water stations supplied data necessary to calculate lake evaporation by the energy-budget method. Previous digital modeling in the lower ACF River Basin provided estimates of leakage rates to and from the Upper Floridan aquifer that occur along the lake boundary.

Study Area

Lake Seminole is located in the lower ACF River Basin in parts of southwestern Georgia and northwestern Florida (fig. 1). The 6,800-square mile (mi^2) lower ACF River Basin includes Lake Seminole and the land area that exchanges ground- and surface-water flow with the lake. In Georgia, the study area includes Baker, Decatur, Early, Grady, Miller, Mitchell, and Seminole Counties; in Florida, the study area includes Gadsden and Jackson Counties (fig. 2).

Physiography

The lower ACF River Basin lies within the Coastal Plain physiographic province in southwestern Georgia, northwestern Florida, and southeastern Alabama, and is drained by the Apalachicola, Chattahoochee, and Flint Rivers and their tributaries (Torak and others, 1996) (fig. 1). The northern extent of the study area is located in the Fall Line Hills district near the updip limit of the Ocala Limestone (Clark and Zisa, 1976). The Fall Line Hills district is highly dissected with steep-hill slopes and streams that lie from about 50 to 250 ft below adjacent ridges. Relief diminishes gradually to the south and east where the Fall Line Hills district grades into the Dougherty Plain (Torak and others, 1996).

The Dougherty Plain is a nearly level lowland that pinches out where the Fall Line Hills district and Tifton Upland meet (Clark and Zisa, 1976). Formed by erosion, land-surface altitude ranges from about 300 ft at the northern extent of the plain to about 77 ft at Lake Seminole, and about 40 ft directly downstream of Jim Woodruff Lock and Dam. Land-surface slopes average about 5 feet per mile (ft/mi) (Hicks and others, 1987). Karst topography defines the landscape of the Dougherty Plain. A major feature near Lake Seminole and upstream along the Flint River is internal drainage, where streams connect the surface- and ground-water flow systems through karst conduits in the streambed. Sinkhole formation is responsible for the development of numerous ponds and wetlands that characterize the region; the bottoms of these ponds are filled with low-permeability sediment and hold water year-round. Underground channels created from dissolution of the Ocala Limestone capture surface water, and account for a substantial percentage of drainage in the Dougherty Plain (Hicks and others, 1987).

To the east, the basin is bordered by a well-defined northwest-facing feature called the *Solution Escarpment* (MacNeil, 1947), which forms a prominent boundary between the Tifton Upland and the Dougherty Plain. The crest of the Solution Escarpment creates a topographic and a surface-water divide between the Flint River to the west and the Ochlockonee and Withlacoochee Rivers to the east (Torak and others, 1996).

The southern limit of the study area terminates in northwestern Florida in the Tallahassee Hills, the equivalent of the Tifton Upland. Here, altitudes range from about 330 ft near the Georgia–Florida State line to about 100 ft south of the study area. Sediments in this area are composed of clayey sand, silt, and clay that end abruptly at the Apalachicola River (Torak and others, 1996).

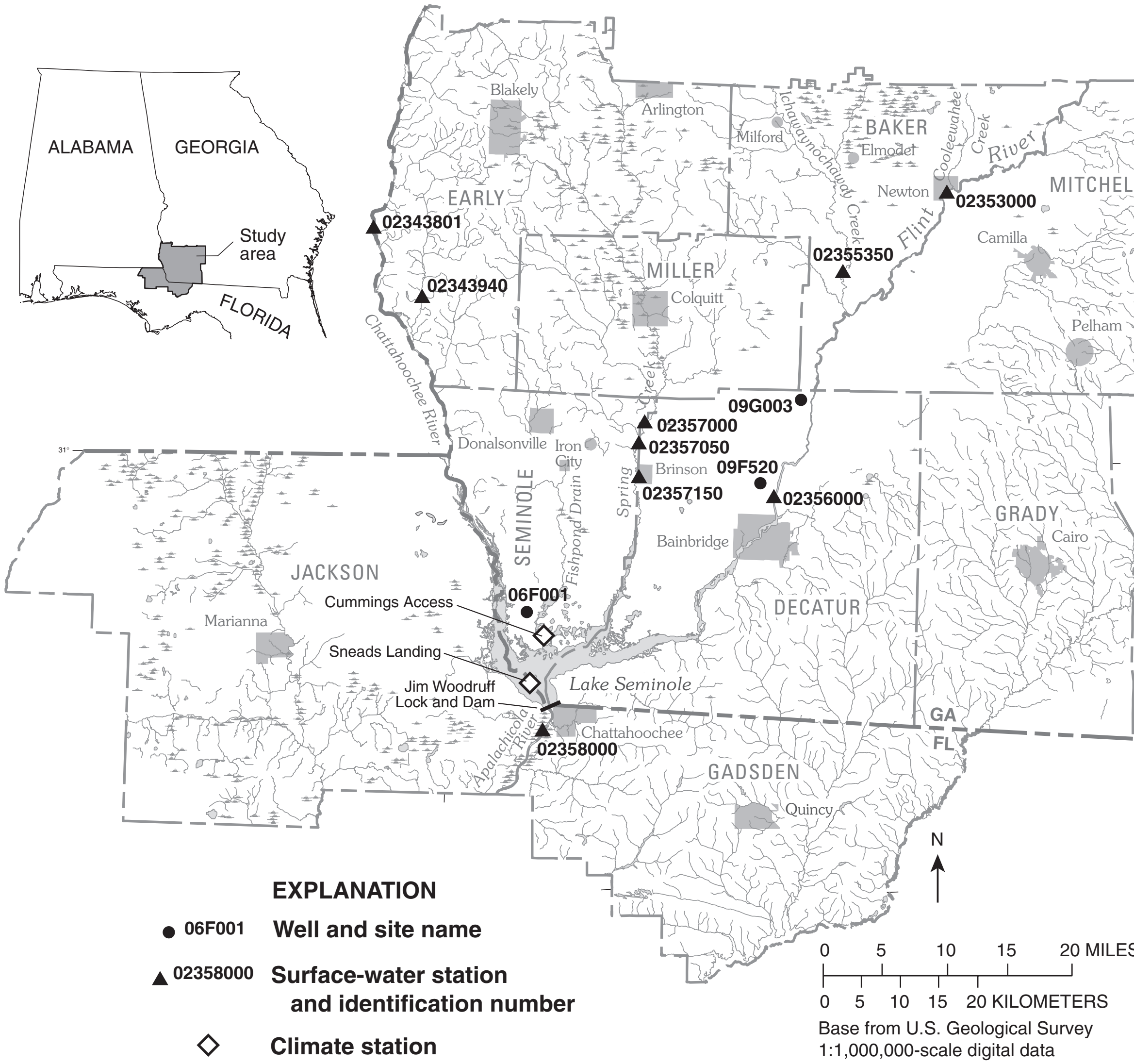


Figure 2. Study area, wells, surface-water, and climate stations near Lake Seminole.

Climate

The climate of the study area is humid subtropical. The mean annual temperature for the 42-year period 1959–2000 at Arlington in Early County, Georgia, is 66.7 degrees Fahrenheit (°F) (table 1; see fig. 2 for location). The coldest months, December and January, average about 51.3°F; occasional freezing temperatures occur during this time. The warmest months, July and August, have an average temperature of about 81.1°F; and temperatures near 100°F are not uncommon. Average annual rainfall is about 53.9 inches; highest monthly rainfall occurs during March; lowest rainfall occurs during October.

Previous Studies

Several hydrogeologic investigations have been conducted in the study area in Georgia. Sever (1965) described the hydrogeology in Seminole, Decatur, and Grady Counties, Georgia, and the water resources of the area surrounding Lake Seminole. Hayes and others (1983) described the hydrology of the Upper Floridan aquifer in the Dougherty Plain. Miller (1986) described the hydrogeologic framework of the Floridan aquifer system; Johnston and Bush (1988) summarized the hydrology of the Floridan aquifer system; Bush and Johnston (1988) evaluated the ground-water hydraulics, regional flow, and ground-water development of the Floridan aquifer system that included the study area. Torak and McDowell (1995) and Torak and others (1996) evaluated the hydrogeology and ground-water resources in the lower ACF River Basin. Albertson (2001), Torak (2001), and Mosner (2001), respectively, described the hydrogeology, water chemistry, and stream-aquifer relations in the lower ACF River Basin near Lake Seminole. Mosner (2002) evaluated stream-aquifer relations, ground-water seepage, and the potentiometric surface of the Upper Floridan aquifer in the lower ACF River Basin.

Selected investigations in the study area in Florida include a report by Vernon and others (1958), describing the geology of the area near Lake Seminole, and a report by Pratt and others (1996), describing the hydrogeology of the Northwest Florida Water Management District.

Several studies have been conducted that examine the instrumentation and methodologies associated with determining lake evaporation and developing lake water budgets. DeBruin (1978) developed a model for determining evaporation from a lake by combining two well-known models, the Priestley-Taylor and Penman equations. DeBruin and Keijman (1979) discussed the application of the Priestley-Taylor model for determining evaporation. Lee and others (1991) discussed the instrumentation required to estimate evaporation and developed a preliminary water budget of a seepage lake in central Florida. Rosenberry and others (1993) evaluated the instrumentation required to produce the energy budget for a lake in Minnesota. The energy-budget model is considered by most investigators to be one of the most accurate methods available for determining evaporation (Thomas C. Winter, U.S. Geological Survey, Denver, Colorado, oral commun., May 2000), because calculations of latent heat represent "true" values of latent-heat flux (Keijman, 1974; Stewart and Rouse, 1976). Lee and Swancar (1996) described the water budget and energy budget of a seepage lake in central Florida and compared methods for determining evaporation from the lake surface. Winter and others (1995) compared 11 analytical equations, including the 4 listed above, that are used to estimate evaporation with similar estimates computed from the energy-budget method. Their results showed that the Priestley-Taylor, Penman, and DeBruin-Keijman models agreed most closely with the energy-budget method when climate data were available near the lake.

Table 1. Climate data for Arlington, Early County, Georgia, 1959–2000 (*see* figure 2 for location).

[°F, degree Fahrenheit. Data from Georgia Automated Environmental Monitoring Network (2002)]

Month	Average maximum temperature (°F)	Average minimum temperature (°F)	Total precipitation (inches)
January	61.9	38.1	5.37
February	65.5	40.8	4.92
March	72.5	46.9	6.09
April	80.4	53.8	3.79
May	86.2	60.6	3.65
June	90.7	67.6	5.1
July	92.3	70.3	4.96
August	91.8	69.8	4.8
September	88.5	65.8	3.84
October	80.6	54.7	2.65
November	72.1	46	3.15
December	64.9	40.1	5.61
Average	**78.9**	**54.5**	**53.9**

Average annual temperature 66.7 °F

Well and Surface-Water-Station Numbering System

In this report, wells are identified with either a numbering system based on USGS topographic maps (Georgia) or a numbering system developed by the Northwest Florida Water Management District (Florida). In Georgia, each 7½–minute topographic quadrangle map has been given a number and letter designation beginning at the southwestern corner of the State. Numbers increase eastward through 39, and letters increase alphabetically northward through "Z," then become double-letter designations "AA" through "PP." The letters "I," "O," "II," and "OO" are not used. Wells inventoried in each quadrangle are numbered sequentially beginning with "1." Thus, the third well inventoried in the Chattahoochee quadrangle is designated 06D003. In Florida, wells are inventoried using a triple-A numbering identification system; each well is assigned a 4-digit identifier following "AAA," for example AAA1640. Springs are considered ground-water sites and, therefore, are identified in the same manner as wells.

Surface-water stations are identified by a numbering system used for all USGS reports and publications since October 1, 1950. The order of listing stations is in a downstream direction along the main channel. All stations on a tributary entering upstream from each mainstream are listed before that station. Each surface-water station is assigned a unique 8-digit number. The station number such as 02351890 includes the 2-digit number "02," which designates the site as a surface-water station, and the 6-digit downstream order number "351890."

Acknowledgments

The authors extend appreciation to the many agencies that contributed to the collection of data used in this report, including the Northwest Florida Water Management District and the U.S. Army Corps of Engineers. Gratitude is extended to all the landowners and land managers who allowed access to wells and springs to collect water-resource data.

Ground Water

The description of the ground-water flow system (fig. 3) is based on previous works by Sever (1965), Hayes and others (1983), Miller (1986), Hicks and others (1987), Johnston and Bush (1988), Torak and McDowell (1996), Torak and others (1996), and Stewart and others (1999). These studies indicate that recharge to the Upper Floridan aquifer is mainly by infiltrated precipitation, either directly through the limestone units of the aquifer or through the undifferentiated overburden. The undifferentiated overburden consists of chemically weathered limestone residuum and alluvium—composed of sand, silt, and clay—that partially confine the Upper Floridan aquifer. Although most layers of similar lithology in the undifferentiated overburden are discontinuous and can be traced for only short distances, a layer of clay is present that might be continuous throughout the lower half of the overburden (Torak and others, 1996). Low vertical hydraulic conductivity and the relative thickness of this clay layer within the overburden create a hydrologic barrier to vertical flow of water to and from the aquifer (Torak and others, 1996). The clay layer affects ground-water flow in the aquifer system by causing ground water to perch in overlying deposits following periods of heavy rainfall, thus decreasing the amount of ground-water recharge to the aquifer from infiltration and controlling the infiltration rate (Torak and others, 1996). Regional ground-water flow is from northwest to southeast; near streams, discharge is toward the stream channel.

Because the river channels cut into the overburden and are in contact with the Upper Floridan aquifer, there is potential for direct hydraulic connection between the aquifer and surface-water features. Sudden changes in river stage, however, do not necessarily correspond to a rise in ground-water level (Torak and others, 1996), indicating little hydraulic interaction between the stream and the Upper Floridan aquifer. When the ground-water level is higher than the stream stage, water in the Upper Floridan aquifer discharges to the stream, and the stream becomes a gaining stream. Conversely, when the ground-water level is lower than the stream stage, the stream may discharge water to the aquifer, and the stream becomes a losing stream. The rate of discharge depends on the hydraulic gradient between the aquifer water level and stream stage, and the permeability of the streambed (Hicks and others, 1987).

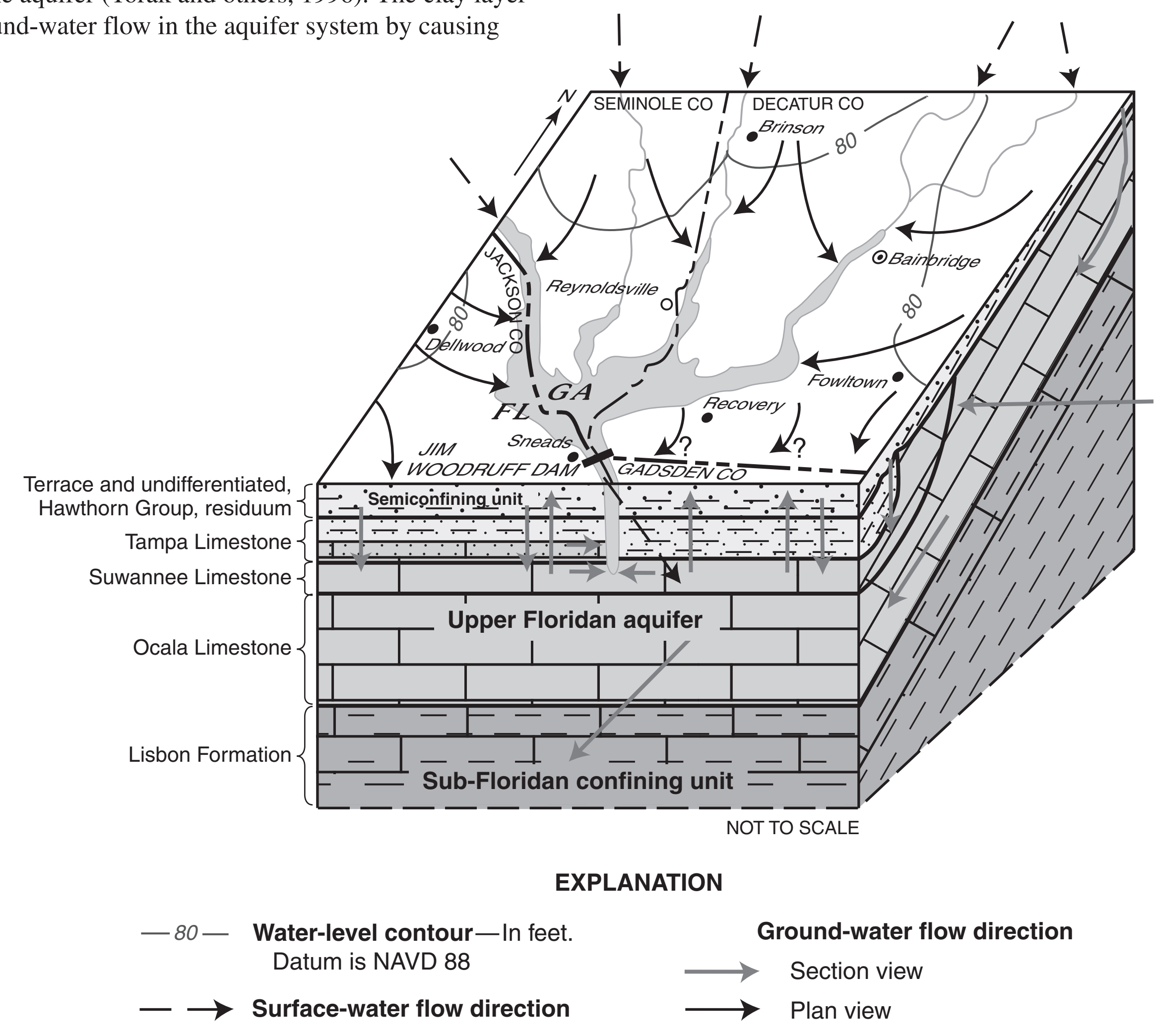


Figure 3. Schematic diagram of the ground-water flow system in the lower Apalachicola–Chattahoochee–Flint River Basin (modified from Torak and others, 1996).

Hydrogeologic Setting

The Coastal Plain Province is underlain by pre-Cretaceous to Quaternary sediments. In the study area, these sediments consist of alternating sand, clay, dolostone, and limestone, which dip gently and thicken to the southeast (Hicks and others, 1987). The Dougherty Plain is characterized by a highly transmissive flow system developed through dissolution and karst processes in the Ocala Limestone, which is the main water-bearing unit of the Upper Floridan aquifer. This flow system is defined by high rates of direct recharge through sinkholes, indirect recharge by vertical leakage through the overburden, and discharge to surface-water bodies such as the Flint and Chattahoochee Rivers, when ground-water levels exceed stream levels (Torak and others, 1996).

The Upper Floridan aquifer is a complex hydrogeologic system with a vertical and areal distribution that adds complexity to the dynamics of stream-aquifer interactions. Dissolution of the limestone by fluctuating ground-water levels and infiltrating rainfall have produced secondary permeability in the Ocala Limestone, making the aquifer highly productive. Solution conduits between the Solution Escarpment and the Flint River transmit large amounts of water from the Upper Floridan aquifer to springs that discharge along streams. The hydraulic gradient and the amount of dissolution and connectivity of conduits determine the relative speed with which water moves through the aquifer.

Throughout much of the study area, ground-water flow within the Upper Floridan aquifer is unconfined or semiconfined and discharges at discrete points by springflow or diffuse leakage into streams and other surface-water bodies (Johnston and Bush, 1988). The high degree of connectivity between the Upper Floridan aquifer and surface-water bodies is limited to the Ocala Limestone and younger units that are in contact with streams in the Lake Seminole area. Geologic units comprising the Upper Floridan aquifer systems are late-Eocene age and younger, forming, in ascending order, the Lisbon Formation, Clinchfield Sand, Ocala Limestone, Suwannee Limestone, Tampa Limestone, Hawthorn Formation, Miccosukee Formation, and surficial deposits of terraced and undifferentiated overburden (fig. 4). Hayes and others (1983), Miller (1986), Hicks and others (1987), Bush and Johnston (1988), Torak and McDowell (1995), and Torak and others (1996) provide details about these geologic units.

<table>
<tr><th rowspan="2">SERIES</th><th colspan="3">GEORGIA</th></tr>
<tr><th colspan="2">Geologic unit</th><th>Hydrologic unit</th></tr>
<tr><td>HOLOCENE AND PLEISTOCENE</td><td colspan="2">Terraced and undifferentiated deposits</td><td rowspan="3">Semiconfining unit</td></tr>
<tr><td rowspan="4">MIOCENE</td><td rowspan="4">Undifferentiated overburden (residuum)</td><td>Miccosukee Formation</td></tr>
<tr><td>Hawthorn Formation</td></tr>
<tr><td>Tampa Limestone</td><td rowspan="4">Upper Floridan aquifer</td></tr>
<tr><td>Suwannee Limestone</td></tr>
<tr><td rowspan="3">EOCENE</td><td colspan="2">Ocala Limestone</td></tr>
<tr><td></td><td>Clinchfield Sand</td></tr>
<tr><td colspan="2">Lisbon Formation</td><td>Lower confining unit</td></tr>
</table>

Figure 4. Correlation chart of geologic and hydrologic units in the lower Apalachicola–Chattahoochee–Flint River Basin (from Torak and others, 1996).

Hydraulic Characteristics

Variations in thickness and hydraulic conductivity determine the ability of geologic units to function as aquifers and to transmit usable amounts of ground water for consumption. These aquifer characteristics also provide a mechanism for vertical leakage between aquifers and surface water, or between subunits within aquifers that are separated by semiconfining units (Torak and others, 1996). Variations in lithology and patterns of secondary porosity result in a complex ground-water flow system characteristic of many karst areas.

Overlying Semiconfining Units

In the lower ACF River Basin in Alabama, Georgia, and the northern panhandle of Florida (Jackson County), the Upper Floridan aquifer is semiconfined above by undifferentiated overburden, consisting of alternating layers of sand, silt, and clay and chemically weathered residuum of the Ocala Limestone. In Gadsden County, Florida, however, the Upper Floridan aquifer is semiconfined by a clay bed at the base of the Tampa Limestone.

The relatively low vertical hydraulic conductivity and the substantial thickness of the laterally continuous clay layer present in some places in the overlying semiconfining units create a hydrologic barrier to vertical flow of ground water to and from the aquifer. This clay layer can have an effect on ground-water flow in the aquifer system, causing ground water to perch following periods of heavy rainfall, thus decreasing recharge of water to the aquifer by infiltration of precipitation (Torak and others, 1996). Where the clay in the lower-half thickness of the overburden is absent, however, flow of water from the surface through the overburden and into the aquifer is relatively unimpeded and rapid.

Upper Floridan Aquifer

The Upper Floridan aquifer generally ranges in thickness from a few feet at the updip limit to more than 700 ft in Florida; near Lake Seminole, the Upper Floridan aquifer ranges from about 150 to 400 ft thick (Torak and others, 1996). The aquifer is confined below by low-permeability sediments of the Lisbon Formation and, generally, is semiconfined above by undifferentiated overburden. The aquifer is exposed along sections of major streams—such as the Apalachicola, Chattahoochee, and Flint Rivers, and Spring Creek—where erosion has removed the overburden (Maslia and Hayes, 1988).

