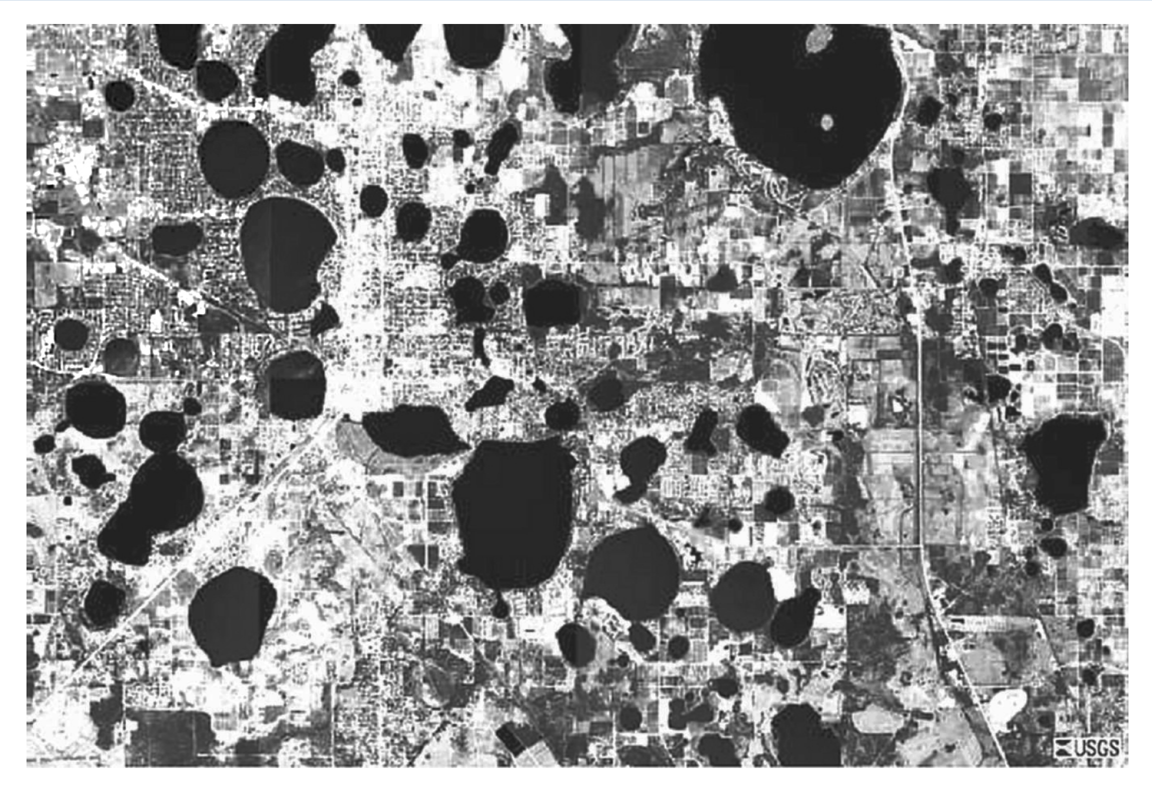
Estimating Ground-Water Inflow to Lakes in Central Florida Using the Isotope Mass-Balance Approach

Water-Resources Investigations Report 02-4192



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

Prepared in cooperation with the
Southwest Florida Water Management District

Cover: Aerial view of lakes in Polk County, near Winter Haven, Florida; from digital orthophoto quadrangle, photographed January 6, 1999, by the National Aerial Photography Program.

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By Laura A. Sacks

U. S. GEOLOGICAL SURVEY
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SOUTHWEST FLORIDA WATER MANAGEMENT DISTRICT

Tallahassee, Florida
2002

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Multiply	By	To obtain
inch (in.)	2.54	centimeter
foot (ft)	0.3048	meter
mile (mi)	1.609	kilometer
acre	4,047	square meter
acre	0.4047	hectare
foot squared (ft^2)	0.09290	square meter
cubic foot (ft^3)	0.02832	cubic meter
cubic feet per second (ft^3/s)	0.02832	cubic meter per second

Temperature in degrees Celsius (ºC) can be converted to degrees Fahrenheit (ºF) as follows:

$$^{o}F = (1.8 \times {}^{o}C) + 32$$

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

Vienna Standard Mean Ocean Water (VSMOW) is the reference for both deuterium and oxygen-18 isotope ratio scales.

Acronyms and Additional Abbreviations used in Report

α	alpha
$\delta^{18}O$	delta oxygen-18
δ_a	isotopic composition of atmospheric moisture
δD	delta deuterium
δ_E	isotopic composition of evaporating water
δ_{Gi}	isotopic composition of ground-water inflow
δ_L	isotopic composition of lake water
δ_{L1}	initial isotopic composition of lake water
δ_{L2}	final isotopic composition of lake water
δ_P	isotopic composition of precipitation
$^{\circ}C$	degrees Celsius
E	evaporation
G_i	ground-water inflow
GIS	geographic information system
h	relative humidity, normalized to lake-surface temperature
in/yr	inches per year
L/min	liters per minute
mL	milliliters
MWL	meteoric water line
p	probability level
P	precipitation
r^2	coefficient of determination
RH	relative humidity
S_i	surface-water inflow
SWFWMD	Southwest Florida Water Management District
T_a	air temperature
T_o	lake-surface temperature
USGS	U.S. Geological Survey
V_1	initial lake volume
V_2	final lake volume
VWM	volume-weighted mean

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Abstract

The isotope mass-balance approach was used to estimate ground-water inflow to 81 lakes in the central highlands and coastal lowlands of central Florida. The study area is characterized by a subtropical climate and numerous lakes in a mantled karst terrain. Ground-water inflow was computed using both steady-state and transient formulations of the isotope mass-balance equation. More detailed data were collected from two study lakes, including climatic, hydrologic, and isotopic (hydrogen and oxygen isotope ratio) data. For one of these lakes (Lake Starr), ground-water inflow was independently computed from a water-budget study. Climatic and isotopic data collected from the two lakes were similar even though they were in different physiographic settings about 60 miles apart. Isotopic data from all of the study lakes plotted on an evaporation trend line, which had a very similar slope to the theoretical slope computed for Lake Starr. These similarities suggest that data collected from the detailed study lakes can be extrapolated to the rest of the study area.

Ground-water inflow computed using the isotope mass-balance approach ranged from 0 to more than 260 inches per year (or 0 to more than 80 percent of total inflows). Steady-state and transient estimates of ground-water inflow were very similar. Computed ground-water inflow was most sensitive to uncertainty in variables used to calculate the isotopic composition of lake evaporate (isotopic compositions of lake water and atmospheric moisture and climatic variables). Transient results were particularly sensitive to changes in the isotopic composition of lake water. Uncertainty in ground-water inflow results is considerably less for lakes with higher ground-water inflow than for lakes with lower ground-water inflow. Because of these uncertainties, the isotope mass-balance approach is better used to distinguish whether ground-water inflow quantities fall within certain ranges of values, rather than for precise quantification.

The lakes fit into three categories based on their range of ground-water inflow: low (less than 25 percent of total inflows), medium (25-50 percent of inflows), and high (greater than 50 percent of inflows). The majority of lakes in the coastal lowlands had low ground-water inflow, whereas the majority of lakes in the central highlands had medium to high ground-water inflow.

Multiple linear regression models were used to predict ground-water inflow to lakes. These models help identify basin characteristics that are important in controlling ground-water inflow to Florida lakes. Significant explanatory variables include: ratio of basin area to lake surface area, depth to the Upper Floridan aquifer, maximum lake depth, and fraction of wetlands in the basin. Models were improved when lake water-quality data (nitrate, sodium, and iron concentrations) were included, illustrating the link

between ground-water geochemistry and lake chemistry. Regression models that considered lakes within specific geographic areas were generally poorer than models for the entire study area. Regression results illustrate how more simplified models based on basin and lake characteristics can be used to estimate ground-water inflow.

Although the uncertainty in the amount of ground-water inflow to individual lakes is high, the isotope mass-balance approach was useful in comparing the range of ground-water inflow for numerous Florida lakes. Results were also helpful in understanding differences in the geographic distribution of ground-water inflow between the coastal lowlands and central highlands. In order to use the isotope mass-balance approach to estimate inflow for multiple lakes, it is essential that all the lakes are sampled during the same time period and that detailed isotopic, hydrologic, and climatic data are collected over this same period of time. Isotopic data for Florida lakes can change over time, both seasonally and interannually, primarily because of differences in net precipitation. The isotope mass-balance approach was most successful for lakes in the central highlands, where lakes have higher ground-water inflow, are deeper, and undergo less isotopic variability, compared to lakes in the coastal lowlands. Results from this study illustrate the large range in ground-water inflow to Florida lakes and underscore the importance of ground water in the water budget of many of Florida's lakes.

INTRODUCTION

Ground-water inflow can be an important part of the water balance of Florida lakes, but inflow can vary substantially between lakes (Grubbs, 1995; Lee, 1996; Lee and Swancar, 1997; Sacks and others, 1998). Differences in the amount of ground-water inflow between lakes can affect lake stage and water quality. A lake with high ground-water inflow may experience less stage decline during a drought than a similar lake with low ground-water inflow. Loading of solutes from the adjacent ground-water catchment may also be higher for a lake with high ground-water inflow, and this can adversely impact lake water quality. For example, if nitrate concentrations were high in the shallow ground water around two lakes, the lake with more ground-water inflow would have higher nitrate loading than the lake with less ground-water inflow.

Ground-water inflow is a difficult term to quantify (Winter, 1995). Detailed basin-scale studies have been used to understand ground-water exchange with several Florida lakes (Grubbs, 1995; Lee, 1996; Swancar and others, 2000; Swancar and Lee, in press). For these studies, ground-water inflow and outflow were quantified using detailed water budgets and ground-water flow models that were calibrated to data from numerous wells in the lake basin. Although these studies provide valuable insight into processes controlling ground-water exchange, they are expensive and their results cannot be readily extrapolated to the larger population of Florida's more than 7,800 lakes. In addition, computing ground-water inflow from saturated ground-water flow modeling has limitations that can cause ground-water inflow to be underestimated when recharge is particularly high (Lee, 1996, 2000; Swancar and Lee, in press). Another approach used to understand ground-water fluxes is to compute net ground-water flow (ground-water inflow minus outflow) as the residual to the water budget. This method is helpful in understanding short-term and seasonal changes in ground-water flow, as well as understanding differences among lakes (Sacks and others, 1998; Swancar and others, 2000; Metz and Sacks, 2002); however, because only net flow (difference between inflow and outflow) is computed, this method cannot be used to precisely quantify both ground-water inflow and outflow. Other methods used to estimate ground-water inflow are also problematic. Flow-net analysis relies on assumptions, such as depth of the flow field and the area contributing flow to the lake, which are not easy to quantify (Belanger and Kirkner, 1994; Lee and Swancar, 1997), and seepage meters rely on point measurements, which must be extrapolated spatially and temporally (Fellows and Brezonik, 1980; Belanger and Montgomery, 1992).

Chemical and isotope mass-balance approaches have been used to estimate ground-water inflow to lakes (Stauffer, 1985; Krabbenhoft and others, 1994). Chemical mass-balance approaches, using a conservative solute such as chloride or sodium, has been successfully used for lakes in undeveloped areas of Florida (Pollman and others, 1991; Sacks and others, 1998). However, land-use practices can greatly alter ground-water chemistry in developed areas, and large uncertainties in ground-water chemistry can cause large uncertainties in computed ground-water inflow (Sacks and others, 1998). This approach also has problems when the concentration of the conservative tracer is similar in both the lake and ground water.

The isotope mass-balance approach (using hydrogen and oxygen isotope ratios) has rarely been used to estimate ground-water inflow to Florida lakes (Sacks and others, 1998). However, it has been used successfully in other areas of the country and world (for example, Dinçer, 1968; Zimmermann, 1979; Krabbenhoft and others, 1990; LaBaugh and others, 1997; Yehdegho and others, 1997). The greatest limitation to using this approach is in quantifying the isotopic composition of water evaporating from the lake. The isotope mass-balance approach is well-suited for estimating ground-water inflow to lakes in geographic proximity (Dinçer, 1968; Krabbenhoft and others, 1994) and has been applied this way in the upper midwest of the United States (Ackerman, 1992; Michaels, 1995). In addition, Sacks and others (1998) indicated that this approach was promising for Florida lakes. The isotope mass-balance approach for multiple lakes is most successful when coupled with detailed data collection at a lake with a known water budget (Dinçer, 1968; Krabbenhoft and others, 1994).

The Southwest Florida Water Management District (SWFWMD) has been legislatively mandated to establish minimum water levels that are acceptable for lakes in its district. These levels need to be established regardless of availability of long-term stage data at individual lakes. To understand differences in lake-level responses between lakes, it is important to understand how ground-water exchange differs between lakes. Thus, a simplified approach was needed to estimate the ground-water component in the water budget of numerous lakes. In October 1998, the U.S. Geological Survey (USGS) began a 4-year cooperatively funded project with the SWFMWD to use the isotope mass-balance approach to estimate ground-water inflow to numerous lakes in central Florida.

Purpose and Scope

The purpose of this report is to estimate ground-water inflow to 81 lakes in central Florida using the isotope mass-balance approach. Specific objectives are: (1) to evaluate the use of the isotope mass-balance approach for estimating ground-water inflow to Florida lakes; (2) to better understand the distribution of ground-water inflow over a broad population of Florida lakes; (3) to gain additional knowledge on the isotope budget of Florida lakes, including the isotopic composition of rainfall and atmospheric moisture; and (4) to evaluate factors that are important in controlling the amount of ground-water inflow to Florida lakes.

Lakes were sampled in the coastal lowlands and central highlands of western and central peninsular Florida, primarily in Highlands, Polk, Hillsborough, and Pasco Counties (fig. 1). Detailed climatic and isotopic data were collected from one lowland and one highland lake. At the highland lake (Lake Starr), ground-water inflow and lake evaporation were independently estimated for a detailed water-budget study (Swancar and others, 2000; Swancar and Lee, in press). At the lowland lake (Halfmoon Lake), climatic and isotopic data were compared to those collected at Lake Starr. The remaining lakes had considerably less data collection. Data were collected primarily during 1999 and 2000. Steady-state and transient formulations of the isotope mass-balance equation were used to estimate ground-water inflow, and a sensitivity analysis was used to evaluate the effects of uncertainty in terms of the equation on computed ground-water inflow. Multiple linear regression was used to generate predictive models of ground-water inflow using lake-specific basin characteristics.

Lake Selection

The 81 lakes selected for this study cover a wide range of lake depths, sizes, and basin topographies (fig. 1). Lakes with little or no surface-water drainage (seepage lakes) were selected over surface-drainage lakes to eliminate the effect of uncertainty in surface-water flows on calculated ground-water inflow. Another selection criteria was that there be at least 8 years of recent stage data (typically monthly measurements) between 1992 and 2000, so that results could be linked with a concurrent study evaluating lake-stage variability. Several lakes that had less stage data were included because additional data were available (for example, an earlier estimate of ground-water inflow). All lakes were within the geographic boundaries of the SWFWMD. Lakes were also favored for selection if they were on SWFWMD's priority list to establish management levels, or if they were used as reference lakes in prior studies (Southwest Florida Water Management District, 1999). Upland lakes in Polk and Highlands

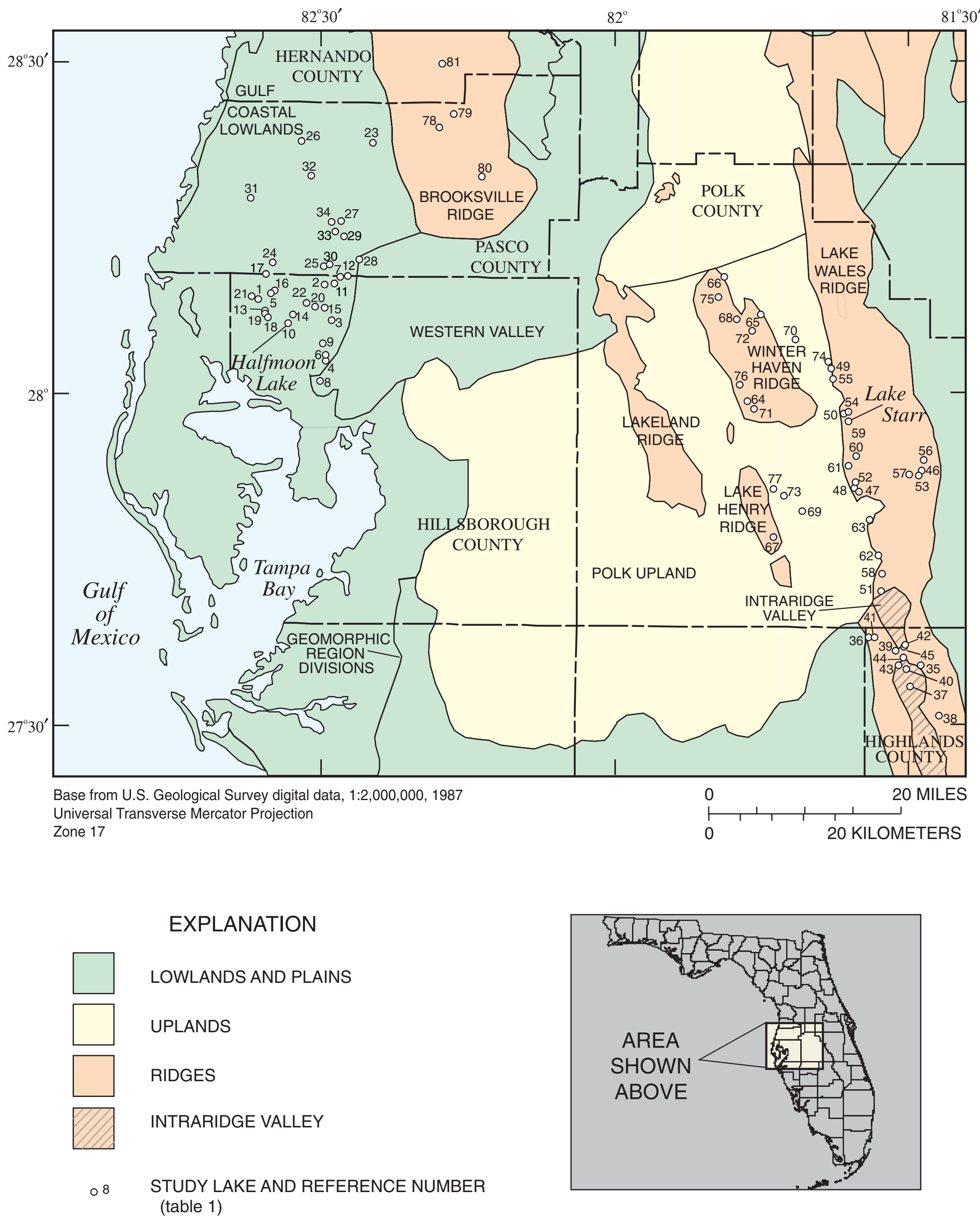


Figure 1. Location of study lakes and relation to geomorphic regions (after White, 1970).

Counties were restricted geographically to central Polk and northern Highlands Counties, because results from Sacks and others (1998) suggested that there may be some geographic variations in the isotopic composition of atmospheric moisture between these areas and southern Highlands County and western Polk County (Lakeland Ridge area).

Of the 81 study lakes, 47 were in ridge and upland areas of the central highlands, and the remaining 34 were in the coastal lowlands north of Tampa Bay. Lakes ranged in surface area from 3 to 5,074 acres, and in maximum depth from 2 to 78 ft (table 1). Lakes in the central highlands tended to be larger and deeper than lakes in the coastal lowlands. The median lake surface area was 64 acres for highland lakes, compared to 32 acres for lowland lakes; the median maximum lake depth was 27 ft for highland lakes, compared to 17 ft for lowland lakes.

Table 1. Location and size of study lakes

Map reference number (figure 1)	Lake name	Section	Township (South)	Range (East)	Maximum lake depth (feet)	Lake surface area (acres)
Hillsborough County lakes in coastal lowlands:						
1	Alice	16	27	17	24	92
2	Allen	10	27	18	25	24
3	Bird	26	27	18	14	25
4	Boat	14	28	18	17	30
5	Calm	14	27	17	25	116
6	Carroll	15	28	18	18	201
7	Deer	1	27	18	31	32
8	Egypt	27	28	18	32	60
9	George	10	28	18	24	26
10	Halfmoon	31	27	18	25	31
11	Hobbs	1	27	18	18	60
12	Hog Island	6	27	19	19	49
13	Juanita	22	27	17	20	23
14	LeClare	30	27	18	16	42
15	Merrywater	22	27	18	8	21
16	Mound	11	27	17	24	75
17	Osceola	3	27	17	17	58
18	Raleigh	27	27	17	17	23
19	Rogers	27	27	17	16	92
20	Starvation	21	27	18	11	27
21	Taylor	16	27	17	20	45
22	Van Dyke	17	27	18	4	13
Pasco County lakes in coastal lowlands:						
23	Big Fish	28	24	19	8	65
24	Black	26	26	17	11	5
25	Camp	34	26	18	7	16
26	Crews	16	24	18	2	267
27	Curve	1	26	18	16	28
28	Gooseneck	29	26	19	13	25
29	King	7	26	19	9	118
30	Linda	26	26	18	11	22
31	Moon	28	25	17	21	112
32	Pierce	9	25	18	15	32
33	Thomas (Pasco)	11	26	18	11	151
34	Wistaria	2	26	18	14	23
Highlands County lakes in central highlands:						
35	Angelo	25	33	28	15	43
36	Chilton	7	33	28	21	21
37	Denton	2	34	28	48	57
38	Dinner (Highlands)	17	34	29	31	365
39	Isis	15	33	28	63	45
40	Lotela	26	33	28	27	778
41	Olivia	6	33	28	46	82
42	Pioneer	11	33	28	35	72
Highlands County lakes in central highlands:--continued						
43	Tulane	27	33	28	78	77
44	Verona	23	33	28	68	41
45	Viola	14	33	28	33	59
Polk County lakes in central highlands (Lake Wales Ridge and Intraridge Valley):						
46	Aurora	13	30	28	37	108
47	Blue (South Lobe)	24	30	27	58	79
48	Blue (North Lobe)	24	30	27	51	17
49	Crystal	21	28	27	20	105
50	Dinner (Polk)	15	29	27	13	15
51	Hickory	17	32	28	13	99
52	Josephine	13	30	27	28	10
53	Little Aurora	13	30	28	34	15
54	Mabel	11	29	27	17	83
55	Menzie	28	28	27	21	18
56	Saddlebag	6	30	29	46	164
57	Saint Anne	14	30	28	33	15
58	Silver	5	32	28	28	129
59	Starr	14	29	27	31	109
60	Wales	1	30	27	19	273
61	Warren	11	30	27	20	3
Polk County lakes in central highlands (other ridge and upland areas):						
62	Clinch	31	31	28	29	1,194
63	Crooked	1	31	27	25	5,074
64	Eagle	1	29	25	27	401
65	Grassy	19	27	26	14	94
66	Helene	34	26	25	27	52
67	Henry	16	31	26	21	64
68	Little Van	26	27	25	9	19
69	Lizzie	1	31	26	23	73
70	Lucerne	2	28	26	16	39
71	McLeod	7	29	26	27	337
72	Medora	36	27	25	23	48
73	Polecat	27	30	26	25	21
74	Sara	17	28	27	15	34
75	Tennessee	9	27	25	17	110
76	Thomas (Polk)	35	28	25	25	62
77	Walker	21	30	26	31	23
Pasco County lakes in central highlands:						
78	Iola	15	24	20	41	99
79	Jessamine	11	24	20	26	33
80	Pasadena	16	25	21	12	259
Hernando County lake in central highlands:						
81	Spring	15	23	20	39	47

Lake Starr and Halfmoon Lake were chosen for additional data collection because they are good examples of highland and lowland lakes, respectively, and because they were the recent focus of USGS basin-scale studies to define ground-water exchange (Swancar and others, 2000; Metz and Sacks, 2002; Swancar and Lee, in press). In addition, Lake Starr has been designated by the USGS and SWFWMD as a "benchmark lake" for long-term hydrologic data collection. These data are intended to assess the response of the lake's water budget to changes in climatic conditions and other hydrologic stresses. Thus, Lake Starr has the additional advantage of ongoing data collection to independently quantify ground-water exchange and lake evaporation.

Description of Study Area

Lakes are a dominant feature in the landscape of the study area (fig. 2). The lakes are situated in a mantled karst terrain, where surficial sands and clays overlie (or mantle) a karstified limestone surface. Most lakes in the study area are thought to be of sinkhole origin (Sinclair and others, 1985; Evans others, 1994; Tihansky and others, 1996), formed when cavities in the underlying limestone cause subsidence in the overlying sands and clays. Lakes are situated in depressions in the surficial deposits and are surficial expressions of the water table in the unconfined surficial aquifer system. The surficial aquifer system is underlain by the intermediate confining unit, which is typically disrupted beneath a lake as the result of

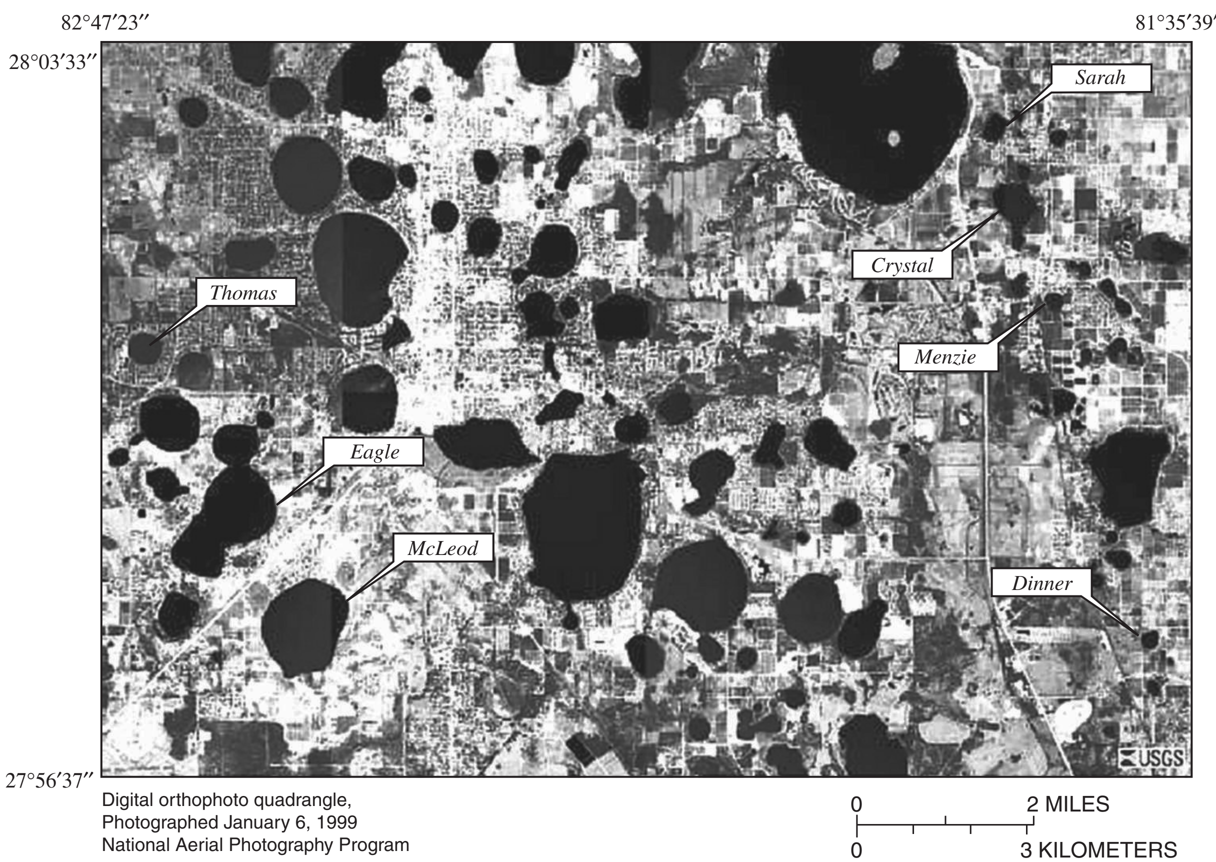


Figure 2. Aerial view of lakes in Polk County near Winter Haven, Florida.

subsidence that formed the lake depression. In the southern part of the study area, the intermediate confining unit may have one or more permeable zones, called the intermediate aquifer system (Southeastern Geological Society, 1986). Beneath the intermediate confining unit/aquifer system is the highly productive, carbonate Upper Floridan aquifer.

The study lakes are in a recharge setting where water levels in lakes and the surficial aquifer system are higher than heads in the Upper Floridan aquifer. Thus, head gradients and flow directions are downward between aquifers. Because of this gradient, vertical ground-water outflow (leakage) occurs through deep parts of the lake bed, particularly where lake sediments are thin (fig. 3). In shallow parts of the lake bed, both ground-water inflow and lateral ground-water outflow can occur, depending on the local setting of the lake. Surficial deposits and the intermediate confining unit are much thinner beneath study lakes in the coastal lowlands (typically less than 50 ft) than in the central highlands, where these deposits can be as great as 400 ft thick beneath lakes in the southern part of the study area (Buono and Rutledge, 1978).

Lakes that were sampled are in several geomorphic divisions (White, 1970) (fig. 1). Lakes termed "lowland lakes" are primarily in the Gulf Coastal Lowlands. Topographic relief in this area is typically low, as the land was submerged below sea level during high sea-level stands in the Pleistocene. Lakes termed "highland lakes" are in ridge and upland areas described by White (1970) as the Central Highlands. Highland lakes are in the Winter Haven, Lake Henry, Lake Wales, and Brooksville Ridges, the Intraridge Valley, and the Polk Upland. (Where a lake was bounded by more than one ridge or upland area, it was placed within the area that bounded the majority of the lake surface.) Ridges are generally parallel with the Atlantic coast and were not submerged during high sea-level stands in the Pleistocene age. Thus, ridge lakes have the potential to be older than lowland lakes (White, 1970; Watts, 1975; Grimm and others, 1993). Griffith and others (1997) defined lake regions based on water quality, soils, physiography, and surficial geology. Lowland lakes are in the Keystone Lakes, Land-o-Lakes, Tampa Plain, and Weeki Wachee Hills lake regions. Highland lakes are primarily in the Northern and Southern Lake Wales Ridge, Winter Haven/Lake Henry Ridges, and Southern Brooksville Ridge lake regions.

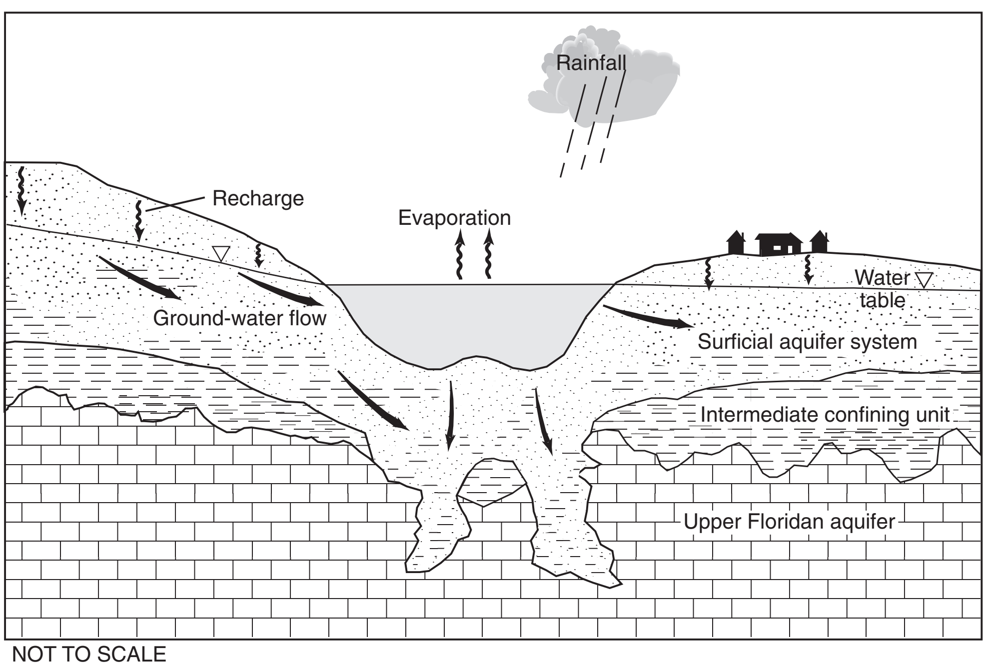


Figure 3. Generalized hydrogeologic section through a central Florida lake.

The climate in the study area is humid subtropical. Average annual rainfall is about 50 in/yr, with about 60 percent of the rain occurring during the summer wet season (June through September). Rainfall from winter frontal activity generally is not as intense as summer rainfall. Annual average lake evaporation is estimated to range between 48 and 57 in/yr (Farnsworth and others, 1982; Lee and Swancar, 1997; Swancar and others, 2000). Evaporation rates for shallow lakes (less than 30 ft) are highest between April and August. Peak evaporation rates from deep lakes can occur later in the fall because of greater heat storage compared to shallow lakes (Sacks and others, 1994). Because of the lack of reliable evaporation data throughout the study area, the spatial distribution of evaporation rates is not known; however, lake evaporation may vary with latitude and proximity to the coast.