The capacity of the Upper Floridan aquifer to store and transmit large quantities of water is attributed to the fractured nature of the Ocala Limestone (Hayes and others, 1983) and associated dissolution of limestone by ground water circulating along bedding planes and fractures, and to interconnected conduits or solution openings (Hicks and others, 1987). Solution conduits transmit a major portion of the ground-water flow and contribute greatly to shaping the potentiometric surface of the aquifer (Hayes and others, 1983).

Computed values of transmissivity for the study area, from field tests of the Upper Floridan aquifer, generally range from about 2,000 to 1,300,000 feet squared per day (ft^2/d) (Hayes and others, 1983; Wagner and Allen, 1984). Although locally accurate, values of transmissivity derived from field tests might not be representative of regional transmissivity because of variability in hydraulic conductivity caused by fracture and solution openings (Torak and others, 1996); the large range of variation in hydraulic conductivity is the result of variations in size and distribution of solution openings. Regional values of transmissivity commonly range from 2,000 to 300,000 ft^2/d (Hayes and others, 1983). Transmissivity is lowest near the updip limit of the Ocala Limestone, where the aquifer is relatively thin. Transmissivity generally increases to the south, where the aquifer thickens, and adjacent to major streams, where flowing water has accelerated the development of solution openings (Maslia and Hayes, 1988).

Lisbon Formation

The lower confining unit in the lower ACF River Basin is the Lisbon Formation. The hard, well-cemented, and argillaceous nature of the limestone of the Lisbon Formation makes it a nearly impermeable base to the Upper Floridan aquifer (Hayes and others, 1983). Because of the relatively low hydraulic conductivity compared with the Upper Floridan aquifer, wells yield only a few gallons per minute from the Lisbon Formation; although southeast of the Dougherty Plain, domestic supplies of water can be obtained (Hayes and others, 1983).

Recharge by upward leakage to the Upper Floridan aquifer across the Lisbon Formation occurs in the northernmost part of the lower ACF River Basin at a flow rate of about 10 cubic feet per second (ft^3/s). Discharge from the Upper Floridan aquifer through the Lisbon Formation occurs in the southern part of the basin at a rate of about 5 ft^3/s, with no leakage in the Dougherty Plain. In contrast, the total lateral-flow component through the Upper Floridan aquifer in the study area is about 4,000 ft^3/s. Thus, in this area, the Lisbon Formation acts as a nearly impermeable lower boundary to the Upper Floridan aquifer (Faye and Mayer, 1996).

Ground-Water Levels

Ground-water levels in the Upper Floridan aquifer and overlying units in the study area fluctuate in response to seasonal and longer cycle variations in recharge from infiltrating precipitation, discharge by pumping and evapotranspiration, and interaction with surface-water features. The natural pattern of higher water-level altitude (or shallow depth to water with regard to land surface) in recharge areas and lower water-level altitude in discharge areas, such as near streams, can be affected by heavy ground-water pumping for agricultural and residential water use. Altitude of water levels in the Upper Floridan aquifer generally ranges from about 190 ft northwest of Lake Seminole to about 65 ft south of the lake. Near the lake, ground-water levels are influenced by lake stage, varying from about 80 ft near the northern impoundment arms of the lake to about 70 ft near the southern part of the Flint River impoundment arm and dam (Mosner, 2002). Magnitude and timing of water-level response to recharge and pumping stresses vary areally within each hydrologic unit and can be large or barely perceptible, and nearly instantaneous or very slow (Torak and others, 1996).

Seasonal Fluctuations

The water level in the semiconfining unit overlying the Upper Floridan aquifer in the lower ACF River Basin usually is highest from January or February through April, declines during summer and fall, and is at a minimum during November through December or January (fig. 5). Beginning in December and continuing through January, water levels in wells generally rise quickly in response to recharge by infiltrating precipitation (Torak and others, 1996). During late spring and summer, however, water-level response to precipitation is subdued because the precipitation either replaces the soil-moisture deficit in the unsaturated zone or is lost to evapotranspiration or runoff (Hayes and others, 1983). Summer precipitation generally is from convective storms, which produce rainfall that is more intense and of shorter duration than rainfall associated with frontal passages during other seasons; most summer precipitation is lost to runoff.

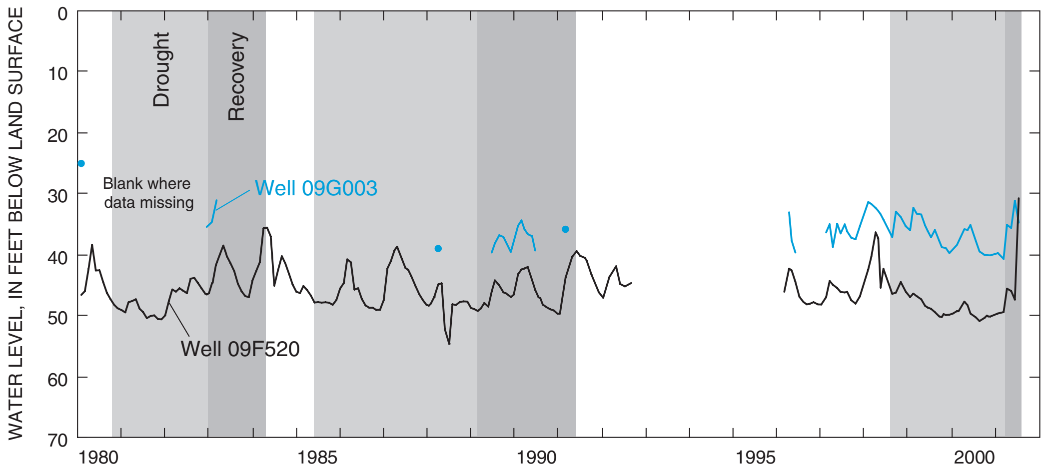


Figure 5. Seasonal water-level fluctuations in the semiconfining unit (well 09G003) and the Upper Floridan aquifer (well 09F520), 1980–2001 (*see* figure 2 for well locations).

Ground-water levels in the Upper Floridan aquifer also fluctuate seasonally in response to precipitation, evapotranspiration, and pumping (fig. 5). Late winter and early spring recharge by infiltrating precipitation, coupled with low evapotranspiration and pumping rates, cause the water level in the Upper Floridan aquifer to reach a maximum during February through April. During the growing season, combined effects of increased ground-water pumping for irrigation, higher evapotranspiration, and decreased recharge, when compared with winter and spring conditions, cause the ground-water level to reach a minimum by late summer through fall. Seasonal water-level fluctuations in the Upper Floridan aquifer range from about 2 ft in the eastern parts of the lower ACF River Basin in Georgia to about 30 ft near Albany, Georgia. Near Lake Seminole, however, ground-water-level fluctuations are subdued because of the strong aquifer-lake hydraulic connection. Near major centers of agricultural and industrial pumping, seasonal water-level fluctuations can exceed 30 ft and are amplified by drought conditions (Torak and others, 1996). Ground-water pumping does not result in formation of distinct cones of depression (Hicks and others, 1987); rather, because of the relatively uniform spacing of wells and pumpage throughout the lower ACF River Basin and the relatively high hydraulic conductivity, the potentiometric surface of the Upper Floridan aquifer is raised and lowered uniformly (Torak and others, 1996).

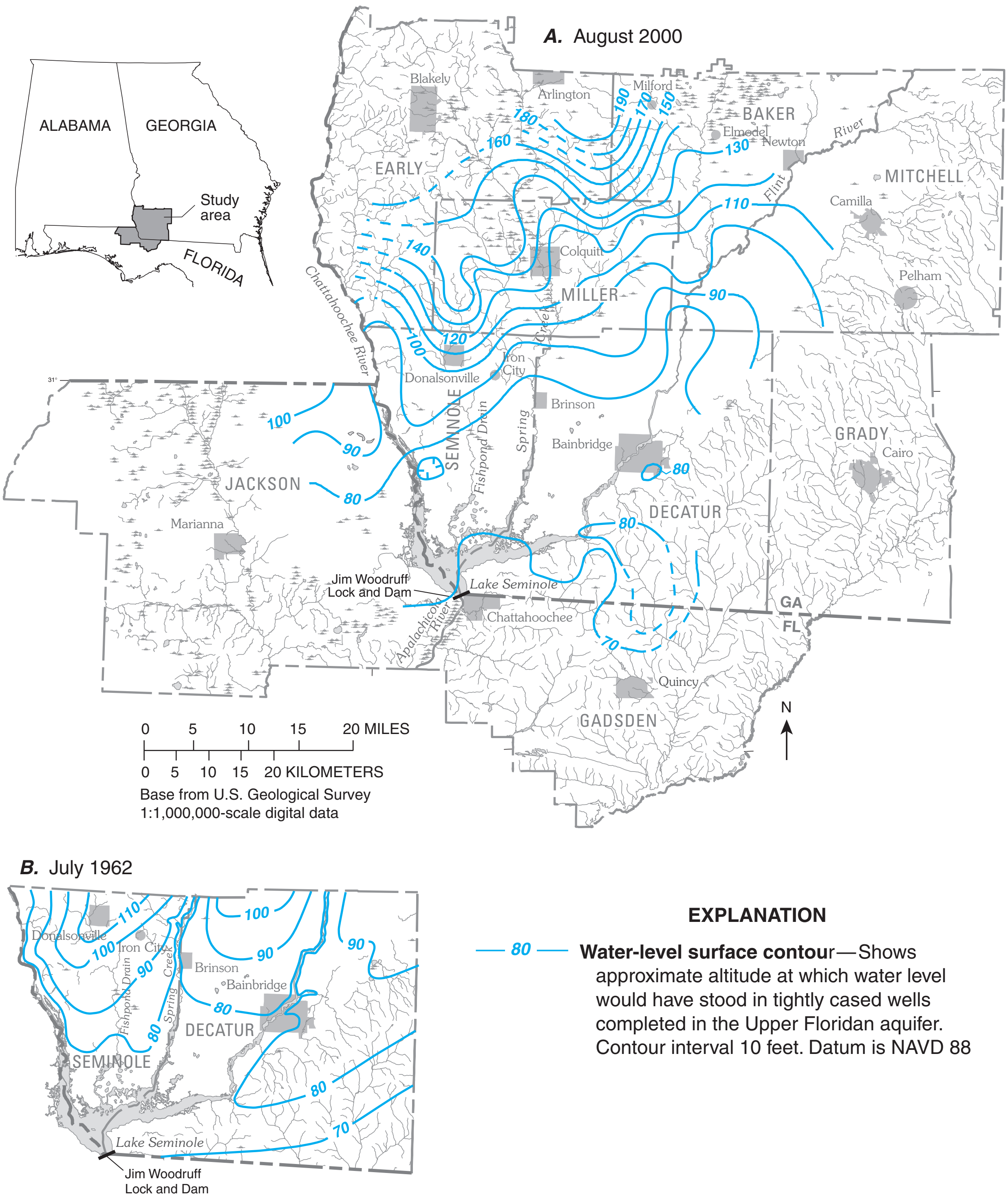
Effects of Drought and Pumping

Long-term declines in water level from drought conditions or pumping are not observed in the major water-bearing units in the lower ACF River Basin. During droughts of 1980–81, 1986–88, and 1998–2001, water levels in wells located in the Dougherty Plain declined to record or near-record lows, but recovered to predrought levels with the return of normal precipitation. Typical water-level response of the Upper Floridan aquifer in Georgia to recent drought conditions is shown in the water-level hydrograph of well 09F520 (fig. 5). During the drought of 1998–2001 record low water levels were recorded in the Lake Seminole area, where ground-water levels also fluctuate seasonally in response to climatic changes and ground-water pumping. Near the lake, however, these fluctuations are subdued as lake stage influences water levels in the Upper Floridan aquifer and undifferentiated overburden.

Predevelopment and recent potentiometric surfaces of the Upper Floridan aquifer (fig. 6) show that pumping rates, ranging from 54 to 119 million gallons per day (Mgal/d) from 1980 to 1999 in Seminole and Decatur Counties, failed to produce long-term water-level declines near Lake Seminole. Lake leakage into the Upper Floridan aquifer minimizes water-level declines caused by drought and pumping.

Surface-Water Influence

Surface-water features have a variable effect on ground-water levels in the area surrounding Lake Seminole. The impoundment of Lake Seminole inundated natural stream channels and other low-lying areas near streams and raised the water level of the Upper Floridan aquifer near the lake to about the same level as the lake. This process decreased the rate of ground-water discharge to streams and to Lake Seminole by reducing the hydraulic gradient between ground water and surface water. Because the pool elevation at Lake Seminole is maintained at an altitude of about 77 ft year-round, water levels in the adjacent aquifer and overlying semiconfining unit also are about 77 ft and remain nearly constant. Prior to impoundment, the hydraulic gradient from the aquifer to the streams was relatively steep, and flow was directed toward the Flint and Chattahoochee Rivers, Fishpond Drain, and Spring Creek. After impoundment, backwater conditions upstream of Jim Woodruff Lock and Dam reduced the hydraulic gradient from the aquifer toward the stream along the Chattahoochee and Flint Rivers for 47 mi, and to a lesser extent for the other two newly formed impoundment arms. The reduced gradient under impoundment conditions produced a corresponding reduced ground-water flow toward each impoundment arm (Jones and Torak, 2003).



EXPLANATION

80 — **Water-level surface contour**—Shows approximate altitude at which water level would have stood in tightly cased wells completed in the Upper Floridan aquifer. Contour interval 10 feet. Datum is NAVD 88

Figure 6. Generalized map of the water-level surface of the Upper Floridan aquifer, (*A*) August 2000 (Mosner, 2002) and (*B*) July 1962 (Sever, 1965).

Downstream of Lake Seminole, ground-water levels in the floodplain are influenced by the stage of the Apalachicola River and upgradient aquifer conditions. These water-level fluctuations are damped, however, by movement of water through the floodplain soils (Leitman and others, 1984). Observations indicate that ground-water-level fluctuations decrease with distance from the river; water levels in wells located more than 1.5 mi from the river are influenced mostly by local rainfall (U.S. Army Corps of Engineers, 1948).

Surface Water

Hydrologic factors that affect surface-water resources play an important role in the evaluation of reservoir-stream-aquifer relations because they influence surface-water interaction with ground water. The drainage network established by streams provides evidence of water-resource availability and stream-aquifer connection; changes in magnitude and duration of streamflow in the absence of direct channel precipitation and runoff indicate active communication between surface water and the underlying ground-water system; and, control structures show human attempts to harness the resource for many purposes (Torak and others, 1996).

Drainage

Lake Seminole is located at the confluence of the Chattahoochee and Flint Rivers, which provide headwater to the Apalachicola River that flows south to Apalachicola Bay and the Gulf of Mexico. The Chattahoochee River Basin drains an area of 8,770 mi^2, beginning in the Blue Ridge Mountains in northeastern Georgia and flowing southwest across the Piedmont and Coastal Plain physiographic provinces before entering Lake Seminole. Near Lake Seminole, the river is deeply incised within in its floodplain and cuts into the underlying limestone aquifer. There are no large tributaries to the Chattahoochee River within the study area, only small streams and creeks—such as Sawhatchee Creek—that mostly drain the undifferentiated overburden to the limestone (Torak and others, 1996).

The Flint River Basin drains an area of 8,460 mi^2, beginning in the Piedmont physiographic province south of Atlanta and flowing south across the Coastal Plain physiographic province before entering Lake Seminole. Major tributaries originate west of the river in the Coastal Plain and include Cooleewahee, Ichawaynochaway, and Spring Creeks. Spring Creek originates north of Colquitt, Georgia, and flows south into Lake Seminole, about 3 mi northeast of the confluence of the Flint and Chattahoochee Rivers. Only minor tributaries exist east of the Flint River draining from the Solution Escarpment, which creates a ground- and surface-water divide and forms the eastern basin boundary (Torak and others, 1996). The area between the Flint and Chattahoochee Rivers is drained internally, defined by streams that emerge and disappear underground. This area seems to be disconnected from the surface-water flow system (fig. 2); but because the Upper Floridan aquifer crops out in the river channels, the distinction between surface and ground water is not straightforward.

The Apalachicola River drains about 2,400 mi^2 of Coastal Plain sediments as it flows approximately 106 mi from Lake Seminole to Apalachicola Bay and the Gulf of Mexico (fig. 1) (Torak and others, 1996). Surface water from Lake Seminole flows through reverse springs into the ground-water system and eventually discharges into the Apalachicola River just downstream of the dam as springflow and diffuse channel leakage (Crilley, 2003).

Streamflow

The Chattahoochee and Flint Rivers provide most of the surface-water flow into Lake Seminole. Because the lake inundated much of the river channel, many tributaries to the Chattahoochee and Flint Rivers now enter the lake directly. Inflow to Lake Seminole was determined from daily streamflow measured at four USGS surface-water stations during the period of study (fig. 2 and table 2). The inflow to Lake Seminole from the Chattahoochee River was measured at the surface-water station near Columbia, Alabama (station number 02343801). The inflow to Lake Seminole from the Flint River was estimated by combining flows from the surface-water stations on the Flint River at Newton, Georgia (station number 02353000), and on the Ichawaynochaway Creek below Newton (station number 02355350). Ichawaynochaway Creek enters the Flint River below the gage at Newton but upstream of Lake Seminole. The inflow to Lake Seminole from Spring Creek, a tributary of the Flint River that now flows directly into the lake, is measured at the gage near Reynoldsville, Georgia (station number 02357150). These four inflows represent 90.5 percent of the entire drainage area of Lake Seminole (table 2).

All possible inflows to Lake Seminole from tributaries that are downstream of active surface-water stations were identified by examining the lake shoreline on 7½-minute USGS topographic maps. Eight tributaries were identified, which may contribute a small amount of ungaged inflow to Lake Seminole (table 3). Five of these tributaries are located along the Chattahoochee River impoundment arm: Cedar Creek (river mile 41.6) (river miles are defined as miles upstream from Jim Woodruff Lock and Dam); Sawhatchee Creek (river mile 35.5); Kirkland Creek (river mile 31.9); Bryans Creek (river mile 29.2); and Irwin Mill Creek (river mile 24). Two of the eight tributaries are located along the Flint River impoundment arm: Big Slough (river mile 33) and Fourmile Creek (river mile 24). One tributary, Fishpond Drain, enters Lake Seminole along its north shore between the Chattahoochee River and Spring Creek. These eight larger tributaries account for about 4 percent of the drainage area of Lake Seminole.

An additional 43 possible sources of surface-water inflows to Lake Seminole were identified by examining the lake shoreline on 7½-minute USGS topographic maps, but corresponding inflows were considered insignificant. Generally, these are

Table 2. Surface-water inflows to Lake Seminole.
[mi², square mile; DA, drainage area]

Surface-water station name	Station number	Drainage area (mi²)	Percent of Lake Seminole DA
Chattahoochee River near Columbia, Ala.	02343801	8,210	47.6
Flint River at Newton, Ga.	02353000	5,740	33.3
Ichawaynochaway Creek below Newton, Ga.	02355350	1,040	6
Spring Creek near Reynoldsville, Ga.	02357150	623	3.6
Total gaged inflows		**15,613**	**90.5**

intermittent streams, streams that drain small areas, and small wetlands along the shoreline. Most of these are along the Chattahoochee River impoundment arm and the south shore of the lake. Because of the proximity of drainage divides to the southeast and southwest of the lake, there is little contributing drainage area to the lake in these areas. There are no other inflows to Lake Seminole along the northeastern part of the lake because internal drainage and karst topography of the Dougherty Plain (fig. 1) provide ample solution conduits for surface water to drain to the underlying Upper Floridan aquifer.

Dams and Navigational Improvements

Inflow to Lake Seminole from the Chattahoochee and Flint Rivers is partially controlled by upstream dams and reservoirs, which are used for multiple purposes—including navigation, water supply, power generation, recreation, flood control, and fish and wildlife enhancement. Beginning in the late 1800s, two dams—City Mills and Eagle Phenix—were built on the Chattahoochee River at the Fall Line near Columbus, Georgia, and Phenix City, Alabama. Initially built for hydromechanical power in the textile mills, the dams have since been converted to hydroelectric projects. The Georgia Power Company also has six relatively small hydropower projects along the 30-mi reach of the Chattahoochee River in the Columbus, Georgia, area. These projects, however, do not contain appreciable storage capacity; therefore, they do not have a notable effect on the flow of the Chattahoochee River (U.S. Army Corps of Engineers, 1988).

The U.S. Army Corps of Engineers constructed and operates five large impoundments in the basin. They are, in order from the Chattahoochee River headwaters: Lake Sidney Lanier (Buford Dam), near Gainesville, Georgia; West Point Lake (West Buford Dam), near West Point, Georgia; Walter F. George Lake (Walter F. George Lock and Dam), near Eufaula, Alabama; Lake George W. Andrews (George W. Andrews Lock and Dam), near Columbia, Alabama; and Lake Seminole (Jim Woodruff Lock and Dam), at the Georgia–Florida State line and headwater of the Apalachicola River. The major stream regulation on the Chattahoochee River is provided by the 38,000-acre Lake Sidney Lanier, located about 50 mi northeast of Atlanta, Georgia. Minor, but important, amounts of storage and flow regulation are provided by West Point Lake and Walter F. George Lake. Lakes Andrews and Seminole are run-of-the-river projects, which largely depend on inflow controlled by upstream impoundments. Lake Seminole's limited storage is capable of short-duration-flow augmentation of the Apalachicola River for navigation purposes (U.S. Army Corps on Engineers, 1988).

Another impoundment, the Morgan Falls Dam, on the Chattahoochee River near Atlanta, Georgia, regulates releases from Buford Dam and assists in maintaining adequate flow for waste assimilation downstream. Georgia Power Company operates this impoundment.

Flint River flows are relatively unimpeded with only two reservoirs along its length before entering Lake Seminole: Lake Blackshear (Warwick Dam), operated by the Crisp County Power Commission; and Lake Chehaw (Flint River Dam), operated by the Georgia Power Company. These projects are primarily run-of-the-river and do not affect downstream flows appreciably (U.S. Army Corps of Engineers, 1988).

Table 3. Ungaged inflows to Lake Seminole.
[mi², square mile; DA, drainage area; do., ditto]

Tributary name	River basin	Drainage area (mi²)	Percent of Lake Seminole DA
Cedar Creek	Chattahoochee	30.9	0.18
Sawhatchee Creek	do.	75.6	.44
Kirkland Creek	do.	39.6	.23
Bryans Creek	do.	38.3	.22
Irwin Mill Creek[a]	do.	21.6	.13
Big Slough	Flint River	334.2	1.94
Fourmile Creek	do.	24.4	.14
Fishpond Drain	do.	127.3	.74
Total substantial ungaged inflows		**691.9**	**4.02**

[a]Drainage area of Irwin Mill Creek is approximate.

Jim Woodruff Lock and Dam controls outflow from Lake Seminole. Lake stage varies around a normal pool altitude of 77 ft. At normal pool, the lake has an area of 37,500 acres, impounding the lower 47 mi of the Flint River (to approximately 10 mi downstream of Newton, Georgia) and the lower 47 mi of the Chattahoochee River (to George W. Andrews Lock and Dam near Columbia, Alabama).