Acknowledgments

The author thanks the many lake-front property owners who allowed access to the various study lakes during sampling. Thanks also to Erik Staub (USGS) for help with data collection and compilation, and to Craig Hutchinson and Mark Estes, for help with additional data collection at Halfmoon Lake and Lake Starr, respectively. Thanks to Terrie Lee, Amy Swancar, and David Krabbenhoft of the USGS, and to Curtis Pollman (Tetratek, Inc.) for technical discussions. Tyler Coplen (USGS) was instrumental in the timely analysis of the isotope samples, particularly during sampling design for atmospheric moisture collection. Thanks to Larry Benson (USGS) for discussions on atmospheric moisture sampling methods and loan of equipment. Assistance with project planning, compiling of lake data, and lake sampling were provided by numerous individuals from SWFWMD, including Mike Hancock, Richard Gant, Don Ellison, Roberta Starks, Ken Romie, and Keith Kolasa. Water-quality data were provided by SWFWMD and Polk County, and bathymetric data were obtained from SWFWMD, University of Florida's LAKEWATCH (Department of Fisheries and Aquatic Sciences), and University of South Florida's Hillsborough County Lake Atlas (Florida Center for Community Design and Research). This report was improved by helpful review comments from Tyler Coplen, Edward German, Terrie Lee, and Amy Swancar of the USGS, and from Brian Armstrong and Mike Hancock of SWFWMD.

METHODS

The isotope mass-balance approach has been used to understand the water budget of lakes in various locations around the world (Dinçer, 1968; Gat, 1970; Zuber, 1983; Krabbenhoft and others, 1990; LaBaugh and others, 1997; Yehdegho and others, 1997). This section discusses the theory of the isotope mass-balance approach, sampling methods used in this study, definition of basin characteristics, and the statistical approach used in this report.

Theory

Deuterium and oxygen-18 are naturally occurring stable isotopes of the water molecule. These isotopes are excellent conservative tracers because they are part of the water molecule itself, rather than dissolved constituents that may undergo reactions and dispersion. Deuterium (^{2}H or D) is a heavy isotope of hydrogen (^{1}H is the most abundant form), and oxygen-18 (^{18}O) is a heavy isotope of oxygen (^{16}O is the most abundant form). Only about 0.015 percent of hydrogen on earth is in the form of ^{2}H, and 0.204 percent of oxygen is in the form of ^{18}O (Clark and Fritz, 1997). Because of their low concentrations, these isotopes are not measured directly. Instead, the ratio of the heavy to light isotope is measured (R_{sample}) and reported relative to the ratio in a reference (R_{ref}) in δ notation:

$$\delta_{sample} = 1000\left(\frac{R_{sample}}{R_{ref}} - 1\right), \tag{1}$$

where R is ^{2}H/^{1}H for hydrogen and ^{18}O/^{16}O for oxygen. Results are reported in units of per mil (parts per thousand). Vienna Standard Mean Ocean Water (VSMOW) is the reference for both deuterium and oxygen-18 isotope ratio scales.

The relative amount of D and ^{18}O in the environment varies depending on the water phase and location (for example, elevation, latitude, and distance from the ocean-continent boundary). D and ^{18}O preferentially condense out of the water vapor, so that rainfall is isotopically enriched in D and ^{18}O relative to the water vapor. ^{1}H and ^{16}O (the lighter isotopes) have higher vapor pressures and diffusivities, causing them to preferentially evaporate, compared to D and ^{18}O. As a result, surface water tends to be more enriched in D and ^{18}O (the heavier isotopes) compared to water vapor (or atmospheric moisture). This global balance between evaporation and condensation results in a consistent relation between δD and $\delta^{18}O$ in rainfall around the world (the global meteoric water line, fig. 4; Craig, 1961; Rozanski and others, 1993). Surface water influenced by evaporation is offset to the right of the meteoric water line because of differences in how the two isotopes fractionate during evaporation (fig. 4). A number of excellent references are available that discuss D and ^{18}O in the water cycle (for example, International Atomic Energy Agency, 1981; Gonfiantini, 1986; Clark and Fritz, 1997; and Coplen and others, 1999).

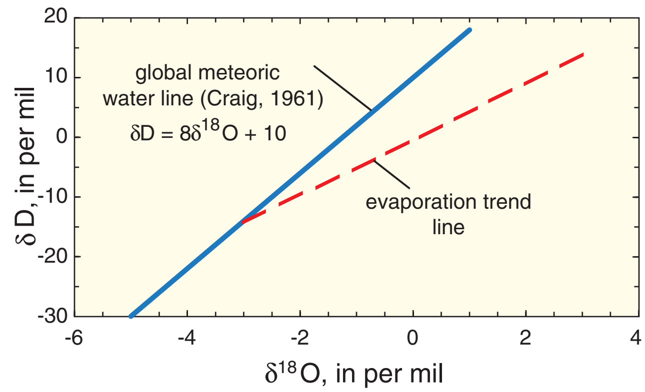


Figure 4. General relation between δD and $\delta^{18}O$ for global meteoric water and water that has undergone evaporation.

Differences in the isotopic composition of lakes in the same geographic and climatic region can give a general indication of lake-water residence time, which reflects differences in water fluxes through the lakes. For example, two seepage lakes of similar volume can have different rates of ground-water inflow and outflow (fig. 5). For the lake with the least amount of ground-water inflow, the longer hydraulic residence time allows more time for evaporation and evaporative enrichment of D and ^{18}O in the lakes, resulting in a lake enriched in D and ^{18}O. Conversely, the lake with more ground-water inflow has a shorter hydraulic residence time and less evaporative enrichment of D and ^{18}O, leading to a lake depleted in D and ^{18}O.

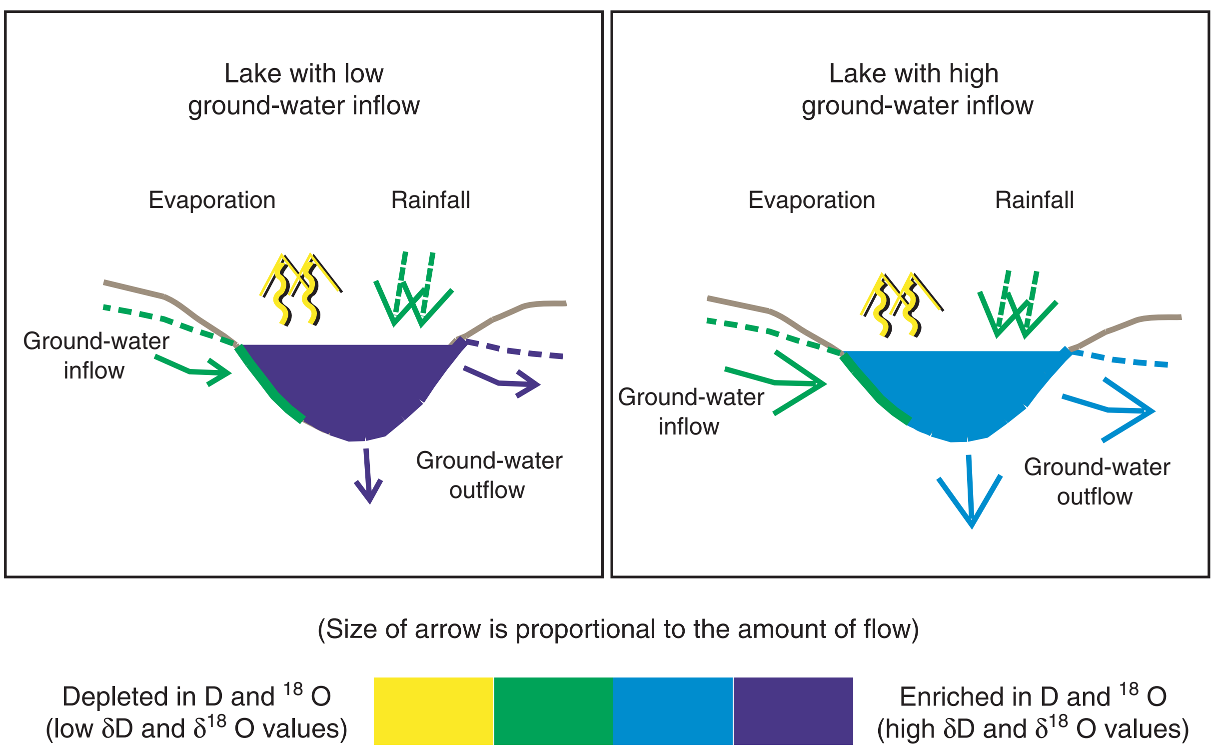


Figure 5. Schematic illustrating differences in isotopic composition between two lakes with different amounts of ground-water inflow.

Isotope Mass Balance

The isotope mass-balance approach combines the general water-budget equation with the isotope mass-balance equation. The lake water-budget equation is:

$$\Delta V = P - E + S_i - S_o + G_i - G_o , \qquad (2)$$

where ΔV is change in lake volume over the specified time period, P is precipitation (rainfall), E is evaporation, S_i is surface-water or stormwater inflow, S_o is surface-water outflow, G_i is ground-water inflow, and G_o is ground-water outflow. The isotope mass-balance equation for δD or $\delta^{18}O$ is:

$$\Delta(V\delta_L) = P\delta_p - E\delta_E + S_i\delta_i - S_o\delta_L + G_i\delta_{Gi} - G_o\delta_L , \qquad (3)$$

where δ is either the δD or $\delta^{18}O$ composition of the lake (δ_L) or the water sources or sinks defined in equation 2. The isotopic composition of surface-water outflow and ground-water outflow are assumed to equal that of the lake. Either δD or $\delta^{18}O$ values can be used in the isotope mass-balance approach to provide independent estimates of ground-water inflow. Hydrogen and oxygen isotope ratios are conservative tracers in the hydrologic cycle (at temperatures observed on the earth's surface) and are not involved in any reactions in the lake. The water-budget (eq. 2) can be rearranged to solve for G_o and substituted into equation 3. This equation can then be rearranged to solve for G_i, resulting in the following equation:

$$G_i = \frac{P\delta_P - E\delta_E + S_i\delta_{S_i} - (P - E + S_i - \Delta V)\delta_L - \Delta(V\delta_L)}{\left(\delta_L - \delta_{G_i}\right)} . \qquad (4)$$

Both G_o and S_o cancel out of equation 4 and, thus, do not need to be known in order to use the isotope mass-balance approach. Most of the study lakes are seepage lakes, and so surface-water inflow also cancels out of equation 4 for those lakes.

Equation 4 simplifies further if the lake is assumed to be in both isotopic and hydrologic steady state. Although these conditions clearly do not exist on a short-term basis, the stage and likely the lake isotopic composition cycle around long-term steady-state conditions. Steady-state assumptions have the advantage of providing results that are more computationally robust than a transient analysis, for which small uncertainties can cause large changes in ground-water inflow results (Pollman and Lee, 1993). The longer the lake residence time, the more reasonable steady-state assumptions will be. Steady-state assumptions are best suited for deep lakes that undergo minimal seasonal changes in lake isotopic composition (Krabbenhoft and others, 1994). Both steady-state and transient analyses are included in this report. For the steady-state analysis, ΔV and $\Delta(V\delta_L)$ are assumed to equal 0 in equation 4.

Linear units were used for this study (volume/lake surface area) to allow for direct comparisons between lakes of different size. However, in doing so, there is a slight bias toward higher ground-water inflow for small lakes. For example, for perfectly circular lakes, a smaller lake has a larger ratio of lake perimeter to surface area than a larger lake. Thus, a small lake has the potential to have more lateral ground-water inflow relative to its surface area than a large lake (Millar, 1971; Fellows and Brezonik, 1980).

Isotopic Composition of Evaporating Lake Water

The greatest limitation to using δD and $\delta^{18}O$ values as conservative tracers is defining the isotopic composition of evaporating lake water (δ_E) (Zimmerman and Ehhalt, 1970; Krabbenhoft and others, 1990). According to theory originally developed by Craig and Gordon (1965), evaporation occurs as both equilibrium and kinetic (nonequilibrium) fractionation processes (Gonfiantini, 1986; Clark and Fritz, 1997). Water vapor in a very thin "boundary layer" directly overlying the water surface, where relative humidity is 100 percent, is in isotopic equilibrium with the lake water. Above this is a transition zone where molecular diffusion (which is a kinetic fractionation process) transports water vapor from the boundary layer to the well-mixed open air column. Lighter isotopes of water vapor are transported faster than heavier isotopes. Because of differences in diffusion rates, kinetic fractionation is more significant for $\delta^{18}O$ than for δD.

Following nomenclature from Krabbenhoft and others (1990) and originally derived by Craig and Gordon (1965), the isotopic composition of evaporating water (δ_E) can be defined as:

$$\delta_E = \frac{\alpha^* \delta_L - h\delta_a - \varepsilon}{(1 - h + 0.001\Delta\varepsilon)}\ , \qquad (5)$$

where δ is the δD or $\delta^{18}O$ composition of evaporating water (E), lake water (L), and atmospheric moisture (a); α^* is the equilibrium isotope fractionation factor at the temperature of the air-water interface (equivalent to $1/\alpha$ defined by Majoube, 1971; calculated from the average lake-surface temperature, T_o); h is relative humidity normalized to average lake-surface temperature (vapor pressure of the air divided by the saturation vapor pressure at the lake-surface temperature), expressed as a fraction; ε is the total fractionation factor, expressed in per mil, and is equal to 1,000 $(1 - \alpha^*) + \Delta\varepsilon$; and $\Delta\varepsilon$ is the kinetic fractionation factor, expressed in per mil (estimated as 12.5 (1 - h) for δD and 14.2 (1 - h) for $\delta^{18}O$; Gonfiantini, 1986). Variables α^*, h, ε, and $\Delta\varepsilon$ are all functions of climatic variables of air temperature, lake-surface temperature, and relative humidity. When using the isotope mass-balance approach for numerous lakes in geographic proximity, it can be assumed that climatic variables are the same within the geographic region, and so δ_E can be calculated as a function of δ_L and δ_a.

To obtain a more accurate estimate of annual average δ_E, values of δ_E can be computed for shorter timeframes (for example, monthly) and then volume-weighted based on the evaporation rate (Krabbenhoft and others, 1990; Yehdegho and others, 1997). However, because Florida has very small seasonal climatic variability compared to more temperate regions, this approach may not be necessary. Sacks and others (1998) concluded that the volume-weighted mean δ_E in central Florida was similar to that computed using annual average climatic variables. For $\delta^{18}O$, there was an average of 3 percent difference between δ_E computed from monthly values and from annual mean values; for δD, the average difference was 14 percent, which is probably related to poorer analytical precision for δD (2 per mil) compared to that of $\delta^{18}O$ (0.2 per mil).

Data Collection

Data collection consisted of climatic, hydrologic, and isotopic data. Lake Starr and Halfmoon Lake were the focus of more detailed data collection; considerably less data were collected from the other study lakes.

Climatic and Hydrologic Data

Climatic data, rainfall, and lake stage were measured continuously at Lake Starr during the study period (January 1999 to July 2000). Climatic data were collected from land and raft climate stations to compute energy-budget evaporation beginning in August 1996 (Swancar and others, 2000). For the period January to July 1996, lake evaporation was estimated from monthly averages for 1997 and 1998. Lake-surface temperature, air temperature, and relative humidity also were used to compute δ_E (eq. 5). Rainfall and lake stage were measured hourly. Net ground-water flow, ground-water inflow, and ground-water outflow were computed monthly and annually between January 1996 and March 2001 (Sacks and others; 1998; Swancar and others, 2000; USGS, unpublished data). A three-dimensional ground-water flow model was used to estimate ground-water outflow at Lake Starr for the 2-year period between August 1996 and July 1998 (Swancar and Lee, in press). Before and after the modeled period, ground-water outflow was estimated based on a good linear relation between simulated monthly ground-water outflow and the monthly average head difference between the lake and a well in the Upper Floridan aquifer (well ROMP 57; ground-water outflow (in/month) = -0.20 head difference (ft) -2.22); Swancar and Lee, in press). Ground-water inflow was then computed as the residual to the water budget. Errors in water-budget terms were also assessed to understand uncertainty in ground-water inflow and outflow (Swancar and others, 2000).