Water-Budget Calculations for Lake Seminole

Water-budget calculations provide a means to quantify contributions of hydrologic components to the total inflow and outflow of water for Lake Seminole, to evaluate the relative importance of each hydrologic component to the water budget, and to estimate errors associated with the measurement and calculation of each component. Error estimates for each hydrologic component indicate the accuracy and reliability by which a component is known or can be measured, and can be used to guide future data collection and measurement techniques.

Lake Seminole receives water from direct precipitation, surface-water inflow, and ground-water inflow, and loses water by evaporation, surface-water outflow, and leakage to the subsurface (fig. 7). These hydrologic components operate continuously to provide a change in lake storage volume, ΔS, which is expressed by the water-budget equation:

$$\Delta S = P + SW_{in} - SW_{out} + GW_{in} - GW_{out} - E, \quad (1)$$

where

ΔS = change in lake storage;

P = precipitation;

SW_{in} = surface-water inflow;

SW_{out} = surface-water outflow;

GW_{in} = ground-water inflow;

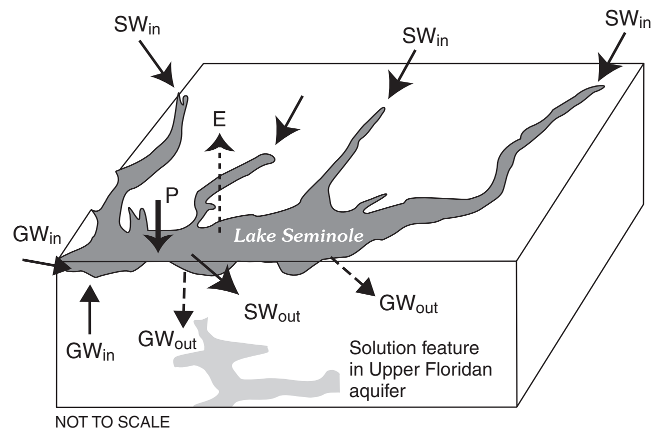
GW_{out} = lake leakage to the undifferentiated over-burden and Upper Floridan aquifer; and

E = evaporation.

Volume for each term is expressed in units of length by dividing by average lake surface area.

Volume calculations for each hydrologic component were computed daily and averaged monthly, and water-budget terms were expressed as monthly rates, in inches. Implicit to the volume calculation for each hydrologic component in equation (1) is an error term, *e*, which is evaluated separately from component calculations; error is discussed later in this report.

When calculating the water budget, each hydrologic component was measured or estimated independently; that is, no component was computed as the residual of other component contributions to the water budget. Inputs to the lake include direct precipitation, inflowing streams, and ground water. Outputs from the lake include lake evaporation, outflow through Jim Woodruff Lock and Dam to the Apalachicola River, and lake leakage. Lake storage was estimated from lake stage and the relation between lake stage and volume. Monthly estimates for these hydrologic components were computed for the study period from April 2000 to September 2001, yielding monthly estimates of change in lake storage. All components were used to calculate error in the lake water budget.



$$P + SW_{in} - SW_{out} + GW_{in} - GW_{out} - E = \Delta S$$

P = precipitation
SW_{in} = surface-water inflows
SW_{out} = surface-water outflows
GW_{in} = ground-water inflows
GW_{out} = ground-water outflows
E = evaporation
ΔS = change in lake storage

Figure 7. The hydrologic cycle for Lake Seminole.

Precipitation

Precipitation was measured at two overwater climate stations (fig. 2) using tipping-bucket rain gages. Daily precipitation values from the two stations were then averaged to account for variability in precipitation over the lake surface. Weather patterns, especially during summer months, include a large number of convective thunderstorms that typically produce local variations in precipitation; several occurrences of significant precipitation variation between stations were noted during the study period. Average monthly precipitation during water year 2001 (October 2000–September 2001) was 0.45 inch higher at the climate station at Cummings Access, Georgia (4.16 inches), than at the climate station at Sneads Landing, Florida (3.71 inches). The variation in the differences in monthly precipitation was large between the two stations during the study period with a mean difference of 0.45 inch and a standard deviation of 0.83 inch. Seasonal variation of precipitation in the study area was similar to patterns in the long-term record, ranging from a low of 0.20 inch during May 2000 to a high of 10.61 inches during March 2001 (table 4).

Table 4. Comparison of long-term average (based on 42 years of record, 1959–2000, Georgia Automated Environmental Monitoring Network, 2002) precipitation to monthly precipitation during the study period.
[All values in inches]

Month and year	Monthly long-term average	Total monthly precipitation	Difference
Apr. 2000	3.79	3.20	0.59
May 2000	3.65	.20	3.46
June 2000	5.10	2.70	2.40
July 2000	4.96	3.16	1.80
Aug. 2000	4.80	1.93	2.87
Sept. 2000	3.84	6.62	2.78
Oct. 2000	2.65	.45	2.20
Nov. 2000	3.15	3.77	.62
Dec. 2000	5.61	2.47	3.14
Jan. 2001	5.37	1.84	3.53
Feb. 2001	4.92	.73	4.19
Mar. 2001	6.09	10.61	4.52
Apr. 2001	3.79	1.32	2.47
May 2001	3.65	1.04	2.61
June 2001	5.10	8.44	3.34
July 2001	4.96	8.67	3.71
Aug. 2001	4.80	4.86	.06
Sept. 2001	3.84	3.06	.78
Oct. 2001	2.65	2.41	.24
Total	**82.72**	**67.48**	**15.24**

Surface Water

Surface-water inflow to Lake Seminole was measured at four surface-water stations (fig. 2; table 2) on the Chattahoochee and Flint Rivers, Ichawaynochaway Creek, and Spring Creek. These inflows have a combined drainage area of 15,613 mi^2, 90.5 percent of the total drainage area of Lake Seminole (17,230 mi^2).

Daily inflow to the Chattahoochee River impoundment arm was measured near Columbia, Alabama (station number 02343801). This surface-water station is located at the George W. Andrews Lock and Dam at river mile 47 (river miles are defined as miles upstream from Jim Woodruff Lock and Dam), which is the most upstream point of Lake Seminole, on the Chattahoochee River impoundment arm, when the lake is at a normal pool altitude of 77 ft.

Daily inflow to Lake Seminole from the Flint River is the sum of flows of the Flint River at Newton, Georgia, and Ichawaynochaway Creek, which enters the Flint River below Newton. The Newton surface-water station (station number 02353000) on the Flint River is located about 72 mi upstream from Jim Woodruff Lock and Dam. The surface-water station on Ichawaynochaway Creek below Newton, Georgia (station number 02355350), is located just upstream of its confluence with the Flint River at approximately river mile 55.

Daily inflow was measured on the Spring Creek impoundment arm near Reynoldsville, Georgia (station number 02357150), using acoustic velocity metering (AVM). AVM was used at this surface-water station to measure velocities in the stream section due to the existence of backwater conditions. Traditional streamgaging methods estimate discharge using stream stage, which is inaccurate under backwater conditions; the AVM measures the velocity of water in the stream channel independent of stage, and, with the stream-channel cross section, can be used to develop a volumetric flow rate.

The U.S. Army Corps of Engineers measured daily outflow from Lake Seminole at Jim Woodruff Lock and Dam and on the Apalachicola River at Chattahoochee, Florida (station number 02358000), about 0.6 mi downstream of the dam. Measured outflow generally is higher on the Apalachicola River than at the dam (fig. 8), which can be attributed to ground-water inflow to the river channel along the reach between the dam and the surface-water station. Outflow generally was lower at the Apalachicola River gage from April through July 2000 than from the dam (fig. 8). Large (greater than 5,000 ft^3/s) differences in daily outflow existed during March and April 2001. Measurement error, if any, cannot be discerned from discharge measured at either the dam or the surface-water station. Therefore, discharge measurements at the Apalachicola River gage were used in water-budget calculations, but were modified to account for the gain in streamflow from ground water and used as outflow from Lake Seminole. Measured outflow was decreased by 550 ft^3/s to account for the streamflow gain due to ground-water discharge (*see* "Lake Leakage" section).

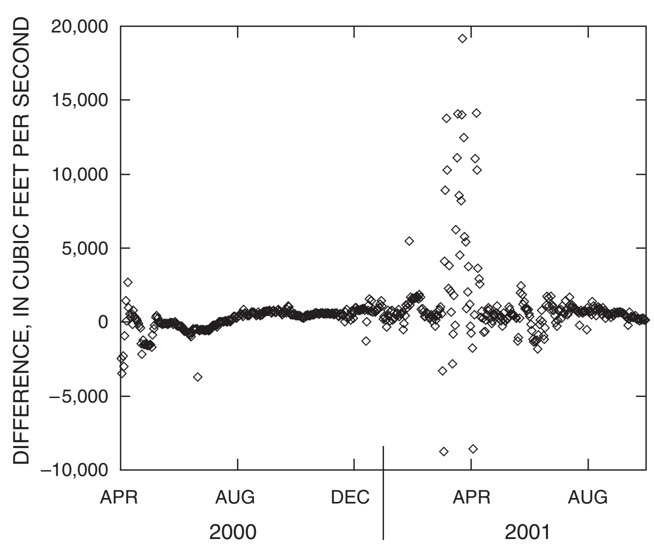


Figure 8. Difference in discharge between the Apalachicola River at Chattahoochee, Florida, and Jim Woodruff Lock and Dam (*see* figure 2 for locations).

Ungaged Inflow to Lake Seminole

A small amount of inflow to Lake Seminole is not accounted for by the four gaged inflows. The drainage area represented by the four gaged inflows represents about 91 percent of the total drainage area to Lake Seminole (table 2), leaving about 9 percent ungaged. Fifty-one minor inflows to Lake Seminole were identified on 7½-minute USGS topographic maps, downstream of the four gaged inflows. Of these, the eight largest ungaged inflows represent about 4 percent of the total drainage area of Lake Seminole (table 3). The remaining 5 percent of unaccounted drainage area is made up of small areas along the lake that include the 43 minor inflows and karst areas north of the lake, which have little surface-water drainage.

Six of the eight largest inflows were measured during drought conditions, the results of which are summarized in table 5. Discharge measurements for three of the inflows on October 23–28, 1986, were made as part of a ground-water modeling study (Torak and others, 1996); discharge measurements were made for six inflows during three synoptic runs performed between October 1999 and August 2000, as part of a ground-water seepage study (Mosner, 2002). The USGS recently installed a gage on Sawhatchee Creek at Cedar Springs (station number 02343940). Daily flows measured from January 19 to September 30, 2002, range from a minimum of 5.8 ft^3/s to a maximum of 206 ft^3/s, with an average flow for this period of 34 ft^3/s.

The inflow to Lake Seminole from the eight largest ungaged inflows was approximated based on limited streamflow information available along with drainage-area size and knowledge of flow characteristics of the tributaries. It is assumed that the 43 minor inflows do not contribute substantial inflow to the lake. The tributary inflows were analyzed in groups, depending on the basin into which they flowed. Five of the eight largest ungaged inflows are located in the Chattahoochee River Basin; of these, Sawhatchee Creek has the largest drainage area (75.6 mi^2, table 3). Average flows measured on Sawhatchee Creek from January 19 to September 30, 2002, were 0.47 percent of the average flows of the Chattahoochee River near Columbia, Alabama, for the same period. The sum of the flows at Cedar Creek, Kirkland Creek, and Bryans Creek (29.3 ft^3/s) measured on April 25, 2000, was similar to the flow of Sawhatchee Creek that same day (36.2 ft^3/s). No flow measurements were available for Irwin Mill Creek, but this basin has the smallest drainage area of the five inflows (21.6 mi^2) and, hence, likely contributes only a small amount of flow. Based on this limited information, it is reasonable to assume that these ungaged inflows constitute approximately 1 percent of the flow in the Chattahoochee River at Columbia, Alabama. Therefore, in order to account for these five inflows, daily inflow was approximated as 1 percent of the daily Chattahoochee River flow.

Table 5. Summary of streamflow measurements for substantial ungaged inflows to Lake Seminole.

[mi^2, square mile; DA, drainage area; ft^3/s; cubic foot per second; %, percent. Kirkland Creek drainage area is approximated as 90 percent of entire Kirkland Creek drainage area]

Streamflow station name	Station number	Drainage area (mi^2)	Percentage of basin DA accounted for in analysis	Percentage of Lake Seminole DA	Streamflow (ft^3/s)	Date(s)
Cedar Creek at State High-way 95 near Calumet, Ala.	02343850	30.3	98.1	0.18	19.4	Apr. 25, 2000
					6.14	July 27, 2000
Sawhatchee Creek at Cedar Springs, Ga.	02343940	64.2	84.9	.37	9.7	Oct. 23–28, 1996
					10.6	Oct. 18, 1999
					36.2	Apr. 25, 2000
					5.21	July 27, 2000
Kirkland Creek at Williams Road, near Jakin, Ga.	02344009	35.6	90	.21	9.86	Apr. 25, 2000
Bryans Creek at State High-way 95 near Crosby, Ala.	02344025	34.7	90.5	.2	0	Apr. 25, 2000
Big Slough at Georgia Highway 97 near Bainbridge, Ga.	02355950	315	94.3	1.83	0	Oct. 23–28, 1996
					0	Oct. 18, 1999
					0	Apr. 24, 2000
					0	July 31, 2000
Fishpond Drain at State Route 285 near Donalsonville, Ga.	02357310	38.8	30.5	.23	0	Oct. 23–28, 1996
					0	Oct. 19, 1999
					0	Apr. 24, 2000

The Flint River Basin contains three of the eight largest ungaged inflows (table 3). Fourmile Creek has the second smallest drainage area of the eight largest ungaged inflows (24.4 mi^2) and probably provides only a minimal amount of flow, although no flow measurements were available for verification. Big Slough and Fishpond Drain have the two largest drainage areas of the eight largest ungaged inflows, 334.2 mi^2 and 127.3 mi^2, respectively (table 3). All low-flow synoptic measurements indicated that there was no flow on these tributaries (table 5). Big Slough normally flows during wet conditions, but was dry at the time flow measurements were made; however, the flow occasionally reverses due to internal drainage of the karst topography and backwater conditions of Lake Seminole. Flows in Big Slough tend to be smaller than expected for the drainage-area size because the karst topography in the basin reduces surface-water drainage.

The synoptic-flow measurements indicating no flow for Fishpond Drain were taken far upstream of where it flows into the lake, thus representing only about 30 percent of the drainage area of Fishpond Drain (table 5). The karst flow system in the lower part of the Fishpond Drain Basin minimizes the potential for surface-water drainage. There is a low potential for ground-water inflow to Fishpond Drain due to small hydraulic gradients between surface water and the aquifer in this part of the basin. In contrast, high ground-water inflow occurs along Spring Creek, where hydraulic gradients are higher than those in adjacent Fishpond Drain, as indicated by numerous springs that are present along Spring Creek. Despite the larger drainage areas of the ungaged inflows in the Flint River Basin, ungaged inflows in the Flint River Basin are lower than ungaged inflows in the Chattahoochee River Basin. Therefore, as an approximation, daily inflow from the three ungaged inflows in the Flint River Basin were calculated as 0.5 percent of the average flows measured on the Flint River near Newton, Georgia.

Error Analysis

The accuracy of streamflow records depends on the stability of the stage-discharge relation, accuracy of stage measurements, accuracy and frequency of discharge measurements, and interpretation of records (Novak, 1985). A determination of accuracy of USGS daily average streamflow records is reported on a water year (October–September) basis for each surface-water station. Daily streamflow records are rated as excellent, good, fair, or poor. An excellent rating indicates that 95 percent of the daily discharges are within 5 percent of the true discharge. A good rating indicates that 95 percent of the daily discharges are within 10 percent of the true discharge. A fair ration indicates that 95 percent of the daily discharges are within 15 percent of the true discharge. A poor rating indicates that the daily streamflow records do not meet the fair rating (McCallum and Hickey, 2000; Franklin and others, 2000; McCallum and others, 2001).

Discharge measurements for the Chattahoochee River (station number 02343801) were reported as good in water year 2000 and fair in water year 2001. During this study (April 2000–September 2001), 66 days (12 percent) of daily streamflow record were estimated. Estimated values are considered poor (McCallum and Hickey, 2000).

Streamflow records for the Flint River (station number 02353000) and Ichawaynochaway Creek (station number 02355350) were reported as good in both water years 2000 and 2001. The Flint River data had only 1 day of estimated flow; Ichawaynochaway Creek had 11 days of estimated flows (2 percent) during the period of study. Flow measurements for Spring Creek (station number 02357150) were reported as good in both water years 2000 and 2001, except when flows were greater than 2,000 ft^3/s, which were considered fair. Flows were greater than 2,000 ft^3/s on 9 days during the study period (1.6 percent) (McCallum and Hickey, 2000, McCallum and others, 2001).

Measured streamflow for the Apalachicola River (station number 02358000) was reported as good in both water years 2000 and 2001. There were only 2 days with estimated flows. Accuracy of flow measurements was not reported for gages operated by the U.S. Army Corps of Engineers at Jim Woodruff Lock and Dam.

Although estimated ungaged inflow in the study area is small (less than 1 percent of total lake inflow), the error in estimating ungaged inflow can be quite large and add to the errors in total surface-water inflow. There are few measurements of streamflow for the eight largest ungaged inflows; these were made on specific dates during low-flow conditions, which are not indicative of all flow conditions. The method of approximating ungaged inflows as a percentage of the gaged inflows assumes that ungaged drainage areas exhibit similar hydrologic response as gaged drainage areas. This is unlikely because the Chattahoochee River flows are regulated, so flow variations on the Chattahoochee River will not be representative of flow variations on unregulated streams in smaller basins.

Lake Storage

A relation between lake stage and storage enabled daily estimates of lake storage to be obtained as a continuous function of lake stage. A regression equation developed from stage and storage data predicted lake storage to within 0.1 percent of the published stage-volume relation for the range of lake stage measured during the study period (fig. 9). The U.S. Army Corps of Engineers takes daily measurements of lake stage at midnight. These measurements are available at *http://water.sam.usace.army.mil/acfframe.htm*. Stage-area relations also were used to calculate evaporation by the energy-budget method and empirical equations.

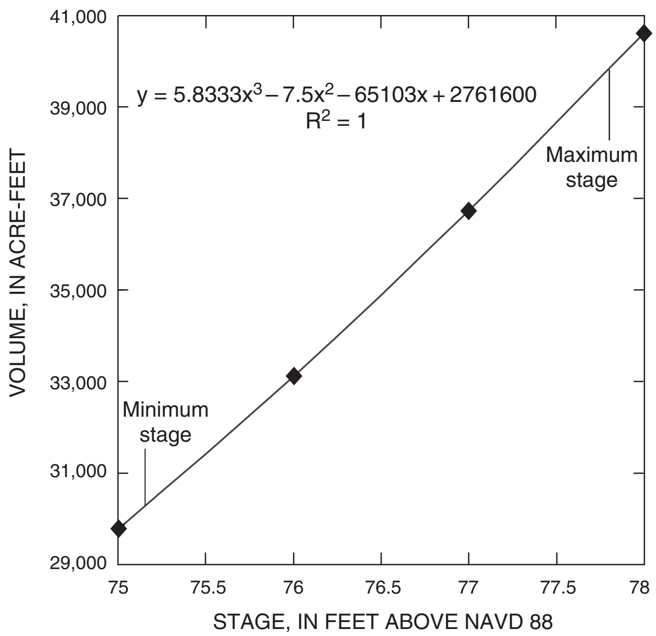


Figure 9. Regression defining the stage-volume relation for Lake Seminole.

Ground-Water Inflow and Lake Leakage

Ground-water inflow to Lake Seminole consists of flow across the lake-aquifer boundary from the aquifer into the lake and stream channels located below the surface-water stations. Lake leakage consists of ground water that flows across the lake-aquifer boundary from the lake into the aquifer. Rates of ground-water inflows and lake leakage were obtained from results of a previous digital simulation of October 1986 conditions (Torak and others, 1996). The model developed for that simulation was based on an application of the USGS modular finite-element model, MODFE (Cooley, 1992; and Torak, 1993a,b). The simulation results were used to approximate ground-water conditions in the Upper Floridan aquifer around and beneath the lake during the current study period. Simulated rates of ground-water inflow along stream channels and impoundment arms were adjusted monthly based on calculated stream baseflow to account for seasonal variations in ground-water flow rates. Lake leakage was determined by streamflow gain of the Apalachicola River just downstream of Jim Woodruff Lock and Dam. Streamflow gain is attributed to lake leakage that enters the aquifer and then discharges into the river.

Estimation of Flow Rates

Rates of ground-water inflow to Lake Seminole and lake leakage (fig. 10) were estimated from simulated discharge across finite elements, representing the main body of the lake and across linear and nonlinear head-dependent (Cauchy-type) ground-water flow boundaries. The Cauchy-type boundaries represent streams and impoundment arms of the lake; the lake was represented by finite elements (Torak, 1993a). Cauchy-type boundaries, which are coincident with element sides, represented the channels of the main gaged inflows to Lake Seminole—that is, the Chattahoochee and Flint Rivers, and Spring and Sawhatchee Creeks. On the Chattahoochee River, the surface-water station was located outside of the model area, so Cauchy-type boundaries represented the stream channel from the model boundary into the upper reaches of the corresponding impoundment arm of the lake and Sawhatchee Creek. The Flint River and Spring Creek were represented with Cauchy-type boundaries that extend downstream of the gages and into the corresponding impoundment arms, but not along the pre-impoundment channels in the main body of the lake. Simulated ground-water flow rates represent hydrologic conditions that existed in October 1986, which is the steady-state calibration period for the model. There was a drought in October 1986 similar in intensity to the drought that occurred during the study period; therefore, simulated ground-water flow around Lake Seminole and lake leakage for October 1986 conditions were assumed to closely approximate the flow conditions that existed during the study period.

The model simulated small-magnitude, areally distributed flow (ground-water inflow and lake leakage) across finite elements that represent the main body of the lake. Simulated ground-water inflow to the main body of Lake Seminole was about 34 ft^3/s, and lake leakage was about 14.7 ft^3/s. This ground-water inflow was included in the total ground-water inflow to the lake but was not varied seasonally like the other simulated ground-water inflows (see below). Because the areally distributed inflow rates are relatively small, seasonal variation would have little effect on the overall lake water budget. Lake leakage was approximated using another approach (see below) because the rate of lake leakage appears to be underestimated according to ground-water model results.

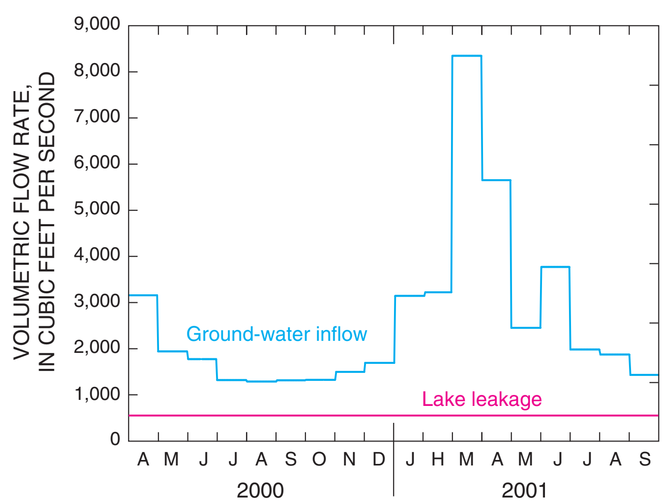


Figure 10. Lake Seminole ground-water inflow and lake leakage derived from simulated results for the period April 2000–September 2001.

All Cauchy-type boundaries representing streams simulated ground-water inflow to stream channels and impoundment arms from the Upper Floridan aquifer (table 6). Unit discharge across the Cauchy-type boundary, or ground-water inflow per unit of channel length, generally is larger for upstream reaches along the Chattahoochee River and Flint River impoundment arms than for the downstream reaches, even though boundary coefficients generally are higher for downstream reaches than upstream reaches. Boundary coefficients are a function of hydraulic conductivity, streambed width, and streambed thickness, and describe how easily water flows across the boundaries for a given hydraulic gradient. Thus, for inflow rates to be higher in the upstream reaches than in the downstream reaches, hydraulic gradients toward the stream in the upstream reaches need to be much higher than those in the downstream reaches to overcome the low boundary coefficient values in the upstream reaches. This indicates that ground-water inflow along these impoundment arms is more sensitive to hydraulic gradients in the aquifer close to the streams than to boundary coefficients.

Seasonal variations in ground-water inflow across Cauchy-type boundaries along each impoundment arm were simulated on a monthly basis during the study period. The inflows were varied according to changes in monthly baseflow estimates of streamflows measured at gages along the impoundment arms using HYSEP, a hydrograph separation program (Sloto and Crouse, 1996). A scaling factor was developed for each impoundment arm, defined as the ground-water inflow downstream of the gage derived from the October 1986 ground-water simulation (Torak and others, 1996) divided by the estimated October 1986 baseflow from HYSEP. (Since streamflow data were unavailable in October 1986 for Spring Creek near Reynoldsville, Georgia, station number 02357150, the average of the estimated baseflows for October 2000 and 2001 was substituted for the October 1986 value, because 2000 and 2001 both were drought years similar to 1986.) Monthly ground-water inflow was calculated as the product of the monthly estimated average baseflow from HYSEP and the scaling factor, to proportionally adjust ground-water inflow to the amount of upstream baseflow. Scaling factors for the Chattahoochee and Flint Rivers and Spring Creek were 0.448, 0.426, and 0.173, respectively. Thus, baseflow contributes a larger percentage of total streamflow during low-flow conditions than during high-flow conditions, a result that is consistent with that found by Mosner (2002).

Table 6. Summary of Cauchy-type boundaries used to estimate flow between the Upper Floridan aquifer and stream channels.

[ft/d, foot per day; ft^3/s, cubic foot per second, mi, mile; $ft^3/s/mi$, cubic foot per second per mile; do., ditto]

Zone[a]	Boundary type	Stream name	Boundary coefficient (ft/d)	Ground-water inflow (ft^3/s)	Length of channel (mi)	Ground-water inflow per channel length ($ft^3/s/mi$)
37	Linear	Chattahoochee River	2.0	10.5	3.1	3.3
38	do.	do.	60	185.6	10	18.5
39	do.	do.	100	142.7	9	15.8
40	do.	do.	150	45.7	5	9.2
12	do.	Flint River	120	15.8	.8	20.2
13	do.	do.	100	156.7	16.8	9.3
14	do.	do.	100	81.5	4.3	18.9
15	do.	do.	120	275.3	13.8	20
16	do.	do.	170	64.9	7.4	8.8
17	do.	do.	200	36.7	10.1	3.6
36	Nonlinear	Spring Creek	32	24.8	6.7	3.7
41	do.	Sawhatchee Creek[b]	1.3	8.1	6.3	1.3
42	do.	do.	1.8	1	1.4	.8
43	do.	do.	2	3.8	3	1.3

[a] Zones refer to numbered stream reaches used in the report by Torak and others (1996) and are numbered separately for each boundary type. Ground-water inflow and channel length are adjusted to reflect the percentage of Zone 36 located downstream of the gage.

[b] Sawhatchee Creek is a tributary to the Chattahoochee River that enters downstream of gage.

Although results of digital modeling by Torak and others (1996) indicated that leakage from the main body of the lake occurred at a rate of 14.7 ft^3/s during October 1986, there is evidence that such leakage occurs at a much greater rate. Dye tracing performed by the U.S. Army Corps of Engineers near sinkholes and other solution structures in the bottom of the lake along the western shore near the dam indicated that lake water discharges from the lake through some of these features, which act as conduits. Some of this lake water enters these features at a distance from 0.5 to 0.7 mi upstream of the dam and flows through the Upper Floridan aquifer around the dam, emerging at a spring just downstream of the dam. Dye-tracing results indicate a travel time from about 5 to 7 hours for lake water to enter the spring downstream of the dam (James H. Sanders, Jr., U.S. Army Corps of Engineers, Mobile District, written commun., August 2001). This spring flow joins ground-water discharge from a diffuse area to the south, and discharges to a sinkhole, only to reemerge in a conduit in the bottom of the Apalachicola River, about 280 yards downstream of the dam. Flow rates measured using acoustic Doppler current profiling (ADCP; Lipscomb, 1995) at this conduit on October 21, 1999, and April 17, 2000, were about 140 and 210 ft^3/s, respectively. The USGS surface-water station on the Apalachicola River at Chattahoochee, Florida (station number 02358000), is located about 0.6 mi downstream of the dam and measures streamflow that consists of dam outflow and channel leakage, such as the leakage observed from this conduit, as well as diffuse channel inflow from the aquifer. Water chemistry of flow from the conduit indicated that the discharging water was a mixture of about 93 percent lake water and 7 percent ground water (Crilley, 2003), illustrating that most of the water discharging to the river at this boil originated as lake water. The ground-water component could be the result of mixing with ground water already present in the aquifer and/or hydrochemical interaction of the lake water with the limestone aquifer, altering the water chemistry toward a ground-water signature.

A comparison of lake-outflow measurements during periods of good record, beginning during August 2000 and excluding the outliers during March and April 2001, indicates that flows at the Apalachicola River gage are consistently about 550 ft^3/s higher than outflows from the dam (fig. 8). This streamflow gain most likely is due to lake leakage that flows around the dam through the ground-water flow system and enters the channel farther downstream, before the surface-water station. Hence, lake leakage was estimated as 550 ft^3/s. Likewise, lake outflow, as measured from the Apalachicola River gage, was decreased by the same 550 ft^3/s, which represents streamflow gain below the dam due to ground-water discharge that originated as lake leakage. The value for lake leakage and ground-water discharge to the river are reasonable because they are larger than the flow from the River Boil; additional inflows to the river from other boils or diffuse flow are possible along this stretch of the river. In general, it is expected that ground-water discharge and lake leakage would not change appreciably because the hydraulic gradient from the aquifer to the stream along this reach is relatively constant, although variation in lake leakage was observed in the two measured flows from the boil (140 and 210 ft^3/s).

Error Analysis

Limitations to the ground-water model could produce misleading results for this water-budget component. When the model was developed in the late 1980s, limited field data were available to properly assign aquifer properties and boundary conditions as model inputs near Lake Seminole. The paucity of field measurements in the Lake Seminole area limited the ability of the model to simulate local stream-lake-aquifer interaction; therefore, calibration was affected in this area on a regional scale. The model simulated drought and pumping conditions during October 1986, which may not be completely representative of the conditions that occurred during the current study. In addition, seasonal variations in ground-water conditions were not simulated by the steady-state model for October 1986 conditions.

The hydrologic framework of the model is based on a conceptualization of flow through a heterogeneous porous media rather than more complex flow through discrete conduits, which are characteristic of the karst limestone comprising the Upper Floridian aquifer in the Lake Seminole area. Hence, values of hydraulic conductivity that were assigned to the model represent the combined effects of discrete conduits and the limestone matrix. The distribution and connectivity of discrete conduits could have a large effect on the hydraulic conductivity and flowpaths in the Lake Seminole area. A careful examination of aerial photographs of the lake area before the dam was constructed identified sinkholes that, if interconnected, could provide preferred paths for ground-water flow. Sever (1965) identified numerous springs in the bottom of the lake, indicating the potential for such conduits for ground-water flow. A previous study involving the Upper Floridian aquifer in the Albany, Georgia, area used sinkhole and rock-cavity distributions to delineate hydraulic-property zones for simulation (Torak and others, 1993). Applying this approach to the Lake Seminole study area by delineating discrete flow conduits and evaluating hydraulic properties into a zoned arrangement for model input would improve the model. The only limitation to applying this approach to the Lake Seminole study area would be the availability of hydrogeologic information for characterization.

Despite the limitations of the digital model prepared by Torak and others (1996), it is the best method to quantify ground-water inflow to Lake Seminole and lake leakage. Several assessments can be made to test whether the flows computed by the model are reasonable. Ground-water flow estimates from the Cauchy-type boundaries that were simulated along the Spring Creek impoundment arm were verified by performing a streamflow survey between the Reynoldsville, Georgia (station number 02357150), and the measurement site at Brinson, Georgia (station number 02357050) (fig. 2), on September 15, 2000. Measurements of stream discharge were made at seven sites, and discharge was measured at four springs along this reach. All tributary flow entering Spring Creek occurred along spring runs that resembled creeks. Discharge from these runs was measured and traced to spring origins upstream of the main channel. Other discharge measurements that were performed in Spring Creek between the spring runs indicated diffuse channel leakage, which was attributed to ground-water inflow. Flow increased from about 31 ft^3/s near Brinson, Georgia, to about 131 ft^3/s near Reynoldsville, Georgia, a net gain of about 100 ft^3/s. Of this, about 52 ft^3/s was measured from four springs. The remaining flow was derived either from unmeasured, in-channel springs or from diffuse ground-water inflow along the wetted perimeter of the channel.

Although ground-water conditions on the day of the survey were not the same as the simulated conditions of October 1986, both conditions represent drought periods. The model simulated 26 ft^3/s of ground-water discharge along this stream reach (Cauchy-type boundary). This is almost one-quarter of the 100 ft^3/s obtained from the streamflow survey, indicating that the model could underrepresent ground-water inflow along this reach of the Spring Creek impoundment arm. Despite this apparent incongruity between model results and flow measurements, the model simulated 37 ft^3/s of streamflow gain between Iron City, Georgia, and Brinson, Georgia; on September 15, 2000, a gain of 31 ft^3/s was measured between Iron City and Brinson.

Simulated ground-water flows along the Flint River impoundment arm can be verified by examining increases in flow along the Flint River between Newton and Bainbridge, Georgia. Unfortunately, the USGS surface-water stations at Ichawaynochaway Creek near Newton, Georgia, and at the Flint River at Bainbridge, Georgia, were not in use at the time data were collected for model calibration during October 1986. The synoptic data collection, however, that was performed during October 23–28, 1986, to obtain measurements for calibration of the model (Torak and others, 1996) adequately documented the flow conditions of the time. The gain of the Flint River between Newton and the auxiliary surface-water station upstream of Bainbridge, Georgia (station number 02355700), excluding tributary flow from Ichawaynochaway Creek, was 290 ft^3/s. Simulated ground-water inflows along this reach of the Flint River were about 80 percent greater, 530 ft^3/s, indicating that although simulated ground-water inflows are the same order of magnitude as the measured inflows, the model may have overestimated the inflow of ground water along this reach during October 1986. Ground-water inflows could not be assessed in this manner along the Chattahoochee River impoundment arm because upstream and downstream surface-water stations did not exist to define a reach that was simulated with the model.

Measurements of streamflow gains from ground-water inflow along a gaged segment of the Flint River were compared with baseflow estimates derived from HYSEP to determine if it is reasonable to scale simulated ground-water inflow to the stream by monthly HYSEP-derived baseflows to mimic seasonal variations in ground-water inflows. Near the end of the study period during August 2001, the USGS reinstated the surface-water station on the Flint River at Bainbridge, Georgia (station number 02356000). This station, along with the gages on the Flint River at Newton, Georgia (station number 02353000), and on Ichawaynochaway Creek below Newton, Georgia (station number 02355350), allows the calculation of ground-water inflow along the reach of the Flint River between Newton and Bainbridge. This reach contains only one ungaged tributary, Big Slough, which flows only during extreme rainfall events and sometimes reverses flow. There was a linear relation between measured streamflow gains from ground-water inflow between Newton and Bainbridge and monthly HYSEP baseflow calculated at Newton for the period August 2001 through May 2003 ($r^2 = 0.50$). This suggests that proportional scaling of simulated ground-water inflow is a reasonable approximation for obtaining monthly ground-water inflow along the Cauchy-type boundaries to the lake. The slope of the relation (0.19) was about half as much as the scaling factor used to scale simulated ground-water flow (0.43).

The simulated ground-water flow across Cauchy-type boundaries not only contains the flow between Newton and Bainbridge but also the flow downstream from Bainbridge. This additional flow cannot fully account for the differences between the scaling factor and the slope of the relation between baseflow and ground-water inflow because the simulated ground-water inflow downstream from Bainbridge accounts for only 11 percent of the total simulated flow to the reach. Hence, it is likely that the scaling factor approximated by October 1986 conditions overestimates the ground-water inflows along the Flint River during the study period.

Because lake leakage was estimated as the average difference in streamflows between Jim Woodruff Lock and Dam and the stream gage on the Apalachicola River at Chattahoochee, Florida (station number 02358000), when streamflow measurements appeared reasonable, the estimates are subject to the same measurement errors as streamflows. Apalachicola River flows were reported as good in both water years 2000 and 2001, indicating that measurement errors were within 10 percent of the true discharge 95 percent of the time. Accuracy of flows was not reported for the gages operated by the U.S. Army Corps of Engineers at the dam. Lake leakage, set to a constant 550 ft^3/s, represents 4.1 percent of the average flow at the Apalachicola River gage. Hence, the estimate of lake leakage is within the error of the streamflow measurements used to determine it. Furthermore, there is no reason to suspect that all lake leakage discharges into the small section of river reach between the dam and the surface-water station on the Apalachicola River. Ground-water flowpaths delineated by Jones and Torak (2003) indicate that additional, unquantified lake leakage could discharge to the river downstream of the gage, or enter the aquifer, becoming part of the regional ground-water flow and not entering the river. Lake leakage is likely the least accurate component in the water budget.

Lake Evaporation

Six methods were used to determine lake evaporation: these are the energy-budget method and five empirical equations. The energy budget is one of the most accurate methods for determining evaporation, but the complexity in data collection makes it difficult to use (Harbeck and others, 1958). Although energy-budget estimates for lake evaporation are used in the development of the water budget for this study, those estimates are compared with long-term average annual pan evaporation measurements and estimates calculated using five empirical equations to determine which equations, if any, would be appropriate for use at Lake Seminole. The following empirical equations were used to compute lake evaporation: Priestley-Taylor, Penman, DeBruin-Keijman, and Papadakis, along with estimates calculated by the Georgia Automated Environmental Monitoring Network (GAEMN) using climate-station data and the Priestley-Taylor equation.

The energy-budget method requires large amounts of climatic and hydrologic data to implement; these data are labor intensive and expensive to collect. On the other hand, the better empirical equations require the calculation of stored heat to determine evaporation, and some empirical equations require only basic climate data. These data are readily available from two overwater climate stations that were installed on Lake Seminole during this study and are currently being managed by The University of Georgia as part of the GAEMN (Hoogenboom, 1996). In addition to providing near real-time climate data from the climate stations, GAEMN calculates lake evaporation at each climate station using the Priestley-Taylor equation and posts these estimates at *http://www.griffin.peachnet.edu/bae/* (accessed on June 12, 2004).

Evaporation estimates calculated using the energy-budget method were compared with those calculated using the five empirical equations listed above to determine whether there is a future need for the cost- and labor-intensive-data collection required to support calculation of lake evaporation by the energy-budget method.

Energy-Budget Method

The energy budget requires measurements of heat added to the lake by inflowing surface water and ground water, and direct precipitation, measurements of heat lost from the lake by surface-water outflow and lake leakage, the influence of net radiation, and measurements of the change in heat stored in the lake. The net addition of heat to the lake that does not result in an increase heat stored in the lake is then attributed to evaporation using the equation (Anderson, 1954):

$$Q_x = Q_s - Q_r + Q_a - Q_{ar} - Q_{bs} + Q_v - Q_h - Q_e, \qquad (2)$$

where

Q_x = change in stored energy;

Q_s = incident solar radiation;

Q_r = reflected solar radiation;

Q_a = incoming atmospheric longwave radiation;

Q_{ar} = reflected longwave radiation;

Q_{bs} = longwave radiation emitted by the lake;

Q_v = net energy advected by streamflow, ground water, and precipitation;

Q_h = energy removed from the lake as sensible heat; and

Q_e = energy used in evaporation.

All terms in equation 2 are expressed in calories per square centimeter per day (cal/cm^2/day).

The variables Q_s, Q_r, Q_a, Q_{ar}, and Q_{bs} were not measured directly; instead, net radiation ($Q_n = Q_s - Q_r + Q_a - Q_{ar} - Q_{bs}$) was measured with a net radiometer located at each of the two overwater climate stations (Cummings Access in Georgia and Sneads Landing in Florida) (fig. 2), and the resulting data were averaged to obtain an estimate of net radiation for the entire lake. Because of the fragility of the net radiometer's polyethylene domes, which became brittle in the hot climate, there were periods when the instrument was not functioning. Therefore, a relation was developed between net radiation and total-solar radiation measured at each station (R^2 = 0.8949 for the Cummings Access site, and 0.8958 for the Sneads Landing site; fig. 11) to determine daily net radiation from measurements of total-solar radiation during those periods when net radiation could not be measured.

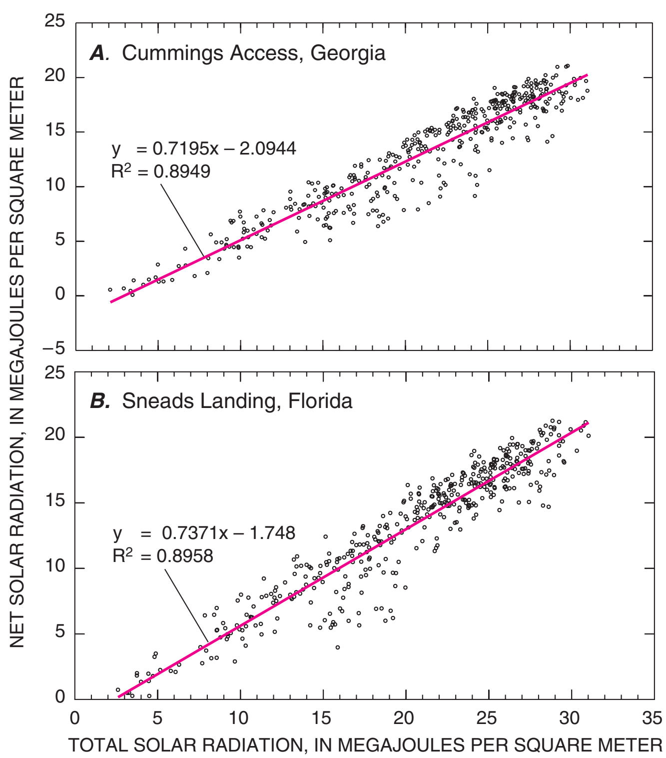


Figure 11. Regression showing relation between total solar radiation and observed net radiation at (*A*) Cummings Access, Georgia, and (*B*) Sneads Landing, Florida (*see* figure 2 for climate-station locations).

The net advected heat, Q_v, to Lake Seminole consists of heat added to the lake by surface- and ground-water inflow and precipitation, and heat exported from the lake by surface-water outflow and leakage. Heat added to the lake by surface-water inflow was calculated from measurements of daily average streamflow at surface-water stations and stream temperature, which was measured using a network of temperature probes that were installed in 26 vertical arrays to obtain temperature profiles at the most upstream locations of the impoundment arms (table 7). When temperature data from the most upstream vertical arrays in the impoundment arms were not available, data from alternate arrays positioned closest to the most upstream arrays were used. A complete description of thermal profiles, temperature-probe installation, and lake-temperature monitoring is given in the appendix.

Heat added to the lake by precipitation was estimated from the volume and temperature of precipitation, which was collected and recorded at the two overwater climate stations on the lake. It was assumed that the temperature of the precipitation was similar to the air temperature. Heat from precipitation at each climate station was averaged on 15-minute intervals as the product of the volume of precipitation, average air temperature, and specific heat and density of water. These values of heat were summed on a daily basis, and the results were averaged for the two climate stations to obtain the average daily heat advected by precipitation. Heat exported from the lake by surface-water outflow was calculated as the product of the discharge and temperature of the Apalachicola River at Jim Woodruff Lock and Dam, and the specific heat and density of water.

Heat advected to the lake from ground-water inflow and lost through lake leakage was calculated as the product of simulated rates of ground-water flow and lake-leakage, ground-water and lake temperature, and specific heat and density of water. Ground-water inflow was assumed to have a constant temperature of 20 degrees Celsius (°C), which was the temperature of ground water measured in springs located beneath the lake (in-lake springs; *see* fig. 4*A* in appendix) and was about the mean annual air temperature of the region (Sever, 1965; fig. 12; *http://www.griffin.peachnet.edu/bae/*—accessed on June 12, 2004).

Daily change in heat stored in the lake, Q_x, was calculated using daily lake volume, changes in lake temperature, volume of the lake, and the specific heat and density of water. Daily changes in heat stored in the lake were calculated separately for each impoundment arm and for the main body of the lake as follows. Initially, the average daily temperature change in each part of the lake was determined by averaging the changes in daily temperature from all working probes located in that part of the lake (table 8). These values were then multiplied by the estimated percentage of lake volume in each corresponding part of the lake. The resulting values were then summed and multiplied by the average daily lake volume (determined as the average of lake volumes estimated at the beginning and end of each day) and summed to obtain the daily change in stored heat for the lake. Probes 1-0 and 25-0, placed specifically in spring runs along the lake bottom, were excluded from calculations of lake temperature because these probes measure the temperature of local ground-water inflow, which is not representative of lake-water temperature in the impoundment arm. Problems estimating Q_x arise from missing data, resulting from temperature-probe failure and degree of representation of measured temperature and estimated volume for different parts of the lake (discussed in the appendix).

Daily volumes of each part of the lake were derived from lake stage by using a polynomial relation and lake bathymetry (fig. 9). Proportions of lake volume were calculated by using the length, width, and depth of channels in each part of the lake (table 8). Although volume estimates are not precise, weighting the temperature by the approximate volume proportions is an improvement over assigning equal weight

Table 7. Summary of surface-water inflows and outflows used in the water and energy budgets and the temperature arrays used to calculate daily average temperatures of surface-water inflows and outflows for energy budget.

Surface-water inflows and outflows	Primary temperature[a]	Backup temperature[b]
Flint River at Newton, Ga., and Ichawaynochaway Creek downstream of Newton, Ga.	Array 2	Probe 3-0
Spring Creek near Reynoldsville, Ga.	Array 24	Probe 23-0
Chattahoochee River near Columbia, Ala.	Array 6	Array 7
Apalachicola River at Woodruff Dam, Fla.	Probe 13-0	Probe 14-4

[a]*See* figure A2 for location of temperature probe arrays.