Data collected at Halfmoon Lake was more limited than at Lake Starr. Lake-surface temperature, air temperature, and relative humidity were measured near the shore of the lake. Rainfall and lake stage were measured approximately weekly. Toward the end of the study (April to July 2000), ground water was pumped into the lake to raise the lake stage (lake augmentation), and during this period the pumped volume was measured. Halfmoon Lake was the focus of a recent water-budget study, and net ground-water flow was computed for a 3-year period between

June 1996 and May 1999 (Metz and Sacks, 2002). A three-dimensional ground-water flow model was used to estimate ground-water inflow and outflow for this period, but the final model results were not available at the writing of this report.

Considerably less data were collected from the other study lakes. Typically, lake stage was measured when the lake was sampled (if not, the stage reading closest in time to the lake sampling was used). Rainfall was estimated from a network of rain gages operated by the National Weather Service (NWS), SWFWMD, and USGS. Rainfall was averaged annually for periods of 1 to 10 years to use in the steady-state isotope equation, and was totaled between sampling dates for the transient formulation of the isotope equation. Energy-budget evaporation rates from Lake Starr were used to estimate lake evaporation between 1996 and 2000. Prior to that, pan evaporation at Lake Alfred in Polk County was used, with pan coefficients based on comparisons between energy-budget evaporation at Lake Starr and pan evaporation at Lake Alfred. For lakes in the coastal lowlands, evaporation rates may be lower than at Lake Starr (Metz and Sacks, 2002); however, no reliable evaporation data exist for lakes in the coastal lowlands. The effect of uncertainties in the evaporation rate on calculated ground-water inflow are discussed later in the report.

Lake stage-volume-area relations were not available for most of the study lakes. When using the transient formulation of equation 4, the initial and final lake volumes need to be defined. For lakes without stage-volume-area relations, volume was estimated based on the relation between maximum lake depth (x) and mean lake depth (lake volume/surface area) (y) for 29 lakes in the study area where lake volume was known ($y = 0.419x + 0.504$; $r^2 = 0.91$; standard error = 1.8 ft). This estimated mean depth (or lake volume in linear units) was used as an estimate of initial lake volume in the transient calculations, and the final volume was assumed to equal the initial volume plus or minus the change in lake stage between samplings, assuming a constant surface area. Change in lake stage between samplings typically was small (less than 1 ft) compared to the whole lake volume, and errors related to assuming a constant lake surface area are assumed to be small. Sensitivity of results to this term is presented later in the report.

Surface-water inflow to three lakes (Crooked Lake, Lake Clinch, and Lake Lotela) was estimated from periodic discharge measurements and discharge at a nearby continuously gaged site. Discharge was measured in three inflow channels to Crooked Lake, three inflow channels to Lake Clinch, and one inflow channel to Lake Lotela, on five separate occasions between August 1999 and March 2000. Flows typically were very low (median 0.5 ft^3/s). These periodic discharge measurements were correlated to measurements at three nearby gaging stations with continuously computed discharge: Tiger, Carter, and Livingston Creeks. Relations between these point measurements and data from the continuously gaged sites were used to estimate surface-water inflow to the lakes.

Overland flow to lakes in the study area is considered to be low because of the permeable nature of the sandy soils. Runoff was deemed important only where a significant amount of storm drainage flowed directly into the lake or large amounts of the basin were paved. Stormwater inflow (runoff) was estimated for several lakes based on drive-by surveys of the lake shore, data from local city or county agencies, and estimates of impervious surfaces contributing stormwater inflow to the lake. Lakes with substantial stormwater inflow are: Boat, Carroll, Egypt, and George Lakes in Hillsborough County; Lake Wales in Polk County; and Lakes Verona and Isis in Highlands County. The ratio of impervious area to lake area was computed, using runoff coefficients between 0.75 and 0.90. This ratio was multiplied by rainfall to estimate stormwater inflow as a fraction of rainfall.

Isotopic Data

Isotopic data were collected from rainwater (δ_P), atmospheric moisture (δ_a), ground-water inflow(δ_{Gi}), surface-water inflow (δ_{Si}), and lake water (δ_L). Unfiltered samples were collected in glass bottles with polyseal caps for isotope analysis. Isotope samples were analyzed by the USGS Isotope Fractionation Laboratory in Reston, Virginia. Hydrogen isotope ratios were determined using a hydrogen equilibration technique (Coplen and others, 1991), and oxygen isotope ratios were determined using a carbon dioxide equilibration technique (Epstein and Mayeda, 1953). The 95 percent confidence interval for analytical uncertainty is 2 per mil for δD and 0.2 per mil for $\delta^{18}O$. To compare differences in the analytical uncertainty of both isotopes, the greater range in values for δD compared to $\delta^{18}O$ needs to be considered. Because of the lower natural abundance of δD, the fractional error in the δD analysis is about twice that for $\delta^{18}O$ analysis (T.B Coplen, U.S. Geological Survey, written commun., 2001).

Rainwater

Rainwater was collected for δD and $\delta^{18}O$ analysis from Lake Starr and Halfmoon Lake between January 1999 and July 2000. The collection gage was a modified storage rain gage, with the funnel emptying directly into a sample bottle. A plastic ball was placed inside the funnel so that water in the bottle was sealed from the atmosphere except during rain events in order to prevent evaporation and isotopic enrichment of the sample (Scholl and others, 1995; Sacks and others, 1998). A mesh screen was used over the funnel to keep out large debris and to contain the plastic ball during high intensity rain events. The sampler was mounted about 5 ft above the ground in an open area. Samples were collected at least weekly and were composited monthly in a glass bottle with a polyseal cap. Annual isotopic composition of rainwater (δ_P) is presented as the volume-weighted mean (VWM; sum of monthly P times δ_P, divided by annual P).

Atmospheric Moisture

Atmospheric moisture was sampled monthly at both Lake Starr and Halfmoon Lake between May 1999 and April 2000. Sampling consisted of pumping ambient air drawn from a height of 6 ft at the lake's edge. The air was passed through super-cooled condensation tubes, similar to methods of Merlivat and Coantic (1975) and Benson and White (1994). Two copper U-tubes were connected in series in a dewar flask containing a dry ice/ethanol slurry, with temperatures less than -70 $^{\circ}$C. Air was pumped at a flow rate of 4 to 5 L/min through the system for 2-3 hours, usually between midmorning and early afternoon. The second U-tube in the series was inspected to ascertain that all of the water vapor was condensed in the first tube. Typically, 9 mL of water was collected in small glass vials with polyseal caps and sent to the laboratory for isotope analysis. Monthly samples at Lake Starr and Halfmoon Lake were collected within one day of each other. Duplicate samples and concurrent samples from different sides of the lakes were also collected. On one occasion, concurrent samples were taken at locations about 20 mi from each lake to evaluate local differences in atmospheric moisture composition. Differences between duplicate samples and concurrent samples typically were very small (within or near analytical uncertainty).

The isotopic composition of atmospheric moisture (δ_a) was also estimated by two additional methods. In the first method, δ_a was back-calculated from the independent water-budget estimate of ground-water inflow at Lake Starr and measured values of other hydraulic, climatic, and isotopic variables in equation 4. This ground-water inflow value was used in equation 4 to calculate δ_E, which was then used in equation 5 and rearranged to solve for δ_a. The other method assumed δ_a was in isotopic equilibrium with rainwater (Turner and others, 1984; Krabbenhoft and others, 1990; Gibson and others, 1999). In this method, δ_a is computed as $\delta_a = \delta_P - \varepsilon$, where $\varepsilon = (\alpha - 1)\ 1{,}000$, and α is defined from equations derived by Majoube (1971) using air temperature.

Ground-Water and Surface-Water Inflow

To estimate the isotopic composition of ground-water inflow in the study area, samples were collected near the shore of 14 lakes twice during the study (October 1999, following the wet season, and March-April 2000, during the dry season). Lakes were selected to geographically encompass both the lowland and highland study areas. Samples were collected using a minipiezometer (or hydraulic potentiomanometer), which consists of a stainless steel, hollow rod with a well point and screen; a manometer board; and tubing to both the rod and the lake (Winter and others, 1988). The rod was pushed into the ground near the lake, either onshore or offshore, and water was pumped alternately from the ground-water and the lake-water sides to the manometer board. The head difference between the lake and the ground water was measured after heads stabilized. Samples were collected for isotope analysis at sites where the ground-water head was higher than the lake, indicating ground-water inflow. Specific conductance of the ground water and lake water also was measured to establish that lake water was not being sampled. For the first sampling, existing near-shore wells were also sampled at Lake Starr and Halfmoon Lake to see if minipiezometer samples represented the shallow ground water. Differences between the minipiezometer and well samples were within analytical error, indicating that the minipiezometer successfully sampled the ground water.

At each lake, one onshore and one offshore sample were typically collected, about 5 to 10 ft from the shoreline. Samples were usually collected from 1.5 to

2 ft below land surface or the lake bottom, and the offshore samples were collected in less than 1 ft of water. In October 1999, 34 ground-water inflow samples were collected from 14 lakes, and in March-April 2000, 22 ground-water inflow samples were collected from 11 lakes. The lower number of samples collected in the March-April 2000 dry season was because ground-water levels were lower than lake levels at several lakes. At these lakes, seasonal flow reversals resulted in ground-water outflow occurring in areas that previously had ground-water inflow.

For the three lakes with surface-water inflows (Crooked Lake, Lake Clinch, and Lake Lotela), samples were collected after discharge measurements were made on two to three occasions, depending on whether there was flow in the channel. Grab samples were collected because inflow channels typically were very small and shallow. The mean isotopic composition of surface-water inflow for each lake was computed by weighting the samples by flow and applying this to the entire study period. Stormwater inflow was assumed to equal that of the VWM isotopic composition of rainwater.

Lake Water

Most lakes were sampled twice: in July or August 1999 (summer 1999) and in January or February 2000 (winter 2000). These sampling periods coincided with periods of high and low lake evaporation, which could represent seasonal extremes in isotopic enrichment or depletion. During the summer of 1999, 79 lakes were sampled, and during the winter of 2000, 2 additional lakes (Lakes Isis and Olivia in Highlands County) were sampled to compare with results from an earlier study (Sacks and others, 1998). About 20 percent of the lakes were sampled by SWFWMD, when the lakes were also sampled for major ions and nutrients. For Crooked Lake, which is the largest of the study lakes, samples were collected from both the larger north lobe and the smaller south lobe (Little Crooked); these lobes were connected during the study. To understand more about year-to-year variability in the isotopic composition of lakes, a subset of nine lakes was sampled for a longer period of time, typically three additional semiannual samples: summer of 1998, winter of 1999, and summer of 2000. Lake Starr and Halfmoon Lake were sampled monthly between January 1999 and July 2000 to gain more understanding of shorter-term variability in lake isotopic composition.

Typically, one representative sample was collected from each lake at a depth of about 1.5 ft below the lake surface. Small lakes (less than about 100 acres) were sampled in approximately the center of the lake, whereas larger lakes were sampled in open water away from the shoreline. Florida lakes are typically well mixed; however, deep Florida lakes (greater than about 30 ft) can thermally stratify in the summer to early fall, and "turn over," or mix, in the winter (Brenner and others, 1990). Before the lakes were sampled, vertical stratification was evaluated by measuring profiles of temperature and specific conductance in the lake. Most lakes had minimal vertical stratification of specific conductance (less than 10 percent change from surface to bottom), except near the lake sediments. For about 13 percent of lake samples, either duplicate samples or samples from different depths or locations in the lake were collected. Differences in δD and $\delta^{18}O$ for these samples were within or very near analytical uncertainty. The only exception is the difference between a shallow (1.5 ft) and a deep (45 ft) sample from Lake Verona from the summer of 1999, which was greater than analytical uncertainty (the deep sample was 3.8 per mil lighter for δD and 0.51 per mil lighter for $\delta^{18}O$). In the winter of 2000, however, the difference between shallow and deep samples at Lake Verona was within analytical uncertainty. Secchi depth was also measured at each lake at the time of sampling as an indicator of lake-water clarity.

Defining Basin Characteristics

Basin characteristics were determined or estimated for each lake from lake-specific data, regional maps, or site-specific geographic information system (GIS) coverages. Compiled data included physical characteristics of the lake and its basin, the lake's hydrogeologic setting, and additional data from the lake and its basin. Geomorphic divisions were defined from White (1970) and Brooks (1981), lake regions were defined from Griffith and others (1997), and surface geology was defined from Scott and others (2001).

Maximum lake depth was taken from bathymetric maps for 52 of the 81 lakes; for 5 additional lakes, fathometer transects or other bathymetric data were available. The lake depth on bathymetric maps was adjusted to reflect stage conditions when the lakes were sampled, as lake stage was typically several feet different at the time of the bathymetric surveys. When bathymetric data were not available, the maximum lake depth observed during lake sampling was used. For those lakes with bathymetric maps, the maximum lake bottom slope and the average lake bottom slope (from the four compass directions) were computed to a standard depth (from the shoreline to a depth of 10 ft for lakes less than 25 ft deep, and to a depth of 20 ft for lakes greater than 25 ft deep).

Lake surface area and perimeter were estimated from GIS soils coverages. Lake basins were defined from topographic maps as the topographic highs (hilltops) around each lake. The surface area and perimeter of each basin were computed, as well as an indicator of basin shape – the ratio of the longest and shortest linear distance from lake shore to the topographic high. Calculated ratios used to illustrate different aspects of lake and basin size included lake surface area/perimeter, basin area/perimeter, basin area/lake surface area, and lake surface area/maximum lake depth. Steepness of land surface in each basin and within 50 m (164 ft) of the lake shore was computed from slope calculations using topographic data from 1:24,000 scale digital map data. Slope computations were made using a grid cell size of 10 m by 10 m (32.8 ft by 32.8 ft) and ARC/INFO GRID software surface modeling functions.

GIS coverages of soils, land use, and wetlands were used to estimate the percentage of soil types and hydrologic classifications, land use, and wetlands for each lake basin and within 100 m (328 ft) of the lake's shore line. These GIS coverages were obtained from SWFWMD. Soils data were digitized from county soils atlases from the Natural Resource Conservation Service (formerly the Soil Conservation Service). Land-use types were photointerpreted from 1:12,000 USGS color infrared digital orthophoto quarter quadrangles taken between the fall of 1995 to the spring of 1996, and interpreted according to the Florida Land Use and Cover Classification System. Wetland delineation was from the U.S. Fish and Wildlife Service National Wetlands Inventory.

Depth and thickness of hydrogeologic units were derived from maps by Buono and Rutledge (1978) and Buono and others (1979). The head in the Upper Floridan aquifer was extrapolated from maps of the potentiometric surface for dry season (May 1999) and wet season (September 1999) conditions (Duerr and Torres, 1999 and 2000). Vertical head difference between the lake and the Upper Floridan aquifer was computed using the lake stage measurement closest in time to when head measurements were made for the potentiometric-surface map. Vertical head gradient was calculated from this head difference and aquifer thickness.

Average and median lake stage (for 8 years of stage data; typically from May 1992-April 2000), as well as the difference between maximum and minimum lake stage, variance, and standard deviation, were computed for each lake from the stage record (most stage data were from Southwest Florida Water Management District, written commun., 2000). A subset of the data using only one observation per month was used so that data were consistent between lakes and between years. Secchi depth and specific conductance data were available for each lake from when the lake was sampled. Other recent water-quality data used were from SWFWMD (34 lakes), Polk County (6 lakes), and USGS (1 lake). Water-quality data were available for 14 of the lakes sampled by SWFWMD from the time that the lake was sampled for isotopes. For the rest of the lakes, the most recent water-quality data were used (typically from the previous 1-2 years before sampling).

Statistical Analysis

Multiple linear (least squares) regression was used to determine whether basin characteristics could be used to estimate ground-water inflow to lakes in central Florida. Multiple linear regression models predict the relation between a response (or dependent) variable and several explanatory (or independent) variables:

$$Y = \beta_0 + \beta_1 x_1 + \dots \beta_k x_k + \varepsilon \ , \qquad (6)$$

where Y is the response variable, β_0 is the intercept, β_1 is the slope coefficient for the first explanatory variable (x_1), β_k is the slope coefficient for the k^{th} explanatory variable (x_k), and ε is the error (or the remaining unexplained "noise" in the data). Basin characteristics data were first plotted graphically to examine their general relations with ground-water inflow computed from the isotope mass-balance approach. A correlation matrix was also used to determine which basin characteristics had statistically significant correlations ($p<0.05$) with ground-water inflow. All basin characteristics with significant correlations were considered in the regression models.

Model selection criteria generally followed those presented in Helsel and Hirsch (1992, p. 315). "Best" models were chosen as those with the lowest Mallow's Cp values. In addition, slope coefficients for all explanatory variables were statistically significant (α level 0.05). Regressions residuals were evaluated for normality by examining probability plots and for uniform scatter by plotting residuals against predicted values. Explanatory variables were transformed based on linearity and scatter in partial residual plots and whether the transformation resulted in a lowered Mallow's Cp value for the model. The influence of possible outliers on regression results was evaluated by comparing residuals to the residuals computed if individual data points were omitted. Multicollinearity was assessed by examining the variance inflation factor and the relation (or r^2) between each explanatory variable and all other explanatory variables in the model. Models were computed for the entire group of lakes and for a subset that had water-quality data. Separate models were also examined for different geographic areas.