[b]Probe *n* – *m*, where *n* is array number and *m* is height above lake bottom.

Table 8. Summary of daily temperature arrays used to calculate daily average temperature of surface-water inflow and outflow for the main body of the Lake Seminole and impoundment arms.

Area of lake	Temperature arrays[a,b]	Percent of lake volume
Flint River impoundment arm	1–5; excluding probe 1-0	16.6
Spring Creek impoundment arm	20–24	3.2
Fishpond Drain	17, 18	1.7
Chattahoochee River impoundment arm	6–8	9.7
Main body of lake	9–16; 19; 25, 26; excluding probe 25-0	68.8

[a]*See* figure A2 for location of temperature probe arrays.

[b]Probe *n* – *m*, where *n* is array number and *m* is height above lake bottom.

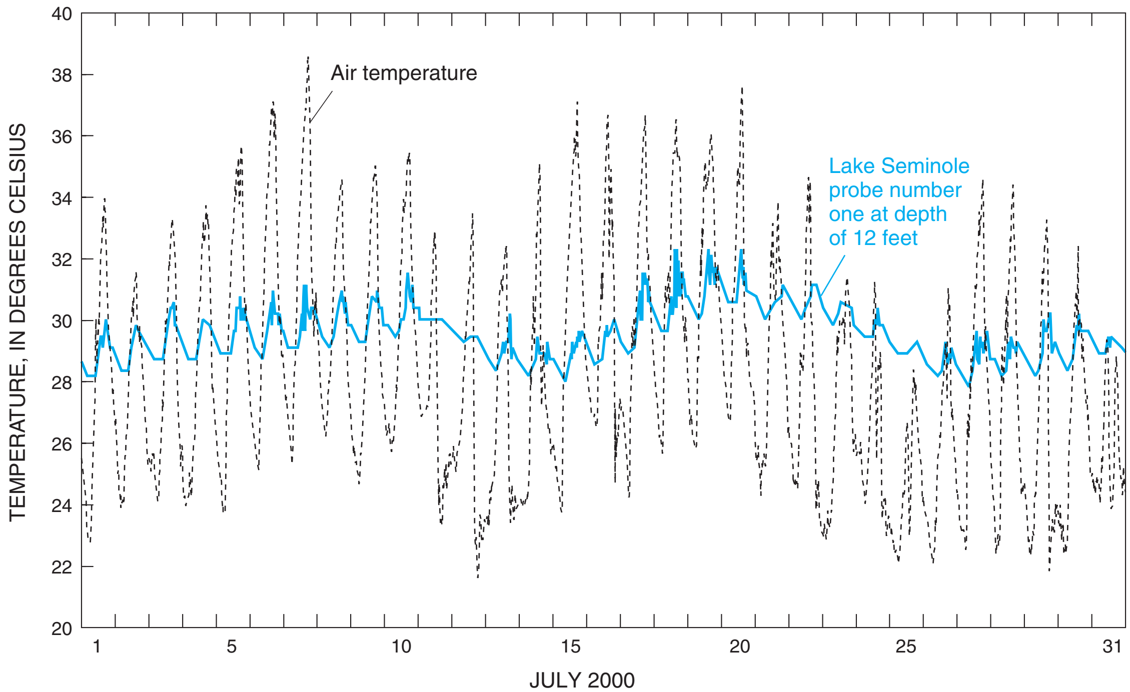


Figure 12. Relation between average air temperature and surface-water temperature at Lake Seminole, July 2000.

to each temperature-probe value. There was no weighting of lake temperature by water depth, even though there is a larger percentage of shallow areas in the lake than deep, because the water column was well mixed and there was little thermal stratification.

Sensible heat, Q_h, and the energy used for evaporation, Q_e, were not measured directly, but were obtained from estimates of the Bowen Ratio, *BR*, defined as the ratio of sensible heat to the energy used for evaporation. The Bowen Ratio is calculated using the equation (Bowen, 1926):

$$BR = Q_h / Q_e$$
$$= 0.00061\ P\ (T_0 - T_a)/(e_0 - e_a), \qquad (3)$$

where

P = barometric pressure in millibars;

T_0 = water-surface temperature in degrees Celsius;

T_a = air temperature at height of 2 meters above water surface in degrees Celsius;

e_0 = saturated vapor pressure at water surface in millibars; and,

e_a = vapor pressure at height of 2 meters above the water surface in millibars.

Using net radiation, Q_n, and the Bowen Ratio, *BR*, the energy-budget equation becomes:

$$Q_e (1 + BR) = Q_n + (Q_v - Q_x). \qquad (4)$$

The energy required for evaporation was measured in calories per square centimeter per day. Evaporation, however, is generally expressed as a rate, for example, volume per unit area per day, and the energy is converted to volume units using the equation:

$$E = Q_e / \rho L, \qquad (5)$$

where E is the evaporation rate expressed in centimeters per day, and

Q_e = energy used for evaporation, in calories per square centimeter per day;

ρ = density of water, in grams per cubic centimeter; and,

L = latent heat of evaporation, in calories per gram.

The latent heat of evaporation, L, varies with water-surface temperature according to the following relation (Maidment, 1992):

$$L = 2.501 - 0.002361 T_s, \qquad (6)$$

where

T_s = water-surface temperature, in degrees Celsius.

Water-surface temperature was obtained from temperature sensors located at the water surface at the climate stations rather than from the shallowest temperature probes in each array, which occasionally were out of the water because of low lake stage during drought conditions during the study period.

The volume of water evaporated from the lake then becomes:

$$E = (Q_n + Q_v - Q_x) / \rho L\ (1 + BR). \qquad (7)$$

Evaporation was calculated on a daily basis using daily average climate data for each climate station. The evaporation estimates from each climate station were then averaged to represent evaporation for the entire lake.

A balanced water budget was a necessary first step to achieving a balanced energy budget. A net loss or gain of water in the lake that is not defined by the storage term, ΔS, corresponds to a deficiency, or excess, of heat that would correspond to the energy available for evaporation calculated by the energy-budget method. Ground-water inflow and lake leakage were the most uncertain of all water-budget components because they were not measured directly; hence, flows and volumes derived from these sources were adjusted to balance the water budget. Volumes of water derived from these sources correspond to heat fluxes that were used in the energy-budget equations.

It was not possible initially to balance the water budget, which, in turn would be used to estimate lake evaporation with the energy-budget method, because the water budget requires that lake evaporation be known. Therefore, an iterative process was used to determine the amount of ground-water inflow and lake leakage required to balance the water budget and compute lake evaporation by the energy-budget method. An initial estimate of lake evaporation was used in the water budget to obtain initial values for lake leakage and ground-water inflow, which were used to balance the water budget. These values were then used in the energy budget to estimate evaporation for each day. The new evaporation estimates were then used to update estimates of lake leakage and ground-water inflow in the water budget. Lake leakage was adjusted before adjusting ground-water inflow, because there was more uncertainty in estimates of lake leakage than ground-water inflow. The new rates of lake leakage and ground-water inflow were used to obtain an updated evaporation rate; this process was repeated until the solutions from both energy and water budgets converged. No further changes in the estimates of lake leakage and evaporation were needed after three iterations.

Empirical Equations

Five empirical equations were used to estimate evaporation: Priestley-Taylor, Penman, DeBruin-Keijman, and Papadakis, and the Priestley-Taylor equation method used by GAEMN. GAEMN also uses the Priestley-Taylor equation to calculate evaporation; differences between the two Priestley-Taylor estimates arise in the manner in which lake-temperature measurements are used to determine heat stored in the lake. GAEMN uses lake temperature measured at the climate stations only; whereas the Priestley-Taylor calculation employs a more detailed heat budget, using temperature measurements collected from the temperature-probe network located throughout the lake (described in the appendix). Calculations of lake evaporation using each empirical method were compared with lake evaporation derived from the energy-budget method to determine the most accurate method, and to determine if GAEMN estimates of evaporation provide an accurate, cost-efficient method of determining evaporation from Lake Seminole.

Priestley-Taylor Equation

The Priestley-Taylor equation determines evaporation as a function of latent heat of evaporation and heat flux in a water body and is defined by the equation (Winter and others, 1995):

$$E = \alpha[s/(s+\gamma)][(Q_n - Q_x)/L], \tag{8}$$

where all terms have been defined previously except,

α = 1.26, Priestley-Taylor empirically derived constant, dimensionless, and

$s/(s+\gamma)$ = derived from slope of saturated vapor pressure at the mean air temperature; γ is the psychrometric constant, dimensionless.

Penman Equation

The Penman equation calculates evaporation based on the energy that is removed from the water-body surface as water vapor and is defined by the equation (Winter and others, 1995):

$$E = [s/(s+\gamma)](Q_n - Q_x) + [\gamma\,(s+\gamma)]\{[15.36(0.5 + 0.01U_2)]\,(e_0 - e_a)\}, \tag{9}$$

where all terms have been defined previously except,

$\gamma(s+\gamma)$ = derived from slope of saturated vapor pressure-temperature curve at the mean air temperature; γ is the psychrometric constant, dimensionless, and

U_2 = wind speed at 2 meters height above the lake surface, in meters per second.

DeBruin-Keijman Equation

The DeBruin-Keijman equation computes evaporation rates by using the moisture content of the air above a water body, heat stored in the water body, and the psychrometric constant, which is a function of atmospheric pressure and latent heat of evaporation, as (Winter and others, 1995):

$$E = [SVP / 0.95\,SVP + 0.63\gamma)]\,(Q_n - Q_x), \tag{10}$$

where SVP = saturated vapor pressure at mean air temperature, in millibars per degree Kelvin, and other terms are as defined previously.

Papadakis Equation

The Papadakis equation does not account for the heat flux at the water surface of the lake to calculate evaporation. Instead, it depends on the difference between the minimum and maximum saturated vapor pressure above the water body and is defined by the equation (Winter and others, 1995):

$$E = 0.5625[e_0 max - (e_0 min - 2)], \tag{11}$$

where

e_0 = saturated vapor pressure at maximum ($e_0 max$) and minimum ($e_0 min$) air temperatures.

Georgia Automated Environmental Monitoring Network

GAEMN estimates of lake evaporation are calculated using the Priestley-Taylor equation; estimates and are reported online at *http://www.griffin.peachnet.edu/bae*. Unlike the calculation of the Priestley-Taylor equation, however, where water temperature was determined by a series of temperature probes located throughout the lake, the GAEMN calculation uses temperature data from temperature sensors that are located at the two overwater climate stations. Temperature data from the climate stations represent the two types of lake habitat that are present, that is, open water (Sneads Landing, Florida) and submerged vegetation (Cummings Access, Georgia), which occur in nearly equal proportions. Therefore, evaporation calculations from the two climate stations were averaged to produce monthly evaporation rates for the lake, which were then compared with evaporation estimates obtained from energy-budget and empirical-equation computations.

Evaluation of Evaporation Estimates and Methods

Annual evaporation calculated by the energy budget for the period April 2000 through March 2001 was 67.2 inches. Monthly estimates range from a minimum of 3 inches during February 2001 to a maximum of 6.9 inches during June 2000, and averaged 5.6 inches per month during the study period (table 9). The long-term average annual pan evaporation estimate for the region is 65 inches (April 2000–March 2001) and is based on measurements from January 1959 to December 1978 (Farnsworth and Thompson, 1982). Annual evaporation calculated during the study using the energy budget is about 2.3 inches greater than long-term average pan evaporation.

Average monthly estimates of evaporation that were computed using the Priestley-Taylor equation ranged from 1.7 inches to 8 inches for the study period with an average monthly rate of 4.8 inches for the period April 2000 through March 2001 (table 9). Monthly differences in evaporation estimates computed using the Priestley-Taylor equation and the energy budget averaged 0.8 inch. The Priestley-Taylor equation underestimated evaporation by 9.7 inches during the study period, an average difference of 14 percent (table 9).

Average monthly estimates of evaporation that were computed using the Penman equation ranged from 1.6 to 6.9 inches for the study period, with an average monthly rate of 4.2 inches during the period April 2000 through March 2001 (table 9). Monthly differences in evaporation estimates computed using the Penman equation and the energy budget averaged 1.4 inches. The Penman equation underestimated evaporation by 17.2 inches during the study period, an average difference of 26 percent (table 9).

Average monthly estimates of evaporation that were computed using the DeBruin-Keijman equation ranged from 2.9 to 9.5 inches with an average monthly rate of 6 inches for the period April 2000 through March 2001 (table 9). Monthly differences in evaporation estimates computed using the DeBruin-Keijman equation and the energy budget averaged 0.4 inches. The DeBruin-Keijman equation overestimated evaporation by 5.1 inches during the study period, an average difference of 8 percent (table 9).

Average monthly estimates of evaporation using the Papadakis equation ranged from 2.7 to 6.7 inches with an average monthly rate of 4.6 inches for the period April 2000 through March 2001 (table 9). Monthly differences in evaporation estimates between the Papadakis equation and the energy budget average 1 inch. The Papadakis equation underestimated evaporation by 11.7 inches during the study period, an average difference of 17 percent (table 9).

Average monthly estimates of evaporation using the GAEMN climate stations ranged from 0.7 to 7.5 inches with an average monthly rate of 4.2 inches during for the period April 2000 through March 2001 (table 9). Monthly differences in evaporation estimates between the GAEMN climate stations and the energy budget average 1.4 inches. GAEMN underestimated evaporation by 17.1 inches during the study period, an average difference of 25 percent (table 9).

Energy-budget estimates of evaporation were largest in the early summer (June) and smallest in the winter (February) (fig. 13). In comparison, the Priestley-Taylor equation estimated evaporation to be largest during the spring (May) and smallest during the winter (December), as did the Penman equation, the DeBruin-Keijman equation, and the GAEMN method. The Papadakis equation estimated evaporation to be largest during the summer (July and August) and smallest during the winter (December). While evaporation estimates computed by the empirical methods, excluding the Papadakis equation, followed the same general seasonal pattern as values derived from the energy-budget method, evaporation computed by the Priestley-Taylor and DeBruin-Keijman equations tended to be larger than the values computed by the energy-budget method during the warm months and smaller than the energy-budget method during the cool months. The Penman-equation estimates were nearly equal to or less than estimates derived from the energy-budget method. Evaporation results from the Papadakis equation had a less discernable pattern; however, evaporation estimates from the Papadakis equation were always less than those obtained from the energy budget; except for February, when the computed evaporation from the energy-budget method was smaller.

Lake-water temperature is a common variable to the energy-budget method and the empirical equations and is used in computing Q_x, the change in heat stored in the lake. This variable influences the seasonality of evaporation estimates because changes in lake-water temperature affect the ease with which water is vaporized from the lake surface. The energy budget further refines evaporation estimates by taking into account the temperature of water entering and leaving the lake by ground- and surface-water inflow and lake leakage, respectively, thus, defining the differences between the empirical equations (Priestley-Taylor, DeBruin-Keijman, Penman, and GAEMN) and the energy-budget method. Each of these

Table 9. Evaporation rates calculated using the energy budget, those derived from empirical equations, and those values posted to the Georgia Automated Environmental Monitoring Network (GAEMN) at *http://www.griffin.peachnet.edu/bae/*

[DBK, DeBruin-Keijman; NA, not applicable; pan, pan evaporation. Evaporation rates in inches per month]

Month and year	DBK equation	Penman equation	Priestly-Taylor equation	Papadakis equation	GAEMN	Energy budget	Pan
Apr. 2000	8.0	5.3	6.1	4.5	5.3	6.8	6.5
May. 2000	9.5	6.9	8.0	5.9	7.5	6.0	7.3
June 2000	8.4	6.2	7.2	5.6	6.9	6.9	7.8
July 2000	8.4	6.4	7.4	6.7	7.1	6.9	7.3
Aug. 2000	7.7	5.8	6.8	6.6	6.3	6.1	7.0
Sept. 2000	5.7	4.1	4.8	4.7	4.4	6.3	6.5
Oct. 2000	6.2	4.3	4.9	5.8	3.8	5.7	5.4
Nov. 2000	3.3	2.1	2.3	3.7	1.6	5.0	3.4
Dec. 2000	2.9	1.6	1.7	2.7	0.7	4.4	2.6
Jan. 2001	3.6	2.0	2.1	3.0	1.4	3.6	2.6
Feb. 2001	3.6	2.3	2.6	3.2	2.1	3.0	3.3
Mar. 2001	5.0	3.1	3.5	3.1	3.0	6.7	5.2
Apr. 2001	7.6	5.1	5.9	4.4	5.3	5.8	6.5
May 2001	8.7	6.2	7.2	5.3	6.7	6.0	7.3
June 2001	7.1	5.1	6.1	4.4	5.9	5.9	7.8
July 2001	7.8	5.7	6.8	4.8	6.5	5.7	7.3
Aug. 2001	6.9	5.0	6.0	4.9	5.6	5.3	7.0
Sept. 2001	6.2	4.5	5.2	4.5	4.5	6.7	6.5
Minimum monthly evaporation (inches)	2.9	1.6	1.7	2.7	0.7	3.0	2.6
Maximum monthly evaporation (inches)	9.5	6.9	8.0	6.7	7.5	6.9	7.8
Study period evaporation (April 2000–September 2001)	116.5	81.6	94.6	83.8	84.7	102.6	107.4
Annual evaporation (April 2000–March 2001)	72.3	50.0	57.5	55.5	50.1	67.2	65.0
Average monthly evaporation (April 2000–March 2001)	6.0	4.2	4.8	4.6	4.2	5.6	5.4
Difference from energy budget (April 2000–March 2001)	5.1	17.2	9.7	11.7	17.1	NA	2.2
Monthly average difference from energy budget	0.4	1.4	0.8	1.0	1.4	NA	0.2
Monthly average percent difference from the energy budget	8	26	14	17	25	NA	3
Monthly average percent difference from long-term average annual pan evaporation	**11**	**23**	**11**	**15**	**23**	**3**	**NA**

equations takes into account the meteorological components of evaporation. The volume and temperature of water that enters and leaves the lake, however, have a substantial affect on the amount of lake water that is evaporated; the energy budget accounts for this with the term Q_v, net heat advected to the lake by streamflow, ground water, and precipitation.

Variation in evaporation estimates between the empirical equations is expected because each equation uses different climatic variables to calculate evaporation. For example, all the equations require some information on heat flux in the lake, except for the Papadakis equation. The Priestley-Taylor equation relies on an estimate of latent heat, the Penman on wind speed, and the DeBruin-Keijman on saturated vapor pressure. Although there is some variation in evaporation estimates between the methods, seasonality of the estimates is common to all methods (fig. 13).

Water-Budget Summary

Inflow to Lake Seminole is dominated by surface-water flows (table 10, fig. 14). Streamflow accounted for 81 percent of the inflow to Lake Seminole. The Chattahoochee and Flint Rivers contributed most of the total inflow, 47 and 28 percent, respectively; tributary flow to the Flint River from Ichawaynochaway Creek contributed 4 percent; Spring Creek contributed the remaining 2 percent. Ground water accounted for 18 percent of lake inflow. The contribution of ground water to the lake actually was slightly higher than this value because streamflow measurements at gages located in the impoundment arms of the Flint River and Spring Creek includes a component of ground-water inflow to the stream channels. This affects only that part of ground- and surface-water inflow

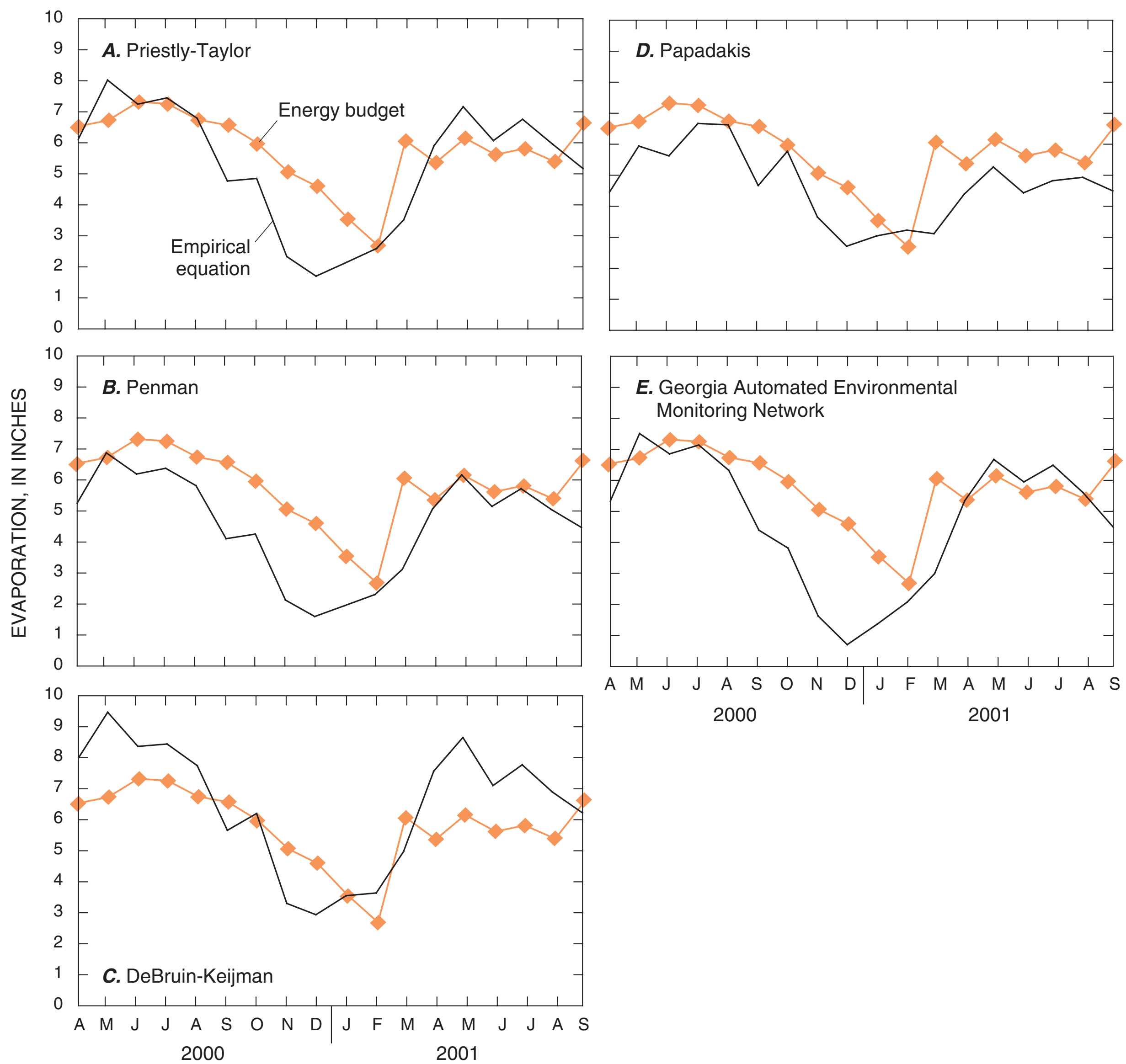


Figure 13. Monthly estimates of evaporation computed using the energy budget and empirical equations: (*A*) Priestley-Taylor, (*B*) Penman, (*C*) DeBruin-Keijman, (*D*) Papadakis, and (*E*) Georgia Automated Environmental Monitoring Network, April 2000–September 2001.

Table 10. Monthly water-budget variables and cumulative error for Lake Seminole.

[GW, ground water; -, outflows greater than inflows. All values in cubic feet per second]

Month and year	Inflows to Lake Seminole					Total inflows	Outflows from Lake Seminole			Total outflows	Change in Storage	Budget difference
	Spring Creek	Chattahoochee River	Flint River	Ichawaynochaway Creek	Precipi-tation		Apalachicola River	Lake leakage	Evapo-ration			
Apr. 2000	361	9,628	5,618	609	166	19,539	16,780	550	345	17,676	-614	2,477
May 2000	202	5,148	1,944	200	9	9,443	7,863	550	307	8,720	-557	1,280
June 2000	121	3,986	1,217	86	130	7,314	4,276	550	351	5,177	314	1,823
July 2000	121	3,113	1,305	131	149	6,137	4,567	550	343	5,460	-184	861
Aug. 2000	97	3,608	1,283	105	89	6,470	5,256	550	312	6,118	-153	504
Sept. 2000	114	3,020	2,105	221	318	7,095	5,339	550	321	6,210	663	222
Oct. 2000	136	2,778	1,426	176	22	5,866	5,109	550	292	5,951	-322	238
Nov. 2000	146	3,342	2,549	305	193	8,033	5,811	550	256	6,617	468	948
Dec. 2000	175	5,929	2,685	449	121	11,051	9,751	550	227	10,528	169	354
Jan. 2001	328	7,517	4,575	706	93	16,362	14,140	550	183	14,873	-211	1,699
Feb. 2001	304	6,126	4,086	538	41	14,317	11,439	550	151	1,2140	342	1,835
Mar. 2001	1,274	28,881	16,977	2,424	555	58,457	56,640	550	341	57,532	112	813
Apr. 2001	1,287	13,017	10,074	1,802	72	31,903	30,313	550	294	31,158	-224	969
May 2001	406	5,535	2,990	518	52	11,952	11,012	550	308	11,869	-199	282
June 2001	392	9,776	5,573	739	457	20,709	18,049	550	302	18,901	416	1,392
July 2001	320	4,672	3,291	274	439	10,976	10,602	550	293	11,445	-246	-223
Aug. 2001	237	4,556	2,058	199	245	9,167	9,035	550	268	9,853	-315	-371
Sept. 2001	183	3,026	2,010	266	156	7,072	6,623	550	340	7,513	-48	-393
Study period average	345	6,870	3,987	542	184	14,548	12,937	550	291	13,763	-37	809
Percent of Lake Seminole water budget	**2**	**47**	**28**	**4**	**1**		**89**	**4**	**2**			**4**

arms and not total lake inflow. Direct precipitation on the lake made up only a minor portion of inflow, about 1 percent, except during storm events when lake inflow from precipitation made up 10 to 20 percent of total lake inflow (fig. 14).

Chattahoochee and Flint River flows varied seasonally (highest in winter and spring) with the highest flows during March 2001, the month with the highest local precipitation, and lowest in the fall and summer (table 10). The Chattahoochee and Flint Rivers had similar patterns of monthly flows during the period of study even though Chattahoochee River flows are controlled by several large reservoirs, and the Flint River has only two small reservoirs upstream of Lake Seminole. In comparison, seasonal flow variation on Spring and Ichawaynochaway Creeks is more subdued than on the Flint and Chattahoochee Rivers (table 10). Streamflow from Spring Creek exhibited less monthly variation in flows than the Chattahoochee and Flint Rivers, and higher flows in the winter and spring than in the summer and fall (table 10, fig 14).

Monthly patterns in ground-water flows (fig. 10) were somewhat similar to the variation in streamflow for the Chattahoochee and Flint Rivers, which was expected because monthly patterns of ground-water inflow were derived from the calculated baseflows of these rivers. Monthly ground-water flows, however, were lower than monthly streamflow. During periods of high inflow to the lake and high streamflow, ground water comprised a smaller percentage of total inflow to the budget than the other inflows (table 10, fig. 14).

Monthly average inputs from direct precipitation constituted a small part of the total inflow (fig. 14). The normal pattern of seasonal precipitation was not observed during the study; instead, the largest amounts occurred during September 2000 and March, June, and July 2001 (table 10, fig. 15).

Outflow from Lake Seminole was dominated by flow from Jim Woodruff Lock and Dam, which accounted for about 89 percent of lake outflow (table 10, fig. 14). Lake leakage constituted about 4 percent of the budget. Evaporation, determined by the energy budget, made up about 2 percent of total lake outflow. The small change in lake storage that occurred during the study period could not accommodate the 4-percent deficiency in outflow, compared with total inflow. This difference between inflow and outflow represents measurement error, as described previously, as it relates to stream-discharge measurements, ground-water inflow, precipitation, and lake evaporation and leakage.

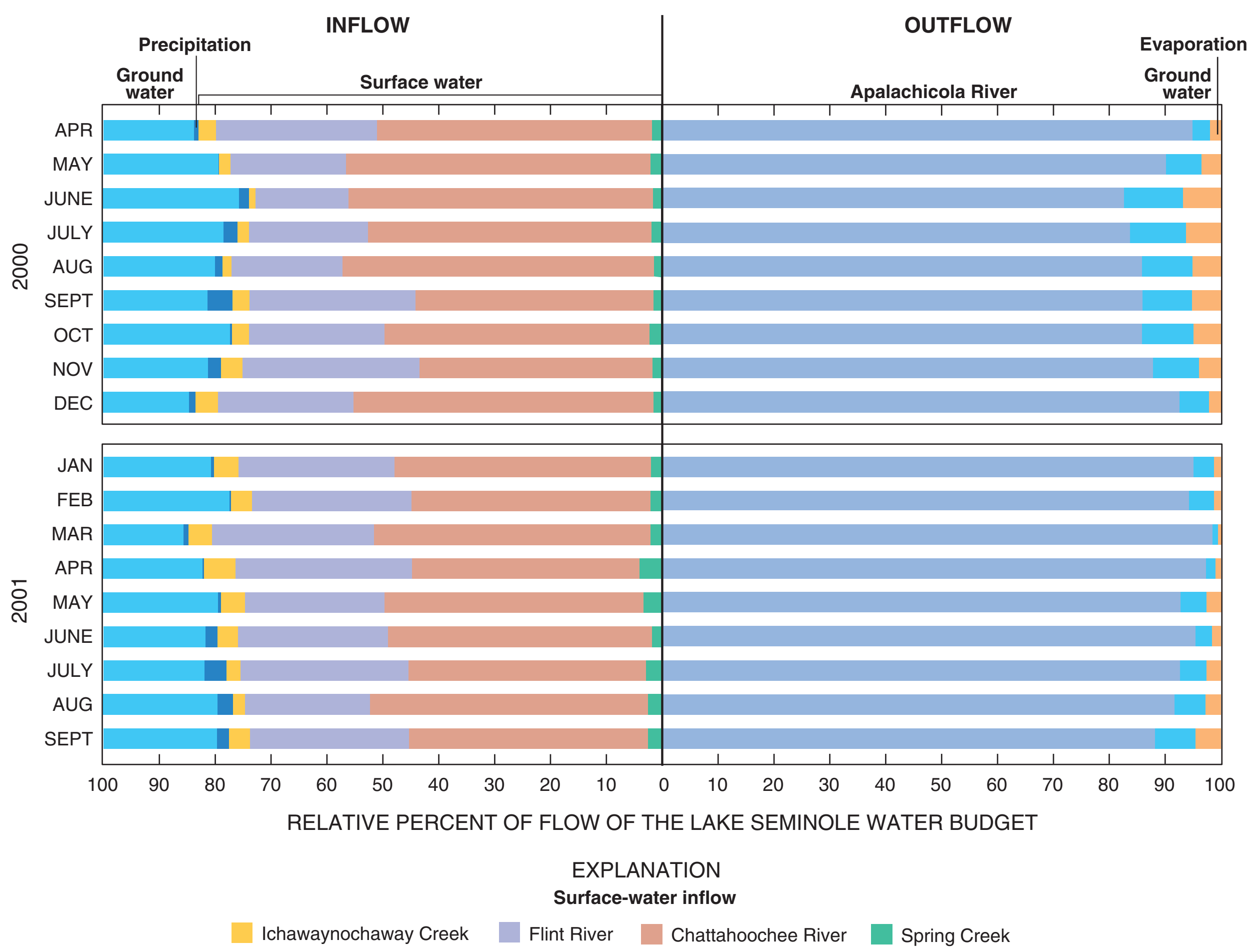


Figure 14. Percent contribution of Lake Seminole water-budget components.

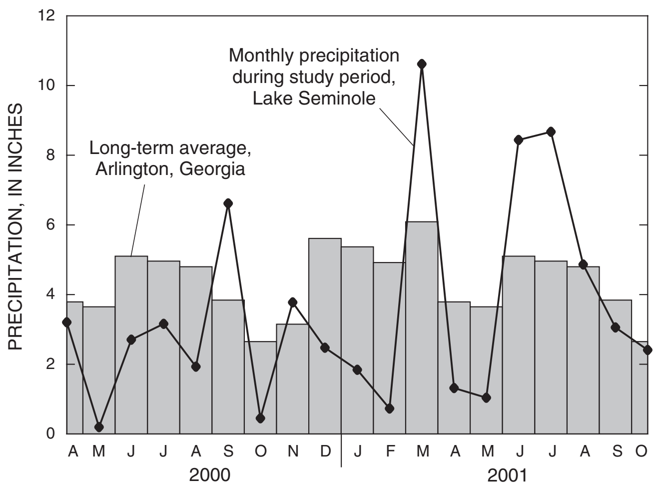


Figure 15. Monthly average precipitation during the study period compared with long-term average precipitation at climate station in Arlington, Georgia, 1960–2002 (*see* figure 2 for climate-station location).

There were fewer monthly variations in lake outflow than inflow, which was nearly constant from June through November 2000 (table 10). Differences in the pattern between surface-water outflow and inflow from the Chattahoochee and Flint Rivers are due largely to fluctuations in reservoir storage and release during the study period. Monthly lake leakage did not show a seasonal pattern and was estimated to be 550 ft3/s (table 10).

Evaporation estimates calculated by the energy budget exhibited a seasonal pattern in response to energy inputs, highest during the summer and lowest during the winter, with the highest rate occurring during June 2000 (6.9 inches) and the lowest occurring during February 2001 (3 inches) (fig. 13 and table 9). The methods used to calculate evaporation included, the energy budget, which was used in calculating the water budget of Lake Seminole, and five empirical equations: Priestley-Taylor, Penman, DeBruin-Keijman, Papadakis, and GAEMN. There was considerable variation among evaporation estimates computed with the empirical equations, with differences in the estimates between the equations and the energy-budget method ranging from an underestimation of 5.1 inches and overestimation of 17.2 inches during the period April 2000 through March 2001 (table 9). Because evaporation accounts for only 1 percent of the lake outflow, its effect on the total water in the budget ranges from 0.3 to 0.71 percent (table 11), depending on the method used to estimate evaporation. This percentage of total water is insignificant compared with the measurement error for surface water, which usually ranges from 5 to 15 percent, depending on streamflow-measurement accuracy.

Water-Budget Error

Differences between monthly inflow and outflow, taking into account changes in storage of the lake, ranged from a deficit of 393 ft^3/s, or 1.4 percent of the budget, during September 2001 to a surplus of 2,477 ft^3/s, or 8.7 percent of the budget, during April 2000 (table 10). Most of the error in the water budget is attributed to surface-water flow, which accounts for 81 percent of the inflow and 89 percent of the outflow in the budget, with measurement errors that range from 5 to 15 percent, depending on flow conditions.

Unlike other studies that developed lake water budgets (Lee and Swancar, 1996; Swancar and others, 2000), the proportion of surface-water flow to the other components of the Lake Seminole water budget (ground water, precipitation, and evaporation) is so large that these other components contribute only minor amounts of water to lake inflow and outflow. Because surface water encompasses such a large percentage of inflow and outflow for Lake Seminole, other hydrologic components seemingly are insignificant to the development of a water budget. While quantifying these hydrologic components is essential for developing a better understanding of reservoir-stream-aquifer interaction and for improving hydrologic monitoring in the region, error in their measurement is inconsequential to the Lake Seminole water budget. A sensitivity analysis of the water budget to changes in each component was used to illustrate the importance of individual measurement accuracy.

Sensitivity Analysis

The sensitivity analysis demonstrates the relative importance of individual components to the overall accuracy of the water budget. Accurate measurement of precipitation and simulation of ground-water inflow becomes less important when the volume of water derived from these water-budget components is compared with the volume of surface water that flows into Lake Seminole. If each component is varied by the same percentage, then their effect on the water budget fluctuates according to their relative contribution, or importance to lake inflow or outflow. For example, if ground-water flow and precipitation are increased by 50 percent each, then the total inflows to the water budget increase by 5.3 and 0.4 percent, respectively; but if surface-water flow is increased by 50 percent, then the total inflows to the water budget increases by about 20 percent (fig. 16). Because the water budget is dominated by surface water, better accuracy in estimating inputs to the water budget would be attained by improving the accuracy by which surface-water inflows are measured. Improvements to estimating ground-water inflow or measuring lake precipitation would result in minimal, if not insignificant, improvements to quantifying lake inflow.

Table 11. Average study period percent difference in water budget of Lake Seminole comparing empirical equations, long-term average annual pan evaporation, and the energy budget.

[DBK, DeBruin-Keijman; GAEMN, Georgia Automated Environmental Monitoring Network; pan, pan evaporation]

Month and year	DBK equation	Penman equation	Priestley-Taylor equation	Papadakis equation	GAEMN	Pan[a]
Apr. 2000	0.45	0.37	0.11	0.61	0.35	0.08
May 2000	1.60	.19	.81	.31	.53	.73
June 2000	.85	1.13	.17	1.65	.53	.94
July 2000	.99	.79	.13	.54	.14	.56
Aug. 2000	.77	.69	.05	.07	.30	.70
Sept. 2000	.82	2.02	1.51	1.59	1.80	.19
Oct. 2000	.20	1.41	.92	.16	1.77	.26
Nov. 2000	1.39	2.27	2.12	1.13	2.66	1.22
Dec. 2000	.78	1.41	1.36	.88	1.82	.90
Jan. 2001	.03	.57	.51	.20	.77	.33
Feb. 2001	.42	.18	.05	.24	.29	.13
Mar. 2001	.14	.31	.28	.31	.32	.13
Apr. 2001	.35	.08	.07	.19	.03	.12
May 2001	1.12	.05	.48	.34	.27	.53
June 2001	.40	.15	.11	.35	.08	.52
July 2001	.94	.01	.48	.39	.36	.71
Aug. 2001	.79	.15	.31	.21	.13	.88
Sept. 2001	.28	1.47	.99	1.45	1.46	.10
Average study period difference in water budget using empirical equation in lieu of energy budget, in percent	0.30	0.71	0.31	0.56	0.60	0.17

[a]Long-term average annual pan evaporation estimates for the region are based on measurements from January 1959–December 1978 (Farnsworth and Thompson, 1982)

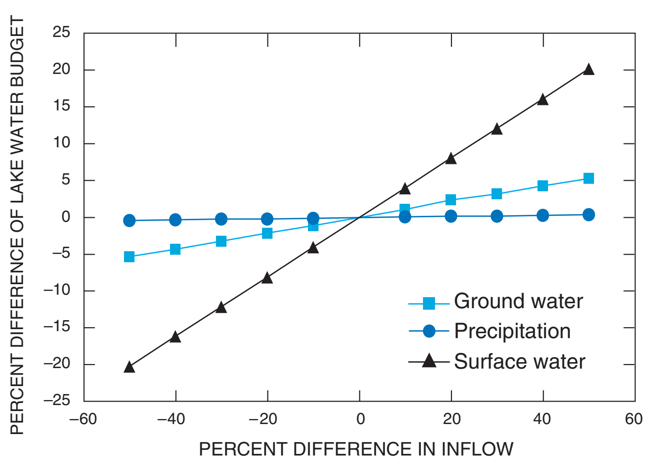


Figure 16. Relative sensitivity of the Lake Seminole water budget to component inflow.

A similar sensitivity analysis of lake outflow components indicates that surface-water flow dominates the water budget to such an extent that a rationale for quantifying lake evaporation and leakage more accurately than by the methods used in this study simply cannot be substantiated. That is, very little improvement in the accuracy of lake outflow would be gained by improving the current methods used to compute lake evaporation and leakage. This is not to say, however, that lake evaporation and leakage are not important to understanding reservoir-stream-aquifer interaction. If evaporation and lake leakage decrease by 50 percent, then the water budget decreases by 0.8 and 1.5 percent, respectively; however, if surface-water outflow decreases by 50 percent, the water budget decreases by more than 21 percent (fig. 17). Thus, accuracy in measuring surface-water outflow is critical for improving the accuracy of the water budget for Lake Seminole.

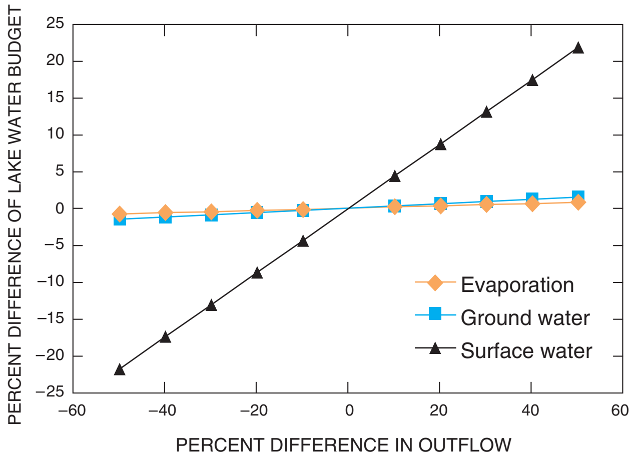


Figure 17. Relative sensitivity of the Lake Seminole water budget to component outflow.

Summary

The stream-aquifer-lake flow system consists of inflow to Lake Seminole by precipitation, the Chattahoochee and Flint Rivers, Spring and Ichawaynochaway Creeks, Fishpond Drain, and other ungaged streams and ground-water flow from the Upper Floridan aquifer. Outflow from the lake consists of lake evaporation and leakage to the aquifer and flow through Jim Woodruff Lock and Dam.

Water-budget calculations for the period from April 2000 to September 2001 indicate that surface-water inflow and outflow dominate the water budget, comprising 81 percent of inflows and 89 percent of outflows. Ground water makes up 18 percent of total inflow, based on estimates of October 1986 flow rates that were modified by hydrograph-separation techniques. Lake leakage comprises 4 percent of total outflow and was derived from streamflow measurements at Jim Woodruff Lock and Dam and on the Apalachicola River, at the surface-water station at Chattahoochee, Florida. Precipitation was estimated as 1 percent of total inflow, and evaporation made up about 2 percent of total outflow. Errors in streamflow measurements represented a component of flow that is larger than the other (minor) components of the water budget, namely, ground-water inflow, lake leakage, evaporation, and precipitation. Because surface water dominates the water budget, alternative, economical, and efficient methods were used to determine ground-water inflow, lake leakage, and evaporation rates for Lake Seminole.

Comparisons of monthly evaporation rates calculated by the energy-budget method with estimates obtained from five empirical equations—Priestley-Taylor, Penman, DeBruin-Keijman, Papadakis, and the estimate provided by the Georgia Automated Environmental Monitoring Network—and reported long-term pan evaporation rates indicate that the energy-budget method provided the closest match between evaporation rates and pan evaporation. Evaporation estimates derived from the empirical equations were less accurate, in comparison with the energy-budget method, ranging from 8 to 26 percent of the rates obtained from the energy budget. The dominance of surface water in the Lake Seminole water budget, however, made the choice of technique used to calculate lake evaporation superfluous; the effect of the evaporation estimate on the total water volume in the budget ranged from 0.3 to 0.71 percent, depending on the evaporation method used. This percentage variation in total water in the water budget is within the measurement error for streamflow. Twenty-percent variation in streamflow caused about an 8-percent change in total water in the water budget.

The evaporation method used by GAEMN is the most time- and cost-efficient means of determining lake evaporation on a monthly basis for the water budget of Lake Seminole. GAEMN uses the Priestley-Taylor empirical equation, which provides evaporation estimates with sufficient accuracy to be used in water-budget computations. Additionally, GAEMN provides the climate data needed for the evaporation computations, performs the computations, and makes these results available online (*http://www.griffin.peachnet.ed/bae/*).

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Appendix — Methods and Instrumentation

The water budget of Lake Seminole can be described as a means of accounting for hydrologic components that contribute to the change in lake storage, which can be expressed as the difference between the volume of water flowing into the lake and the volume of water flowing out of the lake. Individual components used to construct the water budget (fig. 7, eq. A1) can be obtained using various methods. The most complicated of the components to measure accurately is evaporation because of data-collection needs. The energy-budget method, the most accurate method for estimating evaporation, requires large amounts of climatic and hydrologic data to implement; these data are labor intensive and expensive to collect. On the other hand, some empirical equations, which are less accurate than the energy budget, require only basic climate data, which are readily available from two overwater climate stations that were installed on Lake Seminole during this study. The accuracy of several methods were compared and weighed against cost and effort to determine the method of estimating evaporation that would be used in the construction of the water budget of Lake Seminole. Excluding ground water, which was determined using a ground-water flow model developed for the Lake Seminole study area (Torak, 1993a,b), each of the components was measured directly.

$$\Delta S = P + SW_{in} - SW_{out} + GW_{in} - GW_{out} - E \qquad \text{(A1)}$$

ΔS = change in lake storage

P = precipitation

SW_{in} = surface-water inflows

SW_{out} = surface-water outflows

GW_{in} = ground-water inflows

GW_{out} = lake leakage

E = evaporation

Overland flow was virtually nonexistent and was ignored as part of the water budget because the Dougherty Plain was internally drained due to karst solution features. The size and irregular geometry of Lake Seminole adds complexity to evaporation estimates, which are the main focus of the water budget, because evaporation ultimately could be an important component of the overall budget.