CLIMATIC, HYDROLOGIC, AND ISOTOPIC DATA

Climatic Data

Annual average air temperature, lake-surface temperature, and relative humidity were very similar at Lake Starr and Halfmoon Lake (fig. 6). Air and lake-surface temperatures were coolest in January and warmest in July and August. Relative humidity was high year round (monthly average greater than 65 percent), but typically was highest between June and October.

Monthly average climatic data were normally distributed, and the parametric t-test for paired samples was used to test whether there was a statistical difference in average climatic data at Lake Starr and Halfmoon Lake. Average air temperature (at both Lake Starr climate stations) and relative humidity (at Lake Starr land climate station) were not statistically different at the two lakes ($\alpha = 0.05$ for all statistical tests presented in this report). T-test results indicated that average lake-surface temperature was statistically different ($p = 0.009$), but this difference (0.38 °C) is within instrument measurement errors (about 0.3 and 0.5 °C at Lake Starr and Halfmoon Lake, respectively). In addition, the annual average (February 1999 to January 2000) difference in lake-surface temperature at the lakes was 0.1 °C (fig. 6), which is less than the measurement error. Thus, lake-surface temperature apparently was not significantly different at the two lakes.

Because of similarities in climatic data from Lake Starr and Halfmoon Lake (about 60 mi apart), climatic conditions for lakes that are closer together (for example, within the coastal lowlands or the central highlands) are also assumed to be similar. In addition, it is hypothesized that climatic conditions are comparable in both geographic areas for using the isotope mass-balance approach.

Hydrologic Data

During 1999, rainfall totals were 43.2 and 47.0 in/yr at Lake Starr and Halfmoon Lake, respectively. These totals are slightly lower than 30-year (1970-99) annual average rainfall at nearby sites (48.4 in/yr at Mountain Lake near Lake Starr and 49.0 in/yr at Section 21 well field near Halfmoon Lake). The first half of 2000 (January to June 2000) experienced a larger rainfall deficit than 1999 (about 11-in. and 6-in. departures from the 30-year average for that period at Lake Starr and Halfmoon Lake, respectively). Annual rainfall averaged over periods of 4 to 10 years was close to the 30-year average at Lake Starr (49 to 51 in/yr), and was slightly higher than the 30-year average at Halfmoon Lake (54 to 59 in/yr). At the other study lakes, 4-year annual average rainfall (1996-99) ranged from 45 to 64 in/yr, with an average of 55 in/yr. Lakes with the highest rainfall were in the coastal lowlands, where 1997-98 had unusually high rainfall (greater than 65 in/yr at many sites).

Energy-budget evaporation rates at Lake Starr ranged from 55.1 to 57.6 in/yr for 1996 to 1999, with a 4-year average of 56.5 in/yr. These values are comparable to open-water evaporation rates in south Florida (German, 2000) and other studies in Florida where evaporation was computed using the energy-budget method (Sacks and others, 1994; Lee and Swancar, 1997). Although these evaporation rates are higher than those reported on regional maps (for example, Farnsworth and others, 1982), the regional maps may not be accurate because they are based on pan evaporation data, with pan coefficients derived from a small number of data-collection sites. The energy-budget method is considered to be one of the most accurate methods of computing lake evaporation (Winter, 1981).

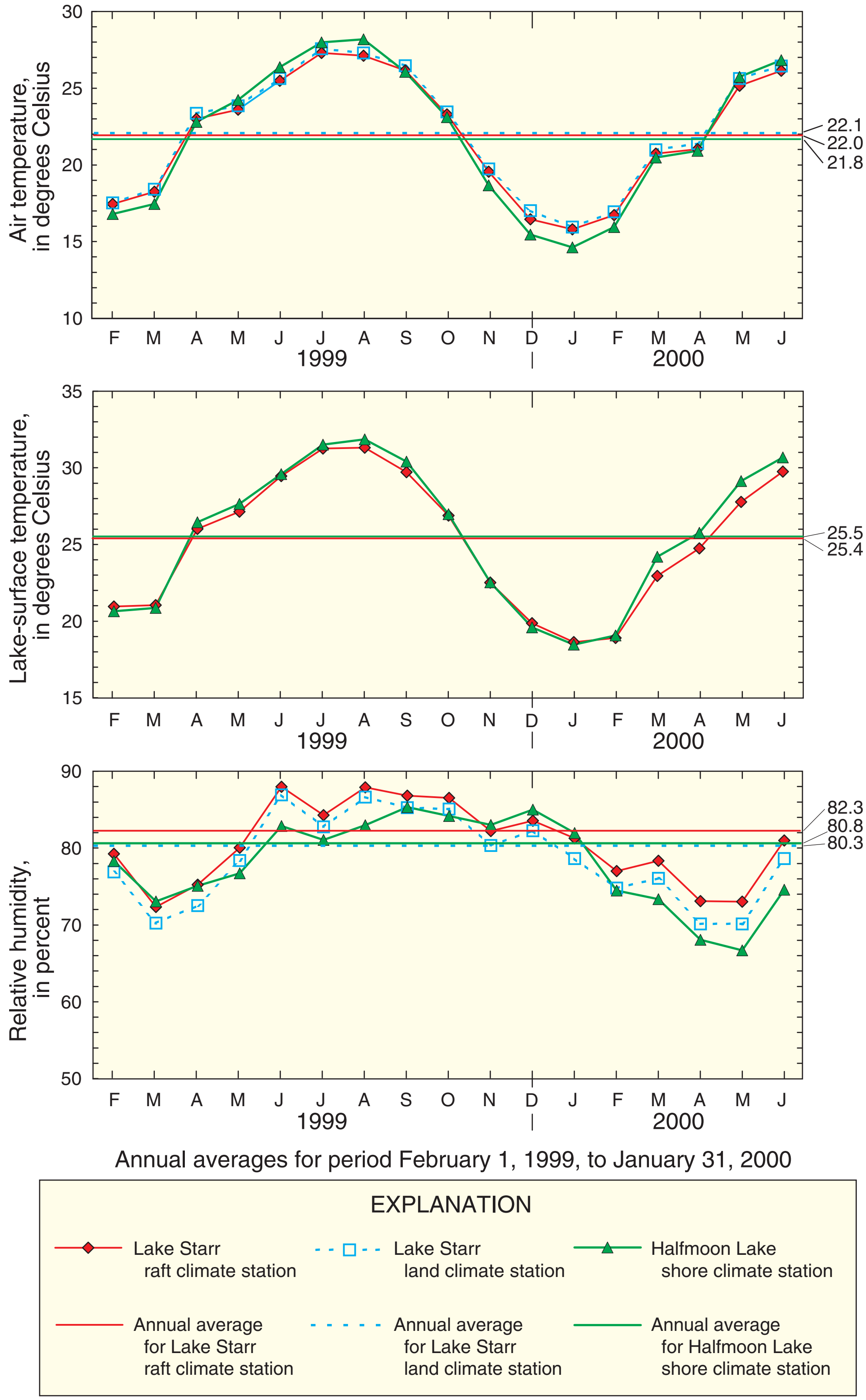
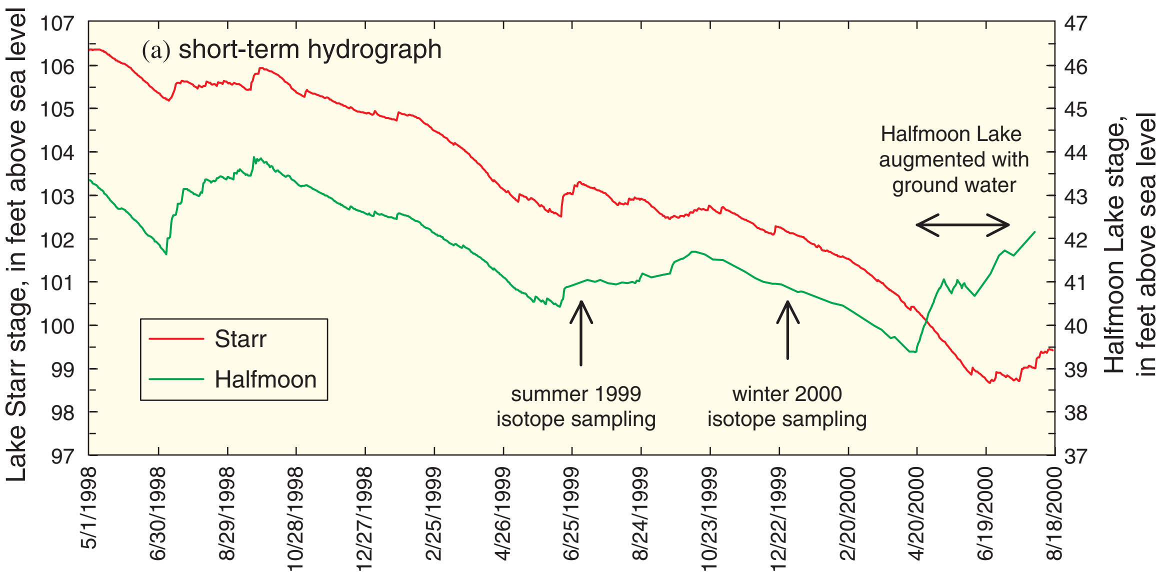


Figure 6. Comparison of monthly average air temperature, lake-surface temperature, and relative humidity at Lake Starr and Halfmoon Lake, February 1999 through June 2000.

The stage of both Lake Starr and Halfmoon Lake declined during much of the study (January 1999 through July 2000; fig. 7a). Most of the stage decline at Lake Starr occurred between January and May 2000, when rainfall was more than 10 in. below the 30-year average and ground-water inflow was more than 8 in. below the median ground-water inflow for this period (1996-99). During the study, Lake Starr's stage varied by 6.28 ft, and Halfmoon Lake's stage varied by 3.05 ft. The stage of Halfmoon Lake would have dropped more if it had not been augmented by ground water between April and July 2000. Over longer periods of time, seasonal stage variation is encompassed within longer term stage variation (multiple years) in response to rainfall deficit or surplus. Both lakes had stage variations ranging between 8 and 9 ft over the almost 20 years that stage has been monitored (fig. 7b). The period when the larger group of lakes was sampled (summer 1999 to winter 2000) was characterized by relatively stable stages at Lake Starr and Halfmoon Lake (fig. 7a). The stage of most lakes varied by less than 1 ft between the two sampling periods.

Annual ground-water inflow to Lake Starr ranged from 14.8 to 39.5 in/yr between 1996 and 2000 (table 2). Ground-water inflow averaged over periods



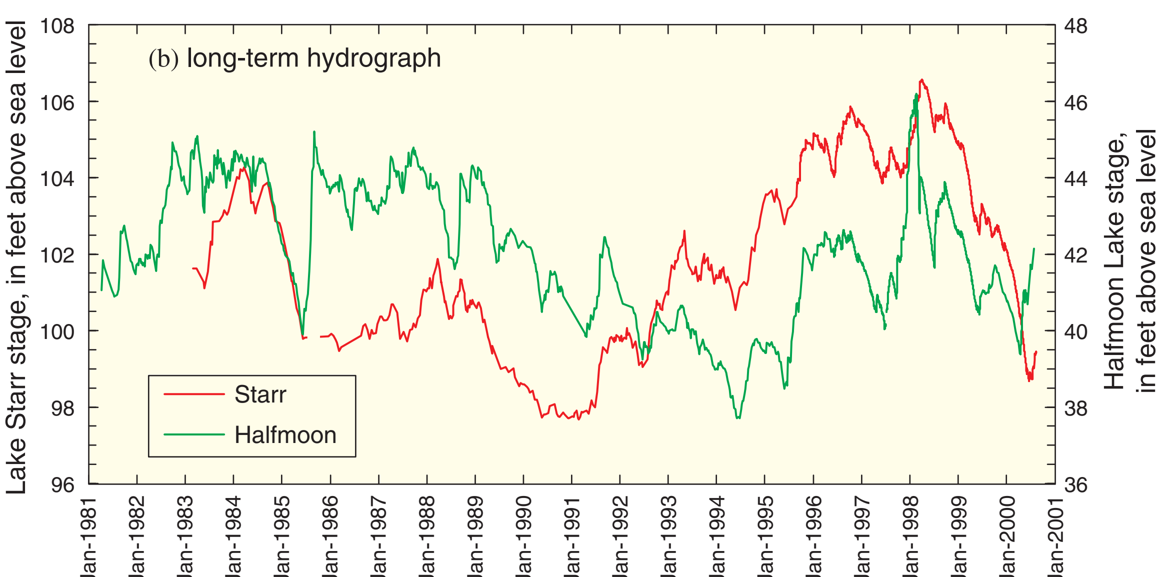


Figure 7. (a) Short-term and (b) long-term hydrographs of the stage of Lake Starr and Halfmoon Lake.

between 2 and 4 years (through 1999) ranged from 26.8 to 29.4 in/yr, respectively. Ground-water inflow varied monthly and was lower during the dry season than the wet season because of lower water-table elevations (fig. 8). Annual ground-water outflow (leakage) was less variable than ground-water inflow, and ranged from 26.0 to 30.8 in/yr. Ground-water outflow is primarily controlled by the head difference between the lake and the Upper Floridan aquifer. Annual net ground-water flow for 1999 was -16.0 in/yr (ground-water outflow exceeded inflow). Net ground-water flow averaged over periods between 2 and 4 years (through 1999) was more balanced and ranged from –1.6 to 1.5 in/yr.

Net ground-water outflow also predominated at Halfmoon Lake for 1999 (-15.5 in/yr) (table 2). Net ground-water flow averaged over 2 to 3 years (through 1999) was more balanced and ranged from –0.2 to 8.0 in/yr. Beginning in April 2000, Halfmoon Lake was temporarily augmented with ground water in an experiment to see how lake stage would respond. The effect of augmenting the lake was notable in the monthly water budget, when net ground-water outflow (negative net ground-water flow) doubled between March and May 2000, and was higher than any of the other months (fig. 9). The effects of augmentation can also be seen in annual net ground-water outflow (June 1999-May 2000), when it was higher (more negative) than the previous 3 years (table 2). Ground-water outflow increases and inflow decreases when a lake is augmented because lake stage is artificially elevated above the water table (Belanger and Kirkner, 1994; Metz and Sacks, 2002). Independent ground-water inflow data were not available for Halfmoon Lake for the study period.

Surface-water inflow was estimated for three lakes, and stormwater inflow was estimated for seven lakes. Surface-water inflow was computed by relating instantaneous discharge measurements to discharge at a nearby site with continuously gaged discharge. Total inflow to Crooked Lake (sum of inflows at three channels) had the best relation to flow at Carter Creek (USGS station number 02270000), about 22 mi to the southeast; total inflows to Lake Clinch (sum of inflows at three channels) had the best relation with Tiger Creek (USGS stations number 02268390), about 9 mi

Table 2. Annual water budget for Lake Starr (1996-2000) and Halfmoon Lake (June 1996-May 2000)

[Units in inches, normalized over lake surface area; P, precipitation; E, evaporation; Aug, augmentation of lake with pumped ground water; S_o, surface-water outflow or pumping from lake; ΔV, change in lake volume; G_i, ground-water inflow; G_o, ground-water outflow; G_{net}, net ground-water flow; --, not available]

Year	P	E	Aug	S_o	ΔV	G_i	G_i error[1]	G_o	G_o error[2]	G_{net}	G_{net} error
Lake Starr:											
1996	54.7	56.8	0	1.2	7.4	37.2	8.3	26.5	5.3	10.7	6.4
1997	54.7	56.5	0	2.2	-6.2	26.0	8.7	28.2	5.6	-2.2	6.7
1998	44.9	55.1	0	3.2	0.1	39.5	8.5	26.0	5.2	13.5	6.8
1999	43.2	57.6	0	2.6	-33.0	14.8	9.2	30.8	6.2	-16.0	6.9
2000	31.0	59.7	0	3.8	-48.3	14.9	9.8	30.6	6.1	-15.8	7.7
Halfmoon Lake:											
1997	74.0	52.0	0	0	46.3	--	--	--	--	24.3	13.2
1998	63.9	51.7	0	60.0	-32.7	--	--	--	--	15.1	31.2
1999	47.0	52.0	0	0	-20.5	--	--	--	--	-15.5	8.2
June 96 - May 97	47.2	52.9	0	0	-18.1	--	--	--	--	-12.4	10.7
June 97 - May 98	84.5	50.7	0	42.6	22.1	--	--	--	--	30.9	24.2
June 98 - May 99	49.0	52.3	0	17.4	-24.2	--	--	--	--	-3.5	12.0
June 99 - May 00	38.4	52.0	40.8	0	5.2	--	--	--	--	-22.1	9.9

[1]Root mean square error of remaining water-budget terms (Swancar and Lee, 2000).