Climate Stations and Variables

Two overwater climate stations were installed on Lake Seminole (figs. 2; A1) to record climatic variables used to estimate evaporation (table A1). The University of Georgia as part of the Georgia Automated Environmental Monitoring Network (GAEMN) (*http://www.griffin.peachnet.edu/bae/*) operates and maintains these stations. Measurements of net radiation,

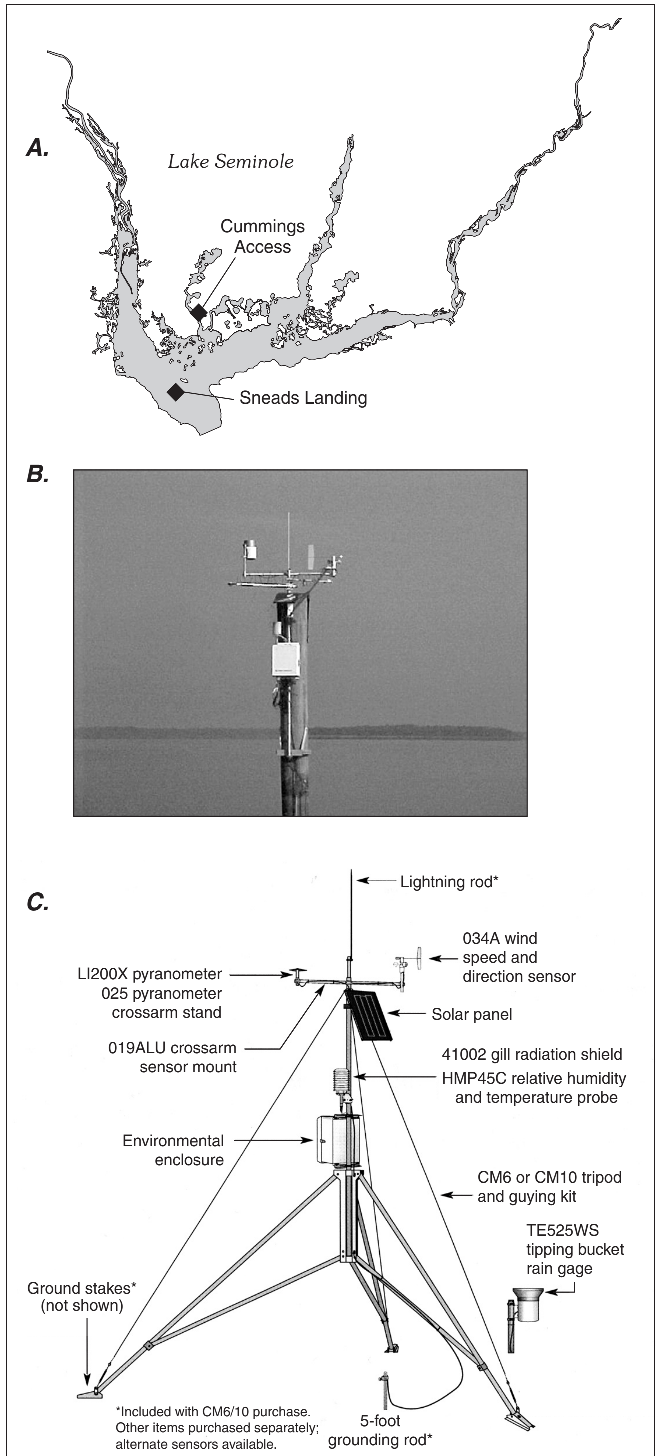

Figure A1. (*A*) Location of climate stations installed over water on Lake Seminole; (*B*) climate-station installation at Sneads Landing, Florida, on Lake Seminole; and (*C*) schematic of instruments used in overwater installation of climate stations (from *www.campbellsci.com*, accessed on November 17, 1999).

Table A1. Instrumentation of the climate stations installed on Lake Seminole.

[ºC, degree Celsius; %, percent; NA, not applicable; m/s, meter per second; mbar, millibar; W/m^2, watts per square meter; Wm^{-2}mV^{-1}; watts per square meter per millivolt]

Instrument type	Model	Purpose	Unit
Datalogger	CR10X	Data collection and storage	NA
Tripod and grounding kit	CM10	Instrument support	NA
Temperature and humidity probes	HMP45C	Collection of temperature and humidity measurements	°C and %
Radiation shield	41002	Shields temperature and humidity probes from direct sunlight	NA
Pyranometer	LI200X	Measurement of solar radiation	W/m^2
Pyranometer base and leveling fixture	LI2003S	Support and leveling of pyranometer	NA
Wind sensor	034A	Measurement of wind speed and direction	m/s
Rain gage	TE525	Precipitation measurement	inch
Barometric pressure sensor	CS105	Measures barometric pressure	mbar
Net radiometer	Q-7.1	Generates a millivolt signal proportional to net radiation	Wm^{-2}mV^{-1}

wind speed and direction, barometric pressure, water surface and air temperature, precipitation, and humidity were made each second and summarized at 15-minute intervals and at midnight daily; monthly averages are shown in table A2. Data were recorded by a datalogger and transmitted by cell phone to a microcomputer at the College of Agriculture and Biological Sciences, Georgia Experiment Station (Georgia Automated Environmental Monitoring Network, 2002).

Two ecological conditions exist in the lake that require separate instrumentation of corresponding climatic conditions. Submerged vegetation covers nearly half of the lake area and affects not only surface-water temperature but also the humidity and net radiation in areas densely vegetated. To account for differences related to these conditions, the climate stations were installed to record data over both openwater and highly vegetated areas. The climate station at Sneads Landing (Florida) was installed over open water, and the climate station at Cummings Access (Georgia) was installed over submerged vegetation. The data from these two climate stations were averaged to determine the climatic conditions for the entire lake.

Thermal Profiles and Temperature Probe Installation

Temperature probes were installed in Lake Seminole to monitor continuously the temperature of the lake in order to calculate the stored heat in the lake. Vertical arrays of temperature probes were distributed at 26 sites to measure the temperature of surface-water inflow and outflow and springflow (ground-water inflow), and to determine the vertical temperature distribution (fig. A2; table A3). Locations for monitoring lake temperature (fig. A2) were selected to provide a somewhat uniform areal distribution; the specific location of each vertical array was dependent on the availability of permanent attachment points for the cable containing the array probes, such as docks, piers, large trees, and channel markers. Some arrays were intentionally located at spring openings or in spring runs—that is, in channels along the lake bottom where spring discharge flows—in order to measure the temperature of ground-water inflow to the lake.

Vertical-temperature profiles were obtained at each array site using from three to five probes, depending on depth, distributed uniformly throughout the water column (fig. A3). About 100 probes were installed during March 3–9, 2000. Onset Optic StowAway® Temp temperature probes were installed, which have a range of -5 to 37°C and an accuracy of 0.2°C. Water temperature was recorded at 24-minute intervals. The probes were affixed to a steel cable that was attached to a stationary structure above or near the water surface such that the position of each probe in the water column did not vary with time. An exception occurred for array 6, which was relocated on April 23, 2001, and for array 8, which was relocated on September 14, 2000. Probe height in the water column was referenced by vertical distance above the lake bottom, and the probes were distributed over somewhat equal vertical intervals such that the top probe was just below the water surface. Individual temperature probes are identified by the array location number (1 to 26), followed by a dash (-), and then the height above the lake bottom in feet. For example, probe 1-12 identifies the probe at location 1, which is 12 feet above the lake bottom (table A3).

Table A2. Monthly statistics for average climate conditions on Lake Seminole.

[mbar, millibar; °C, degree Celsius; m/s, meter per second; cal/cm²/day, calories per square centimeter per day]

Date		Vapor press (mbar)	Average surface temperature (°C)	Average wind (m/s)	Regression net radiation (cal/cm²/day)	Average air temperature (°C)
Apr. 2000	Average	14.34	22.49	2.53	356.92	18.82
	Median	14.41	22.38	2.27	390	19
	Minimum	6.94	20.65	1.38	88.32	11.16
	Maximum	22.4	24.47	5.19	466.56	22.95
	Standard deviation	3.87	.9	.92	107.39	2.93
May 2000	Average	20.57	28.46	2.17	413.32	25.28
	Median	20.02	28.99	2.21	423.94	25.19
	Minimum	12.8	23.62	1.47	269.79	21.6
	Maximum	27.27	30.3	3.42	482.37	28.2
	Standard deviation	3.79	1.55	.48	45.88	1.78
June 2000	Average	23.83	30.2	2.09	367.69	26.74
	Median	25.53	30.22	2.07	389.83	26.74
	Minimum	14.42	28.81	1.51	187.48	24.09
	Maximum	27.7	31.96	2.98	483.92	29.28
	Standard deviation	3.65	.8	.4	93.69	1.29
July 2000	Average	26.1	31.69	1.89	359.74	28.22
	Median	26.22	31.54	1.9	387.86	27.91
	Minimum	21.94	30.24	1.29	107.22	24.65
	Maximum	29.18	33.74	2.94	452.47	31.32
	Standard deviation	1.87	.92	.36	88.29	1.86
Aug. 2000	Average	26.16	31.16	1.71	325.89	27.65
	Median	26.87	31.11	1.67	336.99	27.76
	Minimum	20.59	29.93	1.21	183.69	24.8
	Maximum	29.6	32.45	2.44	418.45	30.07
	Standard deviation	2.38	.7	.28	70.26	1.28
Sept. 2000	Average	23.64	27.8	2.08	252.49	24.61
	Median	25.28	27.7	1.91	297.72	25.6
	Minimum	14.22	24.62	1.03	20.95	19.03
	Maximum	29.64	30.11	3.94	381.67	27.88
	Standard deviation	4.96	1.62	.67	102.15	2.65
Oct. 2000	Average	15.67	22.88	1.6	267.2	19.85
	Median	16.53	22.98	1.35	262.75	20.68
	Minimum	6.36	20.27	.64	139.01	11.65
	Maximum	25.87	26.91	5.88	337.33	26.6
	Standard deviation	4.62	1.81	1.07	49.88	3.35
Nov. 2000	Average	13.48	18.83	2.1	145.85	15.09
	Median	11.59	18.82	1.57	182.41	14.02
	Minimum	5.06	13.69	.94	-6.55	5.73
	Maximum	23.35	24	4.53	252.44	23.97
	Standard deviation	5.54	3.82	1.12	87.95	5.66
Dec. 2000	Average	8.81	12.48	2.28	126.2	8.93
	Median	7.18	12.81	2.09	154.65	7.7
	Minimum	3.34	8.61	.99	-12.56	-.7
	Maximum	20.78	16.61	4.41	218.92	20.36
	Standard deviation	4.9	2.17	.96	77.36	5.32

Table A2—continued. Monthly statistics for average climate conditions on Lake Seminole.

[mbar, millibar; °C, degree Celsius; m/s, meter per second; cal/cm²/day, calories per square centimeter per day]

Date		Vapor press (mbars)	Average surface temperature (°C)	Average wind (m/s)	Regression net radiation (cal/cm²/day)	Average air temperature (°C)
Jan. 2001	Average	8.37	10.25	2.12	152.63	8.92
	Median	7.94	10.84	1.78	167.88	7.88
	Minimum	0	7.19	1.03	.67	.63
	Maximum	18.91	13.8	4.52	243.16	18.67
	Standard deviation	4.2	2.05	.95	67.42	4.96
Feb. 2001	Average	13.03	16.33	2.09	173.39	15.29
	Median	13.25	16.61	1.75	173.81	14.82
	Minimum	5.41	12.63	.88	9.26	8.25
	Maximum	20.16	20.61	5.04	292.31	21.38
	Standard deviation	4.51	2.34	.98	89.19	3.89
Mar. 2001	Average	11.61	17.73	2.33	211.76	14.57
	Median	11.15	17.17	2.28	255.53	14.38
	Minimum	4.62	15.19	1.24	7.2	8.23
	Maximum	22.66	21.38	4.07	369.81	22.1
	Standard deviation	4.51	1.67	.71	130.53	3.87
Apr. 2001	Average	16.28	23.49	2.13	344.03	20.37
	Median	16.24	24.16	1.96	354.6	20.84
	Minimum	6.59	17.76	1.15	107.56	11.98
	Maximum	25.31	26.92	4.07	431.33	24.86
	Standard deviation	4.85	2.67	.69	76.84	3.07
May 2001	Average	18.57	27.09	1.96	376.11	23.66
	Median	17.61	27.31	1.82	394.04	23.9
	Minimum	11.74	24.92	1.25	210.85	20.84
	Maximum	26.29	28.73	3.37	474.13	27.04
	Standard deviation	3.77	.96	.5	69.17	1.44
June 2001	Average	25.07	29.6	1.84	319.89	25.71
	Median	25.4	29.6	1.84	386.14	25.65
	Minimum	18.17	28.09	1.09	36.76	23.51
	Maximum	27.96	31.11	3.02	459.17	27.46
	Standard deviation	2.26	.76	.45	124.4	1.1
July 2001	Average	27.15	30.91	1.8	330.99	27.22
	Median	27.96	31.18	1.76	341.28	27.34
	Minimum	19.97	29.13	1.14	64.77	25.17
	Maximum	29.6	32.31	3.6	435.12	29.46
	Standard deviation	2.23	.8	.42	82.07	1.04
Aug. 2001	Average	27.01	30.22	1.76	290.68	26.58
	Median	27.32	30.46	1.61	307.43	26.43
	Minimum	20.81	28.36	.98	41.74	24.37
	Maximum	30.99	31.75	5.01	422.4	28.6
	Standard deviation	2.14	.94	.77	90.26	1.06
Sept. 2001	Average	21.98	28.17	1.93	266.75	23.81
	Median	24.21	28.34	1.64	278.47	25.08
	Minimum	11.54	23.7	1.04	30.57	16.83
	Maximum	28.03	30.6	3.82	345.58	27.43
	Standard deviation	5.64	1.99	.69	69.15	3.08

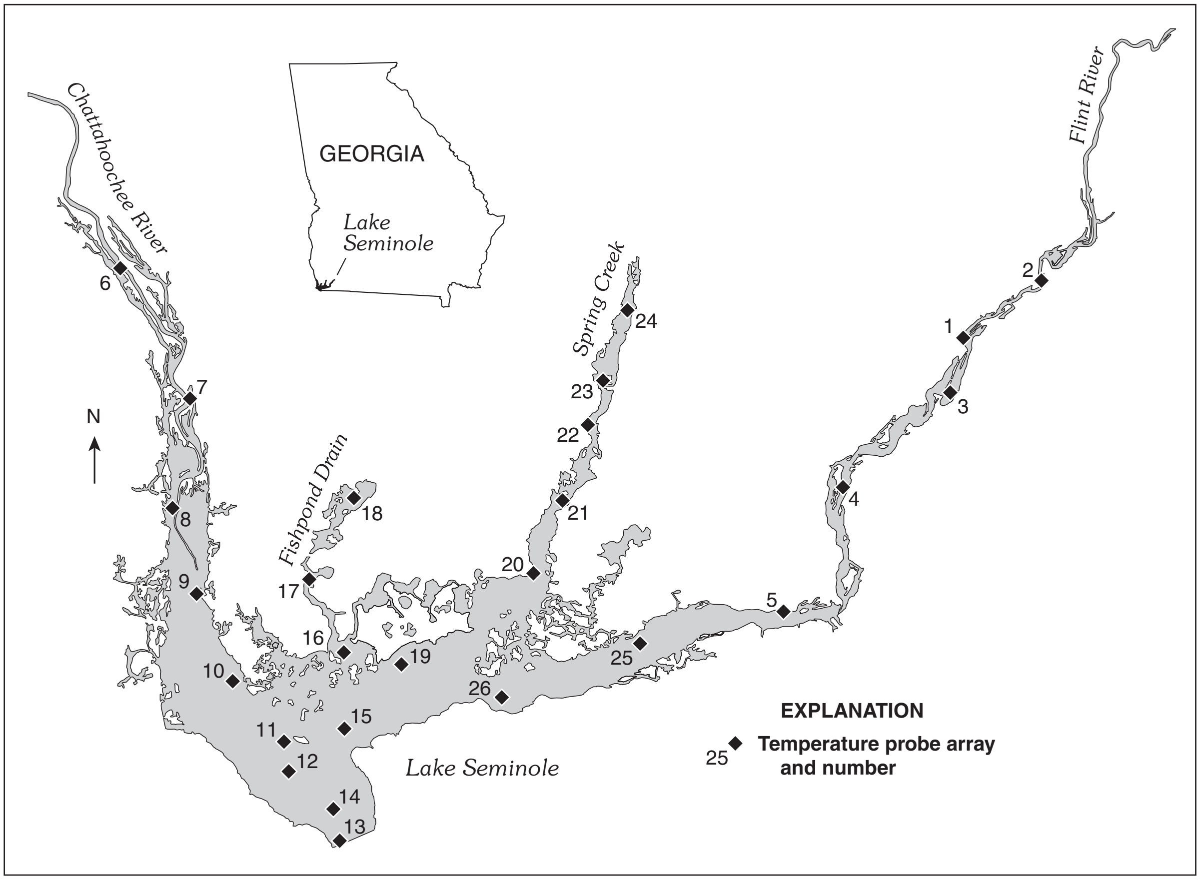

Figure A2. Temperature probe arrays in Lake Seminole, Georgia.

Because of drought, lake stage during the study period often was lower than the stage that existed when the probes were installed. As a result, the top probe on many of the arrays was out of the water, measuring air temperature during periods of low lake stage. In order to prevent the use of air temperature-affected probes, those probes near (0.2 feet or less below) or above the water surface were excluded from analyses. Data also were excluded from malfunctioning probes and temperature spikes that occurred while servicing the probes.

Lake-Temperature Monitoring

Temporal and spatial patterns in lake-water temperature were evaluated to ensure accurate volumetric representation of lake temperature with point-temperature data. These patterns in lake-water temperature determined the appropriate scheme for categorizing and spatially averaging point-temperature data to ensure that estimates for changes in heat stored by the lake can be calculated accurately. Each of the impoundment arms and the main lake body have differing hydrologic conditions that determine their temperature regimes. Therefore, instead of treating lake-water temperature as being areally uniform, the relative temperature and volume in each part of the lake were represented in heat-exchange calculations.

Temporal and spatial patterns in lake-water temperature were assessed using time-series and seasonal-temperature-profile plots. Time-series plots generated for each temperature array illustrate diurnal and seasonal patterns in temperature profiles (fig. A3). Plots of seasonal-temperature profiles are useful for illustrating changes in spatial distribution of temperature in the lake through time (fig. A4; table A4).

Patterns of lake-water temperature profiles in the five distinct lake areas—Flint River impoundment arm (arrays 1–5), Chattahoochee River impoundment arm (arrays 6–8), Fishpond Drain (arrays 17–18), Spring Creek impoundment arm (arrays 20–24), and the main body of the lake (arrays 9–16, 19, and 25–26)—make it necessary to weight temperature data by their representative water volume. For example, the Spring Creek impoundment arm generally is colder than the other regions (fig. A4); and, therefore, the heat budget should account for heat exchange in the corresponding volume of water. Similarly, if the lake is well stratified, then temperature data may be weighted by its representative water volume at different depths.

Table A3. Temperature probe arrays, probe depth, installation data and probe elevation.

[ft, feet; do., ditto]

Probe	Array	Probe height above bottom (ft)	Bottom depth at installation (ft)	Installation date	Probe elevation (ft)
1-0	1	0	16.0	Mar. 3, 2000	60.87
1-7	do.	7	do.	do.	67.87
1-12	do.	12	do.	do.	72.87
2-0	2	0	19.3	do.	57.57
2-5	do.	5	do.	do.	62.57
2-10	do.	10	do.	do.	67.57
2-16	do.	16	do.	do.	73.57
3-0	3	0	10.5	do.	66.37
3-4	do.	4	do.	do.	70.37
3-8	do.	8	do.	do.	74.37
4-0	4	0	13.7	do.	63.17
4-7	do.	7	do.	do.	70.17
4-11	do.	11	do.	do.	74.17
5-0	5	0	9.3	do.	67.57
5-4	do.	4	do.	do.	71.57
5-7	do.	7	do.	do.	74.57
6-0	6	0	20	Mar. 8, 2000	57.05
6-4	do.	4	do.	do.	61.05
6-10	do.	10	do.	do.	67.05
6-14	do.	14	do.	do.	71.05
6-18	do.	18	do.	do.	75.05
6-0	6	0	19.5	Apr. 23, 2001	57.86
6-4	do.	4	do.	do.	61.86
6-8	do.	8	do.	do.	65.86
6-12	do.	12	do.	do.	69.86
6-17	do.	17	do.	do.	74.86
7-0	7	0	12	do.	65.05
7-6	do.	6	do.	do.	71.05
7-10	do.	10	do.	do.	75.05
8-0	8	0	20	do.	57.05
8-4	do.	4	do.	do.	61.05
8-10	do.	10	do.	do.	67.05
8-14	do.	14	do.	do.	71.05
8-18	do.	18	do.	do.	75.05
8-0	8	0	14	Sept. 14, 2000	61.97
8-4	do.	4	do.	do.	65.97
8-8	do.	8	do.	do.	69.97
8-11	do.	11	do.	do.	72.97
8-14	do.	14	do.	do.	75.97

Table A3—continued. Temperature probe arrays, probe depth, installation data and probe elevation.

[ft, feet; do., ditto]

Probe	Array	Probe height above bottom (ft)	Bottom depth at installation (ft)	Installation date	Probe elevation (ft)
9-0	9	0	7	Mar. 8, 2000	70.05
9-3	do.	3	do.	do.	73.05
9-5	do.	5	do.	do.	75.05
10-0	10	0	7	do.	70.05
10-3	do.	3	do.	do.	73.05
10-5	do.	5	do.	do.	75.05
11-0	11	0	10	do.	67.05
11-5	do.	5	do.	do.	72.05
11-8	do.	8	do.	do.	75.05
12-0	12	0	13	do.	64.05
12-4	do.	4	do.	do.	68.05
12-8	do.	8	do.	do.	72.05
12-11	do.	11	do.	do.	75.05
13-0	13	0	12	do.	65.05
13-4	do.	4	do.	do.	69.05
13-7	do.	7	do.	do.	72.05
13-10	do.	10	do.	do.	75.05
14-0	14	0	15	Mar. 9, 2000	62.14
14-4	do.	4	do.	do.	66.14
14-8	do.	8	do.	do.	70.14
14-11	do.	11	do.	do.	73.14
14-14	do.	14	do.	do.	76.14
15-0	15	0	16.5	do.	60.64
15-4	do.	4	do.	do.	64.64
15-8	do.	8	do.	do.	68.64
15-12	do.	12	do.	do.	72.64
15-15	do.	15	do.	do.	75.64
16-0	16	0	8.75	do.	68.39
16-4	do.	4	do.	do.	72.39
16-7	do.	7	do.	do.	75.39
17-0	17	0	8	do.	69.14
17-3	do.	3	do.	do.	72.14
17-6	do.	6	do.	do.	75.14

Table A3—continued. Temperature probe arrays, probe depth, installation data and probe elevation.
[ft, feet; do., ditto]

Probes	Array	Probe height above bottom (ft)	Bottom depth at installation (ft)	Installation date	Probe elevation (ft)
18-0	18	0	6.2	Mar. 9, 2000	70.94
18-3	do.	3	do.	do.	73.94
18-5	do.	5	do.	do.	75.94
19-0	19	0	11	do.	66.14
19-3	do.	3	do.	do.	69.14
19-6	do.	6	do.	do.	72.14
19-9	do.	9	do.	do.	75.14
20-0	20	0	15.8	do.	61.34
20-3	do.	3	do.	do.	64.34
20-7	do.	7	do.	do.	68.34
20-11	do.	11	do.	do.	72.34
20-14	do.	14	do.	do.	75.34
21-0	21	0	17.8	do.	59.34
21-3	do.	3	do.	do.	62.34
21-8	do.	8	do.	do.	67.34
21-13	do.	13	do.	do.	72.34
21-16	do.	16	do.	do.	75.34
22-0	22	0	11.5	do.	65.64
22-4	do.	4	do.	do.	69.64
22-8	do.	8	do.	do.	73.64
22-10	do.	10	do.	do.	75.64
23-0	23	0	8	do.	69.14
23-4	do.	4	do.	do.	73.14
23-6	do.	6	do.	do.	75.14
24-0	24	0	9.5	do.	67.64
24-3	do.	3	do.	do.	70.64
24-6	do.	6	do.	do.	73.64
24-8	do.	8	do.	do.	75.64
25-0	25	0	21.5	do.	55.64
25-6	do.	6	do.	do.	61.64
25-12	do.	12	do.	do.	67.64
25-17	do.	17	do.	do.	72.64
25-20	do.	20	do.	do.	75.64
26-0	26	0	18.85	do.	58.29
26-7	do.	7	do.	do.	65.29
26-14	do.	14	do.	do.	72.29
26-17	do.	17	do.	do.	75.29

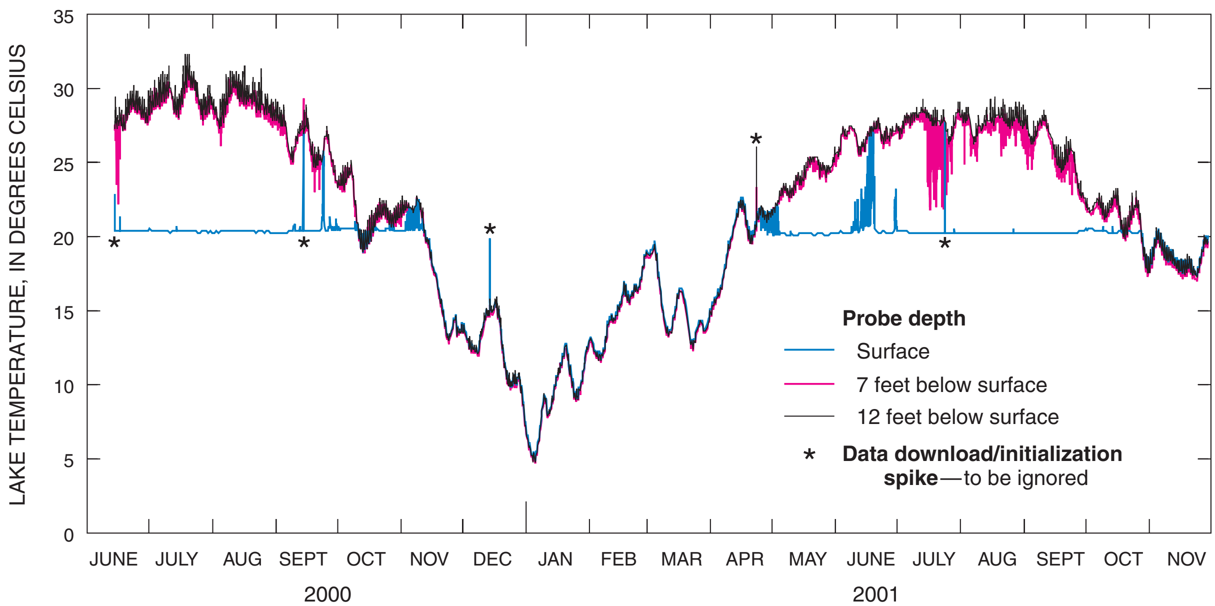

Figure A3. Time-series plot for temperature probe array 1 (*see* figure A2 for location of array 1).