[2]Estimated to be 20 percent.

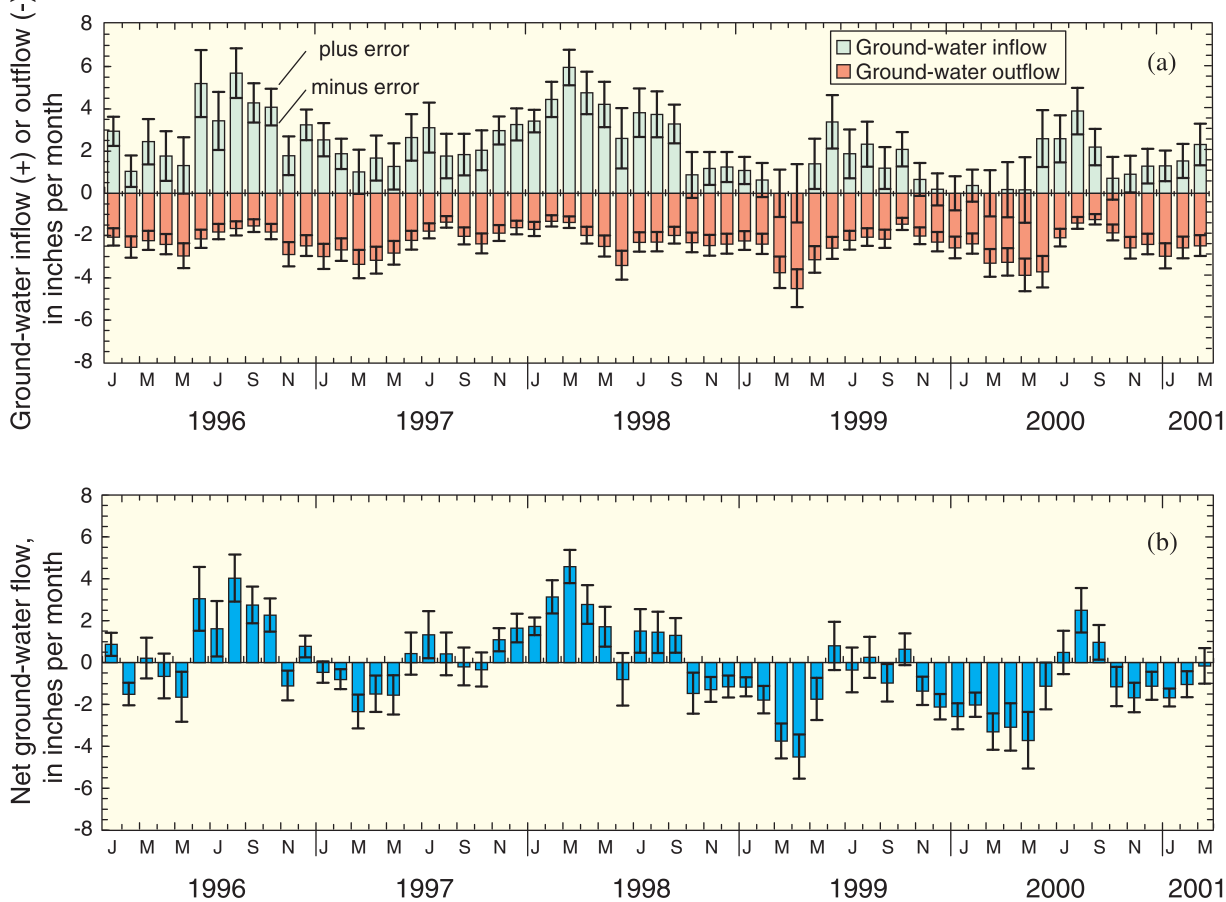


Figure 8. Monthly (a) ground-water inflow and outflow and (b) net ground-water flow at Lake Starr, January 1996 through March 2001.

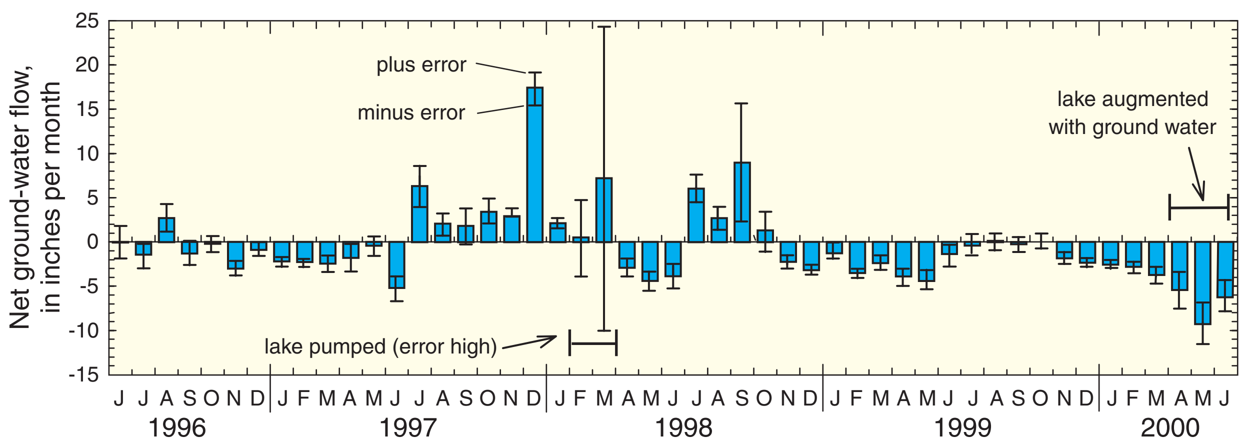


Figure 9. Monthly net ground-water flow at Halfmoon Lake, June 1996 through June 2000.

to the northeast; and Lake Lotela inflow had the best relation with Livingston Creek (USGS station number 02269520), about 9 mi to the north/northeast (fig. 10). Based on these relations, annual surface-water inflow was computed for the three study lakes for the period 4/1/99 to 3/31/00 (table 3) and for the 4-year period 1996-99. Stormwater inflow to the seven study lakes with significant amounts of stormwater inflow was estimated to range from 4 to 37 percent of rainfall on the lake's surface, based on assumed impervious areas. Uncertainties in estimates of surface-water inflow and stormwater inflow are high, probably on the order of 100 percent.

Table 3. Instantaneous discharge in inflow channels to Crooked Lake, Clinch Lake, and Lake Lotela

[Units in cubic feet per second, unless otherwise noted; n/a, not available]

Measurement date	Crooked Lake[1]	Lake Clinch[1]	Lake Lotela
6/2/99	0	n/a	n/a
8/13/99	8.2	1.6	0.10
9/9/99	13	0.94	1.7
10/26/99	15	1.5	4.3
12/8/99	1.9	0.52	0.96
3/1/00	0	0.18	0
Estimate of annual surface-water inflow (4/1/99-3/31/00)[2]			
in 10^7 cubic feet per year	23	3.3	4.4
in inches per year [3]	12	7.6	15

[1]Total discharge from three inflow channels.
[2]Estimated from regression relation with instantaneous discharge from nearest continously gaged site; see figure 10.
[3]Volumetric flow divided by average lake surface area.

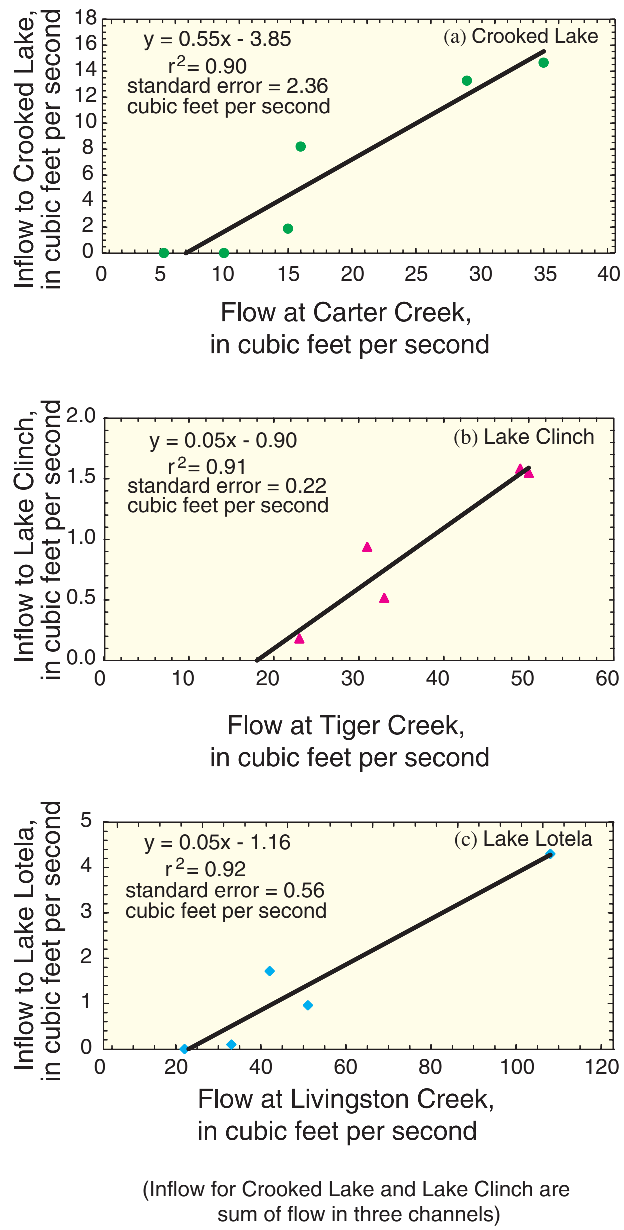


Figure 10. Relations between inflows to (a) Crooked Lake, (b) Lake Clinch, and (c) Lake Lotela, and flow at Carter, Tiger, and Livingston Creeks, respectively.

Isotopic Data

Isotopic data for rainwater, atmospheric moisture, ground-water inflow, surface-water inflow, and lake water were collected for this study. Comparisons were made between data collected at Lake Starr and Halfmoon Lake, as well as temporal changes in isotopic composition.

Rainwater

The annual (1999) volume-weighted mean (VWM) isotopic composition of rainwater was very similar for both Lake Starr and Halfmoon Lake (table 4). Monthly isotopic data for rainwater (app. A) were normally distributed, and so t-tests for paired data were used to compare data at the two lakes. Results indicate that there was no statistical difference between the average isotopic composition of rainwater at the two sites ($p = 0.32$ and 0.40 for δD and $\delta^{18}O$, respectively).

No discernible seasonal trend was observed in the isotopic composition of rainwater (fig. 11), as is commonly found in temperate climates and continental sites (Rozanski and others, 1993). Temperature is presumably not an important control on the isotopic composition of rainwater in Florida because of the relatively moderate, subtropical climate and its proximity to the sea. There was a poor, but statistically significant, inverse relation between monthly rainfall total and the isotopic composition of rainwater at Lake Starr (r^2 = 0.34 and 0.30 for δD and δ^{18}O, respectively); however, these relations were not statistically significant at Halfmoon Lake. Inverse relations between rainfall amount and isotopic composition of rainfall are common for tropical marine sites (Rozanski and others, 1993).

Table 4. Average isotopic composition of rainwater and ground water at Lake Starr and Halfmoon Lake

[Units in δ per mil; δD, delta deuterium, δ^{18}O, delta oxygen-18]

	δD	δ^{18}O
***Rainwater*[1]:**		
Lake Starr	-17.9	-3.78
Halfmoon Lake	-18.9	-3.91
***Ground water*[2]:**		
Lake Starr	-18.2	-3.61
Halfmoon Lake	-18.1	-3.52

[1]Volume-weighted mean for January 1999-early January 2000.

[2]Average of nine samples at Lake Starr and seven samples at Halfmoon Lake.

The δD and δ^{18}O data for rainwater define the local meteoric water line (MWL) (fig. 12). The MWLs from Lake Starr and Halfmoon Lake were very similar, and were not statistically different. Thus, data were combined from both sites, and the local MWL is defined as $\delta D = 7.73\ \delta^{18}O + 11.62$. The slope and intercept of this local MWL is not statistically different from the global MWLs defined by Craig (1961) and Rozanski and others (1993).

Atmospheric Moisture

Atmospheric moisture samples were substantially depleted in D and ^{18}O compared to rainwater, but plot near the local MWL (fig. 13). The isotopic composition of atmospheric moisture (δ_a) at Lake Starr and Halfmoon Lake varied from -67.7 to -112.3 per mil for δD and -10.87 to -16.78 per mil for δ^{18}O (app. B). Data from the two lakes were compared using t-tests for paired data, after determining that the data were normally distributed. Results for the 10 paired monthly samples indicate that there is not a

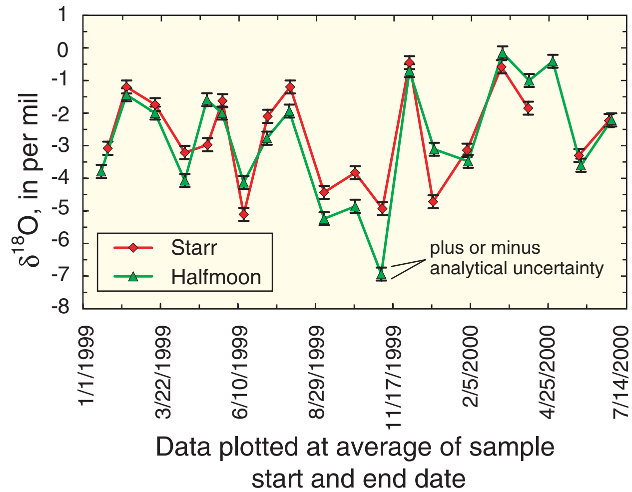


Figure 11. Monthly δ^{18}O values for rainwater at Lake Starr and Halfmoon Lake, January 1999 to July 2000.

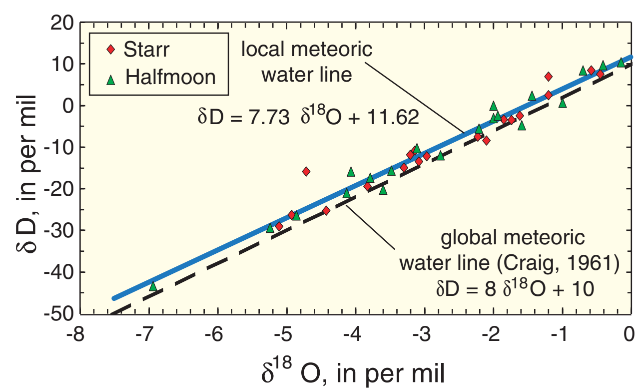


Figure 12. Relation between δD and δ^{18}O for rainwater at Lake Starr and Halfmoon Lake.

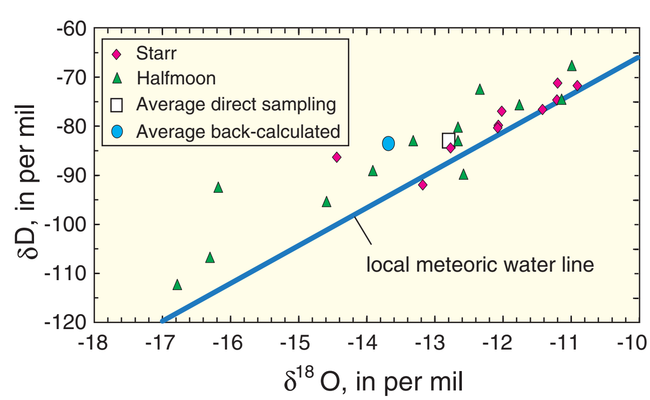


Figure 13. Relation between δD and δ^{18}O for atmospheric moisture at Lake Starr and Halfmoon Lake.

statistically significant difference between the average δ_a at the two lakes for the period sampled ($p = 0.20$ and 0.13 for δD and $\delta^{18}O$, respectively). Less data were available for Lake Starr than for Halfmoon Lake because two Lake Starr sample bottles were broken, and one additional sample was collected at Halfmoon Lake during a cold front (12/1/1999) to evaluate differences in δ_a. Because of the lesser amount of data from Lake Starr and the lack of a notable difference between the sites, a monthly average δ_a was computed using data from both sites where available. From these monthly averages, an annual average δ_a was computed (-83.0 and -12.80 per mil for δD and $\delta^{18}O$, respectively).