A high density of probes was installed in areas of the lake where temperatures were expected to be more variable, such as areas where there were multiple sources of water at different temperatures. For example, the Spring Creek impoundment arm has a large ground-water component to mean annual baseflow, almost 62 percent (Mosner, 2002), which affects lake-water temperature (fig. A4). The Spring Creek impoundment arm represents less than 5 percent of the total lake volume but contains almost one-fifth of the temperature arrays (arrays 20–24; fig. A2).

Temperature arrays also were categorized by water depth, either as shallow (less than 10 feet deep, 11 arrays) or deep (deeper than 10 feet, 15 arrays), to identify temperature patterns associated with water depth. Patterns may be attributed to factors such as light reflection or absorption by the lake bottom and differences in absorption and reflectivity due to the presence of submerged vegetation. Note that because temperature arrays commonly were located based on the convenience of preexisting attachment points, the water-depth categories do not necessarily represent the depth characteristics of the different areas of the lake.

Temporal Variation in Lake-Water Temperature

Time-series plots identify seasonal and diurnal patterns in the variation and stratification of lake-water temperature (fig. A3). In general, patterns of water temperature are similar to fluctuations of average air temperature (fig. 12), high during summer and low during winter. The highest water temperature recorded during the study was about 30°C during July and August of 2000 and 2001; the lowest water temperature recorded was about 5°C during early January 2001. Because there was evidence of stratification in each part of the lake in the summer and winter, temperature was evaluated temporally during these periods.

Winter Lake-Temperature Patterns

During the winter period, November through March, lake water generally was well mixed; however, in deep parts of the lake, where there is less influence by large inflows, there was evidence of stratification. During the winter, two temperature arrays (23 and 25) did not have enough probes functioning to assess patterns in temperature responses. Of the remaining 24 temperature arrays, 13 arrays indicated that waters were well mixed. The arrays representing well-mixed lake water included all five arrays in the Flint River and Chattahoochee River impoundment arms, upstream arrays in the Spring Creek impoundment arm (22 and 24), and arrays in the main body of the lake near the Chattahoochee River impoundment arm (arrays 9–11) (fig. A2). The areal extent of mixing suggests that surface-water inflow into the lake keeps these areas well mixed during the winter period.

Eleven temperature arrays exhibited some degree of thermal stratification during the winter period, with colder, denser water located at the bottom of the water column and warmer, less-dense water at the top (fig. A2). These arrays are located in the main body of the lake (arrays 12–16 and 26) and in the Spring Creek impoundment arm (arrays 20 and 21) (fig. A2). These arrays were not influenced greatly by large inflows and showed a diurnal pattern in temperature associated with air temperature. The magnitude of diurnal variations was highest for probes located close to the surface of the lake as opposed to probes located deeper (fig. A2).

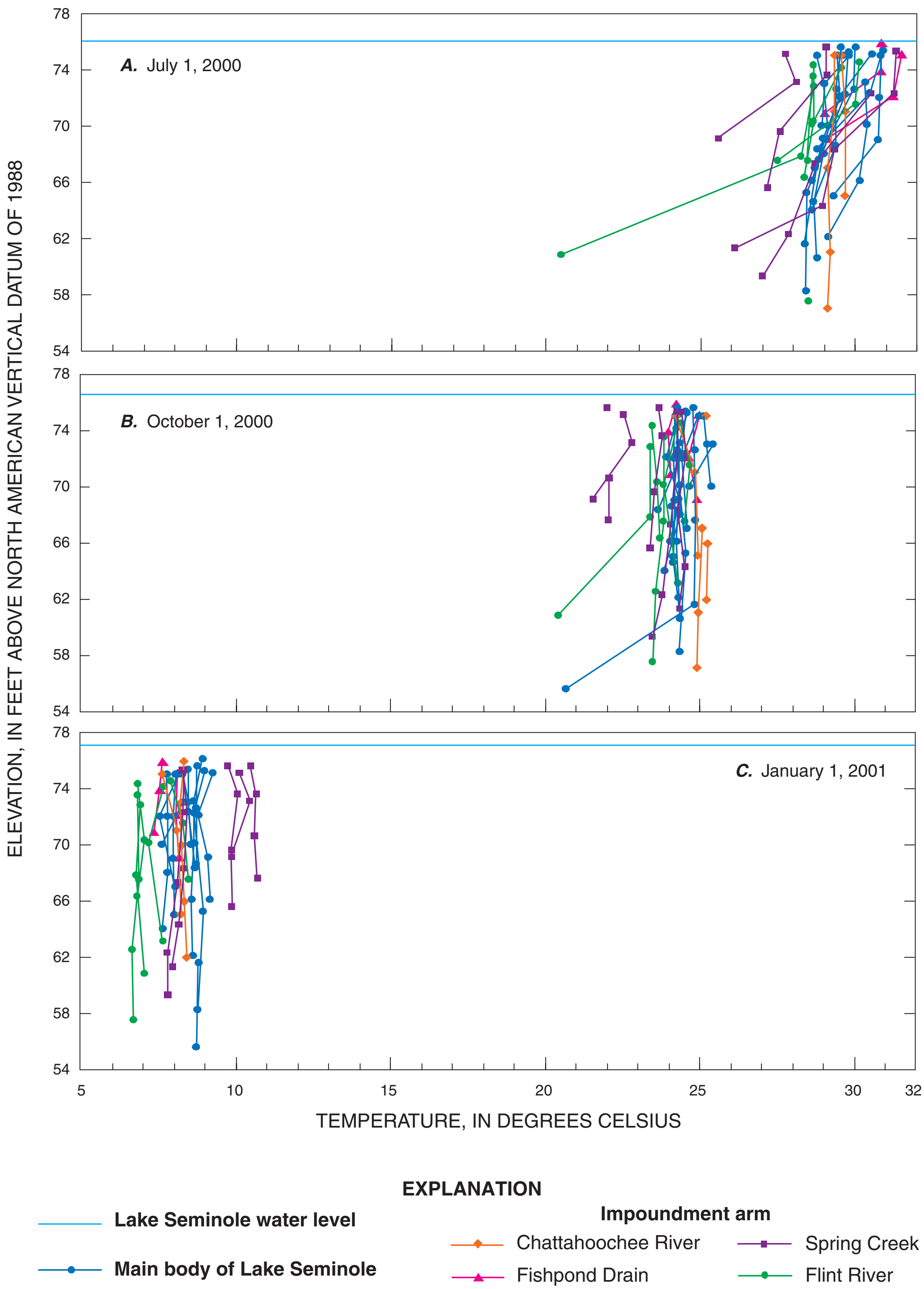

Figure A4. Plots of seasonal-temperature profiles illustrating spatial temperature differences seasonally on (*A*) July 1, 2000, (*B*) October 1, 2000, and (*C*) January 1, 2001.

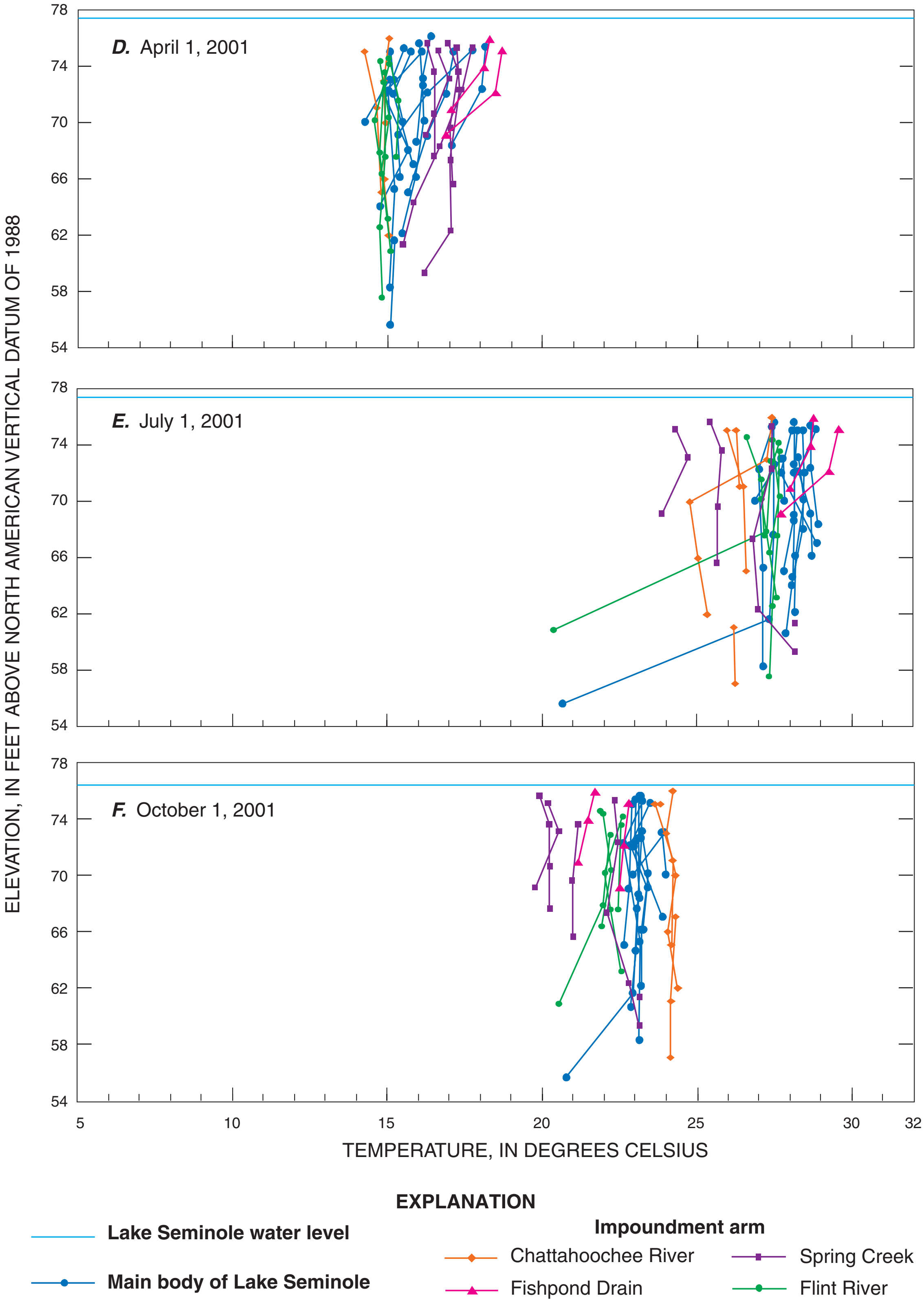

Figure A4—continued. Plots of seasonal-temperature profiles illustrating spatial temperature differences seasonally on (*D*) April 1, 2001, (*E*) July 1, 2001, and (*F*) October 1, 2001.

Table A4. Mean, median, minimum, and maximum temperatures for each area of Lake Seminole during the study period, April 2000–September 2001 (*see* figure A2).

[CRIA, Chattahoochee River impoundment arm; FPD, Fishpond Drain; FRIA, Flint River impoundment arm; MBL, main body of the lake; SCIA, Spring Creek impoundment arm, all temperatures in degrees Celsius]

Probes	Date	Count	Mean	Median	Minimum	Maximum
All	July 1, 2000	84	28.99	29.09	20.48	31.51
CRIA	do.	8	29.37	29.33	29.09	29.71
FPD	do.	6	30.42	30.85	29.01	31.51
FRIA	do.	14	28.16	28.61	20.48	30.13
MBL	do.	40	29.21	29.32	20.67	30.89
SCIA	do.	16	28.45	28.4	25.56	31.32
All	Oct. 1, 2000	92	24.11	24.28	20.42	25.42
CRIA	do.	9	24.95	24.94	24.18	25.25
FPD	do.	6	24.46	24.43	23.99	24.98
FRIA	do.	16	23.66	23.76	20.42	24.67
MBL	do.	41	24.35	24.34	20.67	25.42
SCIA	do.	20	23.47	23.73	21.54	24.56
All	Jan. 1, 2001	88	8.33	8.27	6.63	10.69
CRIA	do.	8	8.17	8.24	7.61	8.39
FPD	do.	6	7.81	7.85	7.35	8.16
FRIA	do.	16	7.21	6.96	6.63	8.46
MBL	do.	37	8.42	8.55	7.52	9.25
SCIA	do.	21	9.21	9.72	7.78	10.69
All	Apr. 1, 2001	88	15.96	15.72	14.26	18.71
CRIA	do.	8	14.82	14.9	14.26	15.07
FPD	do.	6	17.93	18.21	16.9	18.71
FRIA	do.	16	14.95	14.93	14.59	15.35
MBL	do.	37	15.86	15.67	14.27	18.17
SCIA	do.	21	16.76	16.96	15.5	17.74
All	July 1, 2001	84	27.33	27.59	20.38	29.59
CRIA	do.	9	26.54	26.4	25.96	27.42
FPD	do.	6	28.68	28.73	27.72	29.59
FRIA	do.	16	26.92	27.36	20.38	27.68
MBL	do.	40	27.85	28.14	20.67	28.91
SCIA	do.	13	26.18	25.8	23.87	28.17
All	Oct. 1, 2001	85	22.65	22.9	19.77	24.37
CRIA	do.	12	24.11	24.15	23.63	24.37
FPD	do.	6	22.05	22.11	21.16	22.78
FRIA	do.	13	22.08	22.2	20.52	22.6
MBL	do.	38	23.08	23.1	20.77	23.98
SCIA	do.	16	21.25	20.98	19.77	23.12

Summer Lake-Temperature Patterns

From late spring and summer into early fall, April through October, a variety of patterns were identified in the water-temperature time-series data. Most of the shallow probes and many of the deep probes showed a diurnal variation with air temperature. The magnitude of the diurnal fluctuation was high for the shallow probes and low for relatively deep probes.

In the Flint River impoundment arm, the most upstream arrays that were not affected by springflow into the lake bottom (arrays 2 and 3) indicate that waters were well mixed. Diurnal variations were more pronounced in the shallow probes than in relatively deep probes. The influence of inflowing cold ground water along the lake bottom was evident in the thermal profiles for the remaining three arrays (1, 4–5) in the Flint River impoundment arm.

Water temperature along the Chattahoochee River impoundment arm (arrays 6–8) and in the main body of the lake close to the Chattahoochee River impoundment arm (arrays 9–11) remained unstratified during the summer period, which was similar to the winter; the exception being for the uppermost upstream array, which showed some stratification during the summer.

Summer thermal profiles of lake water located in Fishpond Drain (arrays 17–18) and in the main body of the lake near Fishpond Drain (array 16) indicated water temperatures were stratified, as during the winter. Water temperatures at the lake bottom for these arrays showed little or no diurnal patterns. During warming periods (April through August), this water gains heat slowly, lagging behind the temperature response of the water above, suggesting that during the summer, some of the water at depth is from ground-water discharge. Another possible explanation for thermal stratification in Fishpond Drain is the very small, almost negligible, streamflow that enters the lake from this impoundment arm. The small streamflow, coupled with shallow water depth and incidence of submerged vegetation, causes water in Fishpond Drain to move slowly, allowing thermal stratification to occur due to radiational heating.

Along the Spring Creek impoundment arm, lake-water temperature profiles indicate increased stratification with distance downstream toward the main body of the lake. The lower-depth probes in arrays located near the main body of the lake indicate little or no diurnal-temperature variation. These observations suggest that the water was well mixed in the upstream reaches of the impoundment arm. Ground-water inflow to Spring Creek along the length of the impoundment arm adds relatively cooler water (about 20°C) to the bottom of the water column, resulting in increased downstream stratification during the summer period.

Probe 1-0, located along the Flint River impoundment arm, and 25-0, located in the main body of the lake near the Flint River impoundment arm, were installed where springflow enters the lake bottom; summer temperature patterns at these probes are dominated by ground water. When lake temperatures are warmer than ground water, usually from May through mid-November, these probes record nearly constant water temperature, varying between 20 and 21°C, indicating ground-water inflow (fig. A3). These temperatures are slightly warmer than the average ground-water temperature in the region (about 68°F, or 20°C). Occasionally, after a heavy rainfall event, lake water becomes well mixed temporarily with ground water from these springs, resulting in an unstratified thermal profile. At other times, when the temperature of the lake water is slightly warmer than the incoming springflow, some mixing of water not associated with heavy rainfall events occurs. Near the end of the summer period, usually from late October to early November, lake water cools to temperatures that are below that of the inflowing ground water, and the thermal profiles indicate no stratification until the following spring, although temperature data collected at the spring orifices indicate that the springs are still flowing. Temperatures at the remaining arrays located in the main body of the lake (arrays 12–15, 19, 25–26) are stratified during the summer.

Spatial Variation in Lake-Water Temperature

Seasonal differences in water temperature for the five lake areas are illustrated in quarterly plots of temperature and temperature-probe altitude for January and April 2001, and for July and October 2000 and 2001 (fig. A4). Water temperatures recorded at midnight were plotted to reduce the effects associated with diurnal temperature variations.

There were no distinguishing patterns in temperature profiles between shallow- and deep-water temperature arrays (not shown), indicating that water depth is not an important factor to establishing vertical patterns of lake-temperature variation. The lack of any distinguishable pattern in lake temperature in relation to depth also suggests that the presence of submerged vegetation, which is widespread in the shallow waters covering about half of the lake area, does not affect water-temperature profiles appreciably. There were, however, distinct seasonal patterns in temperature profiles among the different lake areas (fig. A4).

On July 1, 2000, most lake-water temperature measurements were similar, ranging between 28 and 31°C (fig. A4*A*). The deep water in the Spring Creek impoundment arm and in the shallow Fishpond Drain was colder than the lake-water temperature in most of the other lake areas. The low temperature along the Spring Creek impoundment arm indicates the influence of ground-water inflow to the lake. Two probes (1-0 and 25-0), identified earlier as being located at or near springs that discharge ground water into the lake along the bottom, recorded the coldest temperatures of all the probes (20.5 and 20.7°C, respectively).

On October 1, 2000, there were small differences in lake temperature among the different areas of the lake; temperatures along the Flint River impoundment arm were slightly colder, and temperatures along the Chattahoochee River impoundment arm were slightly warmer than other areas of the lake (fig. A4*B*). Colder temperatures were apparent along the Spring Creek impoundment arm for the two uppermost upstream arrays (23 and 24) at all depths, due to ground-water inflow. The coldest lake temperatures again were recorded at probes 1-0 and 25-0, which monitor springflow temperature along the lake bottom, with values similar to those measured on July 1, 2000.

On January 1, 2001, lake temperatures were the lowest of any time period evaluated, with temperatures ranging from 6.5 to 10.5°C (fig. A4*C*). Lake temperatures at the four uppermost upstream arrays in the Flint River impoundment arm were colder than temperatures recorded at nearly all of the other probes. The three uppermost upstream arrays in the Spring Creek impoundment arm recorded the warmest lake temperatures of all the other arrays due to ground-water inflow.

On April 1, 2001, most lake temperatures ranged from 14.5 to 17.5°C (fig. A4*D*). Lake temperatures along the Flint River and Chattahoochee River impoundment arms were colder than the main body of the lake; whereas, lake temperatures along the Spring Creek impoundment arm and Fishpond Drain were warmer than the main body of the lake.

On July 1, 2001, there was a more distinct separation in temperature of the different areas compared with July 1, 2000 (compare figs. A5*A*, *E*). With the exception of two probes that monitor springflow along the lake bottom, temperatures ranged from about 24 to 29°C. The coldest lake area was the Spring Creek impoundment arm, followed by the Chattahoochee River and Flint River impoundment arms, and the main body of the lake; the warmest area was Fishpond Drain.

The most distinct separation in temperatures among the impoundment arms occurred on October 1, 2001 (fig. A4*F*); temperatures ranged from 19.5 to 24.5°C. The areas of the lake from coldest to warmest were the Spring Creek impoundment arm, Flint River impoundment arm, main body of the lake, and Chattahoochee River impoundment arm. Temperatures in Fishpond Drain were somewhat similar to those in the Flint River impoundment arm, except with a larger range.

Plots of temperature profiles for the six dates categorized by depth of array (shallow or deep) indicated that there was no pattern in water temperatures associated with depth of water. However, because of the pronounced differences in temperature patterns and representative volumes of each area of the lake (fig. A4), it was necessary to weight the temperatures