Relations were evaluated between δ_a and climatic variables such as air temperature, relative humidity, vapor pressure, wind speed, and atmospheric pressure. For δD, there were no statistically significant ($\alpha = 0.05$) relations between δ_a and these variables. For $\delta^{18}O$, however, there were weak, but statistically significant, relations between δ_a and air temperature, relative humidity, and vapor pressure (best $r^2 = 0.38$ for vapor pressure). Relations for $\delta^{18}O$ data were probably better because of the better analytical precision in the $\delta^{18}O$ analysis. In temperate, continental climates, δ_a has been correlated with humidity and temperature (Schoch-Fischer and others, 1984; White and Gedzelman; 1984). The more moderate climate of Florida may make these relations less apparent. Limited time series samples of δ_a from Miami, Florida, were enriched in D and ^{18}O and had much less variability compared to samples from continental sites (Schoch-Fischer and others, 1984). For the current study, the monthly point sampling of δ_a somewhat restricts our understanding of what controls variability in δ_a. However, these limited data are particularly valuable because such measurements are scarce in locations with subtropical, coastal climates like Florida.

An alternative method of calculating δ_a was back-calculating δ_a from the independent estimate of ground-water inflow from Lake Starr, using equations 4 and 5. Hydrologic data were averaged over 4 years, and steady-state conditions were assumed. The average isotopic composition of lake water was calculated from the July 1999 and January 2000 samplings. The calculated δ_a value for δD (-84.8 per mil) was very similar to the average δ_a value from direct sampling (-83.0 per mil) and was within the 2 per mil analytical uncertainty for δD (table 5). For $\delta^{18}O$, however, the back-calculated value of δ_a (-14.01 per mil) was distinctly lower than the sampled δ_a value (-12.80 per mil), and the difference was much greater than the 0.2 per mil analytical uncertainty for $\delta^{18}O$.

Table 5. Variables used to back-calculate the isotopic composition of atmospheric moisture from the Lake Starr water budget

[P, precipitation; E, evaporation; S_o, water pumped from lake; G_i, ground-water inflow; T_a, air temperature; T_o, lake-surface temperature; h, relative humidity normalized to average lake-surface temperature; δD, delta deuterium; $\delta^{18}O$, delta oxygen-18; δ, isotopic composition of hydrologic term in delta notation; L, lake water; a, atmospheric moisture; α^*, equilibrium isotope fractionation factor (Majoube, 1971); $\Delta\varepsilon$, kinetic fractionation factor: 12.5 (1 - h) for δD and 14.2 (1 - h) for $\delta^{18}O$; ε, total fractionation factor: $1{,}000 (1 - \alpha^*) + \Delta\varepsilon$]

Hydrologic variables[1] (inches per year)	Value	Climatic variables	Value	Isotopic variables (units of per mil, except α^*)	δD	$\delta^{18}O$
P	49.4	T_a (°C)	22.1	δ_L	10.8	2.32
E	56	T_o (°C)	25.4	δ_{Gi}	-18.2	-3.61
S_o	2.3	h (ratio)	0.679	δ_P	-17.9	-3.78
G_i	29.4			δ_E[2]	-29.7	-6.17
				δ_E[3]	-33.6	-8.70
				δ_a[4]	-84.8	-14.01
				δ_a[5]	-83.0	-12.80
				α^* (ratio)	0.92674	0.99074
				$\Delta\varepsilon$	4.0	4.56
				ε	77.3	13.82

[1]From Lake Starr water budget, averaged over 4 years (1996-99).
[2]Calculated from equation 4, rearranged to solve for δ_E.
[3]Calculated from sampled δ_a and other variables in equation 5 (presented as comparison).
[4]Calculated from equation 5 using back-calculated δ_E and rearranged to solve for δ_a.
[5]Sampled (presented for comparison).

Reasons for the difference between sampled and back-calculated δ_a values for $\delta^{18}O$ are not known. If isotopic fractionation of δ_a occurred during sampling due to incomplete removal of moisture, both δD and $\delta^{18}O$ should be similarly affected (Schoch-Fischer and others, 1984; Gibson and others, 1999). Alternatively, another value in equation 5 used to back-calculate δ_a may be ill-defined, with the resulting computed value affecting $\delta^{18}O$ more than δD. For example, Zimmerman (1979) preferred using δD over $\delta^{18}O$ because kinetic fractionation, which is less accurately defined than equilibrium fractionation, is much lower for δD than for $\delta^{18}O$. For temperature and humidity conditions during this study, kinetic fractionation ($\Delta\varepsilon$) accounts for only 5 percent of total fractionation (ε) for δD, but accounts for 32 percent of total fractionation for $\delta^{18}O$. An alternative explanation may be that samples of δ_a were enriched in ^{18}O because of localized recirculated water vapor from lake evaporate. In a study in northern Canada, Gibson and others (1999) concluded that δ_a enriched in ^{18}O was the result of recycled terrestrial moisture (localized evaporate), which is not representative of δ_a as described in the Craig and Gordon (1965) evaporation model. Given the scope of this study, reasons for the difference in $\delta^{18}O$ of atmospheric moisture remain unresolved.

Finally, δ_a was estimated by assuming it was in isotopic equilibrium with rainwater. Using the VWM δ_P from Lake Starr and the average annual air temperature, δ_a for $\delta^{18}O$ was -13.41 per mil, which falls between the back-calculated and sampled δ_a values. In contrast, δ_a for δD was considerably lower (-100.7 per mil) than the sampled and back-calculated values of δ_a (around -84 per mil), which indicates that δ_a is not in isotopic equilibrium with local rainwater for δD. This contradicts results from Sacks and others (1998), but direct sampling of δ_a and a rigorous water budget to independently calculate δ_a were not available for that study.

Ground-Water and Surface-Water Inflow

The average isotopic composition of ground-water inflow at Lake Starr and Halfmoon Lake was virtually identical and was also very similar to the VWM isotopic composition of rainwater for 1999 (table 4). The similarity in the isotopic composition of ground water and rainwater indicates that evaporation does not significantly enrich recharge water in D and ^{18}O. Sandy soils in the study area cause rapid infiltration during rainfall, limiting the time that standing water remains at land surface. Infiltrated water, however, is likely lost through transpiration, which does not cause fractionation of the isotopes as does evaporation (Krabbenhoft and others, 1994; Clark and Fritz, 1997).

Ground-water inflow data from the other 12 lakes had a somewhat larger range in isotopic composition (fig. 14; app. C), with median δD and $\delta^{18}O$ values of –18.0 and –3.57 per mil, respectively. Median values are similar to those from Lake Starr and Halfmoon Lake (table 4). At several lakes, δD and $\delta^{18}O$ values of ground water were offset to the right of the MWL, indicating that they were influenced by evaporation. In these cases, lake water enriched in D and ^{18}O probably seeped into the ground water at an earlier time when the lake level was higher than the water table. Ground-water flow reversals were observed at three lakes between the wet season (October 1999) and the dry season (March-April 2000) samplings.

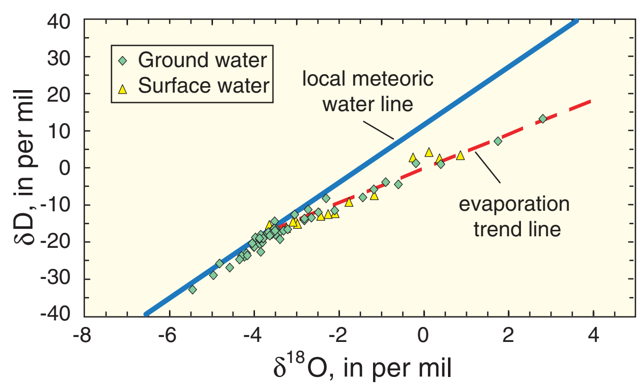


Figure 14. Relation between δD and $\delta^{18}O$ for ground-water and surface-water inflows to lakes.

Surface-water inflow was sampled from seven inflow channels to three lakes (Crooked Lake, Clinch Lake, and Lake Lotela). Surface-water inflow typically was offset to the right of the MWL, indicating the influence of evaporation (fig. 14; app. D). The isotopic composition of surface water can help identify its origins. When surface water originated from an upgradient lake or wetland, its isotopic composition plotted on an evaporation trend line. In contrast, when surface water originated from ground-water seepage, it was closer in isotopic composition to rainwater (meteoric water).

Lake Water

The isotopic composition of Lake Starr and Halfmoon Lake varied seasonally (fig. 15), primarily in response to net precipitation (rainfall minus evaporation). When evaporation exceeded rainfall, the lakes became enriched in D and ^{18}O, and when rainfall exceeded evaporation, the lakes became depleted in D and ^{18}O (fig. 16). Because Halfmoon Lake is shallower and contains a smaller volume of water than Lake Starr, the isotopic composition of Halfmoon Lake responded more quickly to short-term changes in rainfall and evaporation compared to Lake Starr. Because most changes in lake volume for these seepage lakes are in response to net precipitation, changes in lake volume are also highly correlated with changes in the isotopic composition of lake water.

During the spring of 2000, δD and $\delta^{18}O$ values increased at both Lake Starr and Halfmoon Lake (fig. 15). At Lake Starr, rainfall was about 10 in. below the 30-year average between March and May 2000, and ground-water inflow was more than an order of magnitude lower than the average for these months over the previous 4 years (0.3 in. compared to 6.5 in.). In addition, evaporation at Lake Starr for April and May 2000 was about 20 percent higher than evaporation over this same period for previous years (1997-99; period of record for evaporation data). The lack of dilution from isotopically light rainfall and ground-water inflow, along with higher evaporation rates, caused the lake to become enriched in D and ^{18}O (more positive δ values) during this period. Halfmoon Lake was augmented with ground water between April and July 2000, and the direct addition of ground water with lower δD and $\delta^{18}O$ values resulted in lake water becoming noticeably depleted in D and ^{18}O.

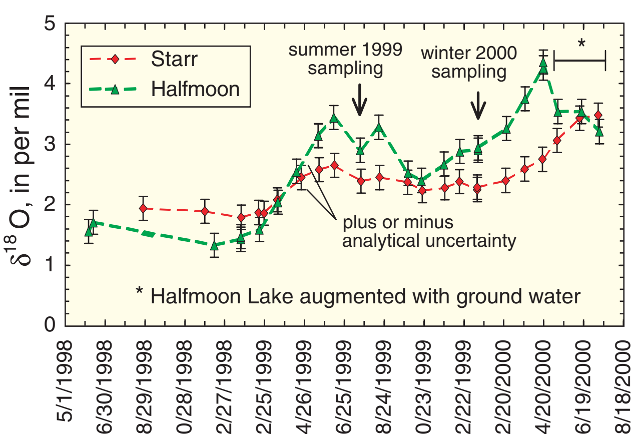
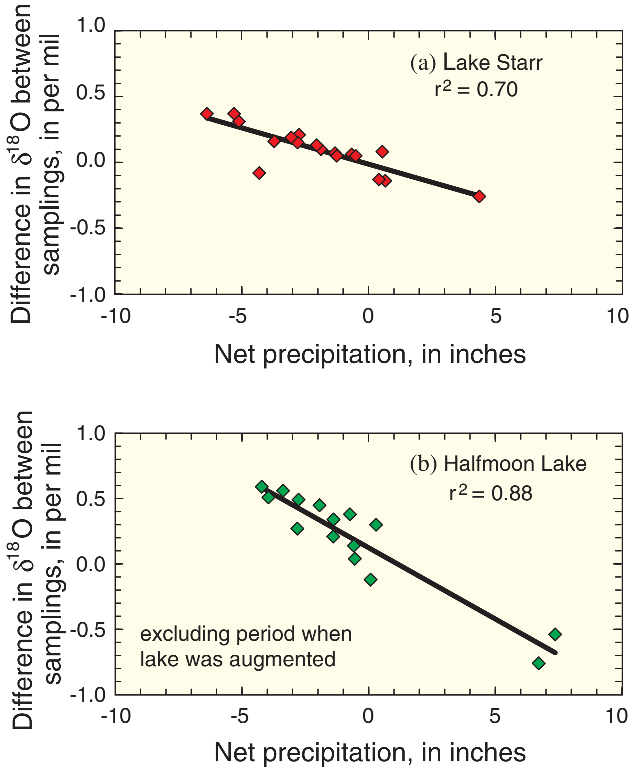


Figure 15. Values of $\delta^{18}O$ for lake water from Lake Starr and Halfmoon Lake between the summer of 1998 and the summer of 2000.



[Relations were poorer for δD (r^2= 0.46 and 0.56 for Lake Starr and Halfmoon, respectively) than for $\delta^{18}O$ because of poorer analytical precision.]

Figure 16. Relation between the difference in lake-water $\delta^{18}O$ values between samplings and net precipitation (rainfall minus evaporation) for (a) Lake Starr and (b) Halfmoon Lake.

Most of the study lakes were sampled twice – once during the summer of 1999 and once during the winter of 2000. For more than half the lakes, δD and $\delta^{18}O$ values of the summer and winter samples were identical within analytical uncertainty (app. E). Thus, this “snapshot” in time corresponded to a period when most of the lakes had a relatively stable isotopic composition. For lakes that were sampled semiannually (between the summer of 1998 and the summer of 2000), the isotopic composition changed more noticeably (fig. 17), with δD and $\delta^{18}O$ values increasing between the winter and summer of 2000 when rainfall was below average. There was less isotopic variability for lakes in the central highlands, which tend to be deeper, than for the shallower lakes in the coastal

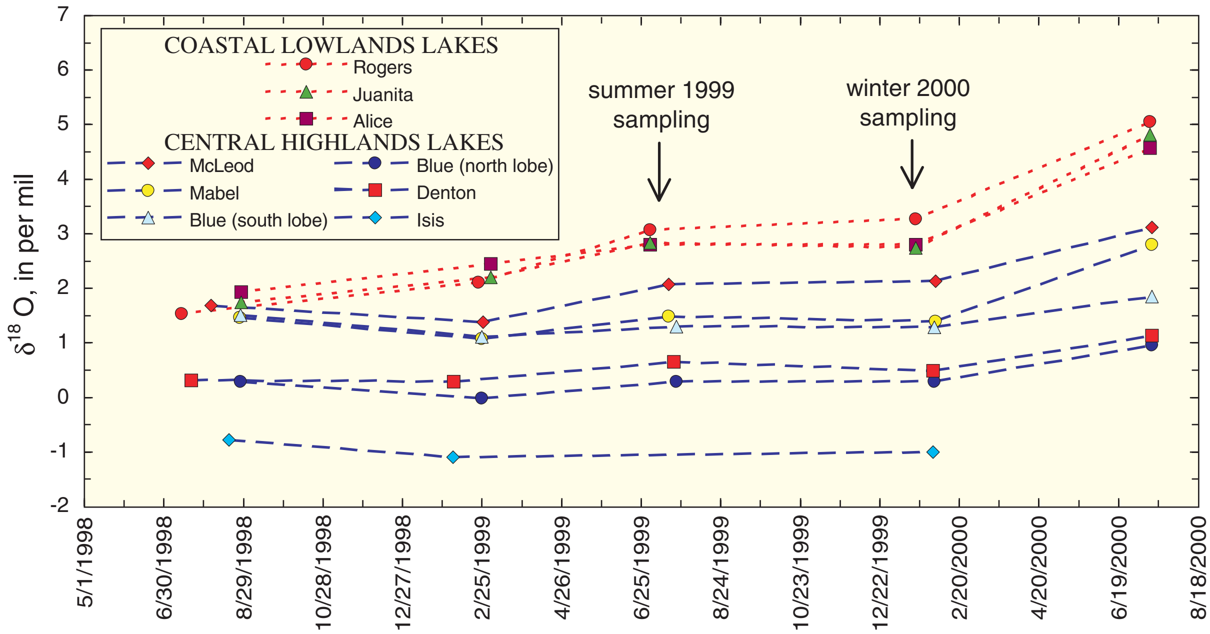


Figure 17. Values of $\delta^{18}O$ for water from lakes sampled semiannually between the summer of 1998 and the summer of 2000.

lowlands. Deeper lakes have larger lake volumes that "buffer" the lakes from short-term changes in net precipitation. In addition, when the water table drops significantly during periods of drought, shallow lakes can have larger reductions in ground-water inflow than deep lakes. Consequently, use of the isotope mass-balance approach between years is more precise for deep lakes than for shallow lakes. Using isotopic data from shallow lakes sampled during different years would likely cause erroneous comparisons of ground-water inflow. For this study, the lakes were all sampled during the same time period, which should allow ground-water inflow conditions to be compared between lakes.

Isotopic data from all the study lakes plot along an evaporation trend line ($\delta D = 4.67\ \delta^{18}O - 0.21$), which is offset to the right of the local MWL (fig. 18). The good relation between lake water δD and $\delta^{18}O$ values on the evaporation line (r^2=0.89) is support for similar atmospheric conditions influencing evaporation at all the study lakes. The evaporation line intersects the MWL at the VWM isotopic composition of precipitation (fig. 18). The isotopic composition of lake water can be used to qualitatively compare ground-water inflow to seepage lakes in the same area. For example, lakes that plot low on the evaporation trend line are depleted in D and ^{18}O because of more flushing with ground water with low δD and $\delta^{18}O$ values. In contrast, lakes that plot higher on the evaporation trend line are enriched in D and ^{18}O, indicating the relatively greater effect of evaporation on the total lake volume. Highland lakes tended to be more depleted in D and ^{18}O than lowlands lakes (fig. 18).

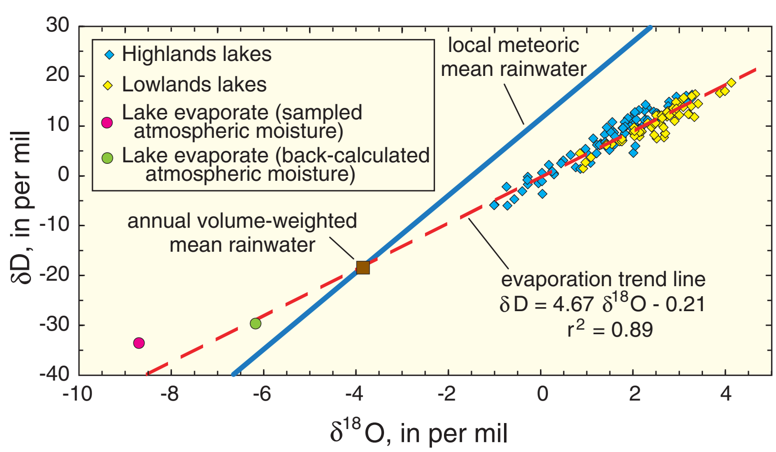


Figure 18. Relation between δD and $\delta^{18}O$ for lake water collected in the summer of 1999 and winter of 2000, and for average evaporate from Lake Starr computed using sampled and back-calculated values for atmospheric moisture.

The slope of the evaporation line is consistent with the theoretical slope of the evaporation line computed from annual average climatic and isotopic data collected at Lake Starr. The theoretical slope can be calculated by:

$$E_{slope} = \frac{h(\delta_a - \delta_I)_D + \varepsilon_D}{h(\delta_a - \delta_I)_{18} + \varepsilon_{18}}, \qquad (7)$$

where E_{slope} is the slope of the evaporation line; h is relative humidity normalized to lake-surface temperature (defined in eq. 5); δ_a is the isotopic composition of atmospheric moisture; δ_I is the isotopic composition of inflows to the lake (volume-weighted precipitation and ground-water inflow); and ε is the total fractionation factor (defined in eq. 5); subscripts D and 18 designate that the term is for δD or $\delta^{18}O$, respectively (Gat, 1971). When back-calculated values of δ_a were used, the calculated slope of the evaporation line (4.68) was virtually identical to the slope of the evaporation line from the larger group of lakes sampled (4.67; fig. 18). The similarity between the theoretical and observed slopes supports the assumption that conditions at Lake Starr are representative of lakes in the greater study area. When the average sampled δ_a values were used in equation 7, the slope was considerably lower (4.34) than the observed slope, because the average sampled $\delta^{18}O$ value was higher than the back-calculated value.

The isotopic composition of water evaporating from a lake (δ_E) should plot on the evaporation trend line, but to the left of the meteoric water line (Clark and Fritz, 1997). The average δ_E at Lake Starr was computed using equation 5, data in table 5, and both values of δ_a (sampled and back-calculated from the Lake Starr water budget). The δ_E value using the back-calculated δ_a plots directly on the evaporation trend line as expected, whereas the δ_E value using the sampled δ_a is offset from this line (fig. 18). This casts further doubt on δ_E results for $\delta^{18}O$ when δ_a was defined from direct sampling.

ESTIMATES OF GROUND-WATER INFLOW

Ground-water inflow was estimated for the sampled lakes using both δD and $\delta^{18}O$ values and both steady-state and transient formulations of equation 4. Further, both sampled and back-calculated values of δ_a were used in the calculations. A sensitivity analysis evaluated uncertainty in terms on the computed ground-water inflow.

Steady-State Results

In order to evaluate the timeframe assumed for the steady-state analysis, ground-water inflow from Lake Starr's water budget was compared to ground-water inflow from the isotope mass-balance approach, using different assumptions about the steady-state timeframe. Rainfall and evaporation were averaged over periods ranging from 1 to 10 years, and δ_a that was used was from direct sampling. The best comparison with ground-water inflow from the water budget was for δD, using hydrologic data averaged over 4 years (table 6). This "steady-state" time period is slightly longer than the hydraulic residence time for Lake Starr of 2.5 years (average lake volume/total inflows for 4-year period 1996-99). These results are consistent with a steady-state timeframe that is at least as long as the lake's residence time. Results using 5 to 10 years of hydrologic data also compared well with ground-water inflow averaged over 4 years. However, independent estimates of ground-water inflow are not available for periods longer than 4 years (before 1996). Similarity of results using the isotope mass-balance approach for the 4 to 10 year timeframe, and their consistency with water budget results, adds confidence to using the steady-state approach.

For Lake Starr, ground-water inflow results using $\delta^{18}O$ data were higher than the water-budget estimates, and they were outside the uncertainty limits of the water budget (isotope mass-balance results between 49 and 65 in/yr, compared to water-budget results ranging from 11 to 38 in/yr; table 6). These higher ground-water inflow values indicate that δ_a values from direct sampling were not correct. As discussed earlier, the δ_a value back-calculated from the water budget was lower (-14.01 per mil) than the average $\delta^{18}O$ value from direct sampling (-12.80 per mil), and is probably more accurate.

Ground-water inflow at Lake Starr was computed using two approaches: (1) δ_E values were computed monthly and weighted annually by the monthly evaporation rate, and (2) δ_E values were calculated from annual average data (δ_a, δ_L, T_o, h) and semiannual lake samples. Results from these two approaches were very similar (table 6). Because of

Table 6. Comparison of ground-water inflow to Lake Starr from water-budget and isotope mass-balance approaches

[δ_E, isotopic composition of water evaporating from lake; in/yr, inches per year; gw, ground water; WB, water budget; δD, delta deuterium, $\delta^{18}O$, delta oxygen-18; δ_a, isotopic composition of atmospheric moisture; --, not applicable]

Method to calculate ground-water inflow	δ_E method	Steady-state timeframe[1]	Ground-water inflow (in/yr)[2]	Range of gw inflow considering WB error (in/yr)
water budget	--	1 year	15	(6 - 24)
δD, steady state	weighted[3]	1 year	41	--
δD, steady state	average[4]	1 year	43	--
water budget	--	2 years	27	(18 - 36)
δD, steady state	weighted	2 years	42	--
δD, steady state	average	2 years	37	--
water budget	--	3 years	27	(18 - 36)
δD, steady state	weighted	3 years	34	--
δD, steady state	average	3 years	39	--
water budget	--	4 years	29	(21 - 38)
δD, steady state	weighted	4 years	32	--
δD, steady state	average	4 years	37	--
δD, steady state	weighted	5 years	31	--
δD, steady state	average	5 years	36	--
δD, steady state	weighted	10 years	32	--
δD, steady state	average	10 years	36	--
water budget		7/20/99-1/11/00	15	(5 - 25)
$\delta^{18}O$, transient	weighted	7/20/99-1/11/00	70	--
$\delta^{18}O$, transient	average	7/20/99-1/11/00	65	--
δD, transient	weighted	7/20/99-1/11/00	34	--
δD, transient	average	7/20/99-1/11/00	39	--
δD, transient	average[5]	7/20/99-1/11/00	27	--

[1]Ending in 1999 (for example, 2 years is 1998-99).

[2]For the transient time period, ground-water inflow was normalized to annual units by dividing by the number of days in transient period (175) and multiplying by 365.

[3]δ_E computed monthly and weighted annually by evaporation rate.

[4]δ_E computed from annual average data.

[5]Using annual average δ_a of -83.0 per mil (May 1999-April 2000), rather than average for transient period (-80.8 per mil).

these similarities and because monthly data are only available for Lake Starr and Halfmoon Lake, δ_E values were computed from annual average data when computing ground-water inflow values for all of the study lakes. Thus, annual average values of ε, Δε, α*, and h (table 5) were substituted into equation 5, and δ_E was solved as a function of δ_L (average of the summer 1999 and winter 2000 samples) and δ_a. The value of δ_a for δD was the annual average sampled from Lake Starr and Halfmoon Lake, and for $\delta^{18}O$ the back-calculated value from the Lake Starr water budget was used. Hydrologic data were averaged over 4 years. Annual (1999) VWM δ_P and average δ_{Gi} from Lake Starr were used, and these values were virtually identical to data from Halfmoon Lake (table 4).

For the δD calculations, ground-water inflow to the 81 lakes ranged from 0 to 285 in/yr, or 0 to 83 percent of total water inflows to the lake (table 7). Median ground-water inflow was 37 in/yr or 41 percent of total inflows. Slightly negative ground-water inflow values were calculated for two lakes, which indicates that these lakes had very low amounts of ground-water inflow, and small uncertainties in terms caused the results to be less than 0. When negative results were obtained, values were set to a very small positive value (0.1 in/yr) for statistical analyses.

Ground-water inflow results using $\delta^{18}O$ values, with δ_a back-calculated from the Lake Starr water budget, were lower than results using δD measurements for many of the lakes (table 7). These differences are because $\delta^{18}O$ results were fixed to the Lake Starr water-budget estimate of ground-water inflow, but δD results were computed independently. Ground-water inflow values derived from δD data were slightly higher than those determined from the Lake Starr water-budget estimate, although they were still within the error bounds (table 6). Thus, ground-water inflow based on δD data may be high for the other lakes as well. Other differences in ground-water inflow based on $\delta^{18}O$ and δD data are probably related to uncertainties in the lake isotopic analyses. Ground-water inflow values derived from $\delta^{18}O$ data ranged from 0 to 258 in/yr (0 to 81 percent of total inflows), with a median of 34 in/yr (or 36 percent of total inflows). The median ground-water inflow value for lakes in the central highlands (48 in/yr or 49 percent of total inflows) was higher than the median for lakes in the coastal lowlands (12 in/yr or 17 percent of total inflows). Negative results were calculated for five lakes (table 7).

Table 7. Ground-water inflow results using steady-state and transient isotope mass-balance approaches

[δD, delta deuterium; $\delta^{18}O$, delta oxygen-18; δ_a, isotopic composition of atmospheric moisture; calc'd, calculated; G_i, ground-water inflow; in/yr; inches per year; %, percent; n/a, not available]

Map reference number	Lake name	Steady state δD and δ_a sampled[1]		Steady state $\delta^{18}O$ and δ_a calc'd[2]			Transient δD and δ_a sampled[3]	
		G_i (in/yr)	G_i (% of total inflows)	G_i (in/yr)	G_i (% of total inflows)	G_i category[4]	G_i (in/yr)	G_i (% of total inflows)
Lakes in the coastal lowlands:								
1	Alice	18	24	9	13	low	11	16
2	Allen	33	34	20	24	low	9	13
3	Bird	33	35	14	19	low	30	38
4	Boat	51	43	37	35	medium	59	54
5	Calm	19	25	12	18	low	20	30
6	Carroll	33	32	16	19	low	54	60
7	Deer	30	33	14	18	low	29	31
8	Egypt	50	37	37	30	medium	30	32
9	George	81	49	49	36	medium	87	58
10	Halfmoon	15	20	2	4	low	23	34
11	Hobbs	20	24	6	9	low	2	3
12	Hog Island	35	36	25	28	medium	24	28
13	Juanita	17	23	6	10	low	27	35
14	LeClare	16	21	1	2	low	24	34
15	Merrywater	23	27	6	9	low	15	23
16	Mound	54	49	37	39	medium	40	41
17	Osceola	19	25	9	14	low	12	18
18	Raleigh	25	30	11	16	low	18	26
19	Rogers	1	1	[5]0	0	low	0	0
20	Starvation	40	41	21	28	medium	46	49
21	Taylor	28	34	15	22	low	3	3
22	Van Dyke	68	55	62	52	high	76	62
23	Big Fish	[5]0	0	[5]0	0	low	1	2
24	Black	49	46	11	16	low	44	52
25	Camp	50	47	13	19	low	29	29
26	Crews	20	26	14	20	low	24	40
27	Curve	14	20	[5]0	0	low	12	16
28	Gooseneck	54	49	40	41	medium	43	38
29	King	14	19	5	8	low	[5]0	0
30	Linda	34	37	17	23	low	25	27
31	Moon	15	22	8	13	low	11	17
32	Pierce	[5]0	0	[5]0	0	low	[5]0	0
33	Thomas (Pasco)	8	12	2	4	low	[5]0	0
34	Wistaria	22	28	[5]0	0	low	[5]0	0
Median, coastal lowlands		24	29	12	17	low	24	29
Lakes in the central highlands (Lake Wales Ridge and Intraridge Valley):								
35	Angelo	43	44	34	39	medium	55	50
36	Chilton	81	61	65	56	high	78	58
37	Denton	94	63	105	66	high	107	66
38	Dinner (Highlands)	45	43	57	49	medium	58	51
39	Isis	274	81	258	80	high	n/a	n/a
40	Lotela	24	25	36	33	medium	12	10
41	Olivia	132	72	130	71	high	n/a	n/a
42	Pioneer	42	43	46	46	medium	46	45
43	Tulane	141	72	160	75	high	163	75
44	Verona	152	67	181	71	high	200	73
45	Viola	68	56	70	56	high	77	58

Table 7. Ground-water inflow results using steady-state and transient isotope mass-balance approaches --(Continued)

[δD, delta deuterium; $\delta^{18}O$, delta oxygen-18; δ_a, isotopic composition of atmospheric moisture; calc'd, calculated; G_i, ground-water inflow; in/yr; inches per year; %, percent; n/a, not available]

Map reference number	Lake name	Steady state δD and δ_a sampled[1]		Steady state $\delta^{18}O$ and δ_a calc'd[2]			Transient δD and δ_a sampled[3]	
		G_i (in/yr)	G_i (% of total inflows)	G_i (in/yr)	G_i (% of total inflows)	G_i category[4]	G_i (in/yr)	G_i (% of total inflows)
Lakes in the central highlands (Lake Wales Ridge and Intraridge Valley):--Continued								
46	Aurora	108	69	92	66	high	144	76
47	Blue (south lobe)	61	51	60	50	high	70	59
48	Blue (north lobe)	132	69	119	67	high	136	74
49	Crystal	34	43	44	49	medium	22	37
50	Dinner (Polk)	70	58	61	55	high	71	65
51	Hickory	71	60	64	58	high	61	54
52	Josephine	285	83	243	80	high	278	85
53	Little Aurora	219	82	172	78	high	196	82
54	Mabel	65	56	61	54	high	48	54
55	Menzie	46	50	36	44	medium	30	45
56	Saddlebag	67	59	61	56	high	86	66
57	Saint Anne	215	82	198	81	high	269	86
58	Silver	43	48	64	58	high	40	44
59	Starr	37	43	29	37	medium	27	40
60	Wales	28	34	38	41	medium	15	25
61	Warren	78	60	36	41	medium	67	62
Lakes in the central highlands (other ridge and upland areas):								
62	Clinch	35	40	42	44	medium	27	30
63	Crooked	11	13	20	22	low	17	19
64	Eagle	17	24	25	32	medium	3	7
65	Grassy	4	7	1	2	low	7	12
66	Helene	12	18	5	8	low	[5] 0	0
67	Henry	39	43	37	42	medium	30	40
68	Little Van	112	66	97	63	high	108	69
69	Lizzie	41	44	39	43	medium	55	55
70	Lucerne	12	21	7	13	low	[5]0	0
71	McLeod	24	32	34	39	medium	12	24
72	Medora	8	12	5	9	low	2	4
73	Polecat	111	68	104	67	high	111	71
74	Sara	33	42	27	37	medium	25	41
75	Tennessee	38	40	36	39	medium	36	45
76	Thomas (Polk)	37	41	29	36	medium	50	53
77	Walker	53	50	54	51	high	80	64
78	Iola	30	35	23	30	medium	1	2
79	Jessamine	168	76	150	74	high	148	82
80	Pasadena	6	10	6	11	low	7	16
81	Spring	52	50	48	48	medium	56	53
Median central highlands		46	50	48	49	medium	55	53

[1]δ_a from direct sampling.
[2]δ_a back-calculated using ground-water inflow from Lake Starr's water budget.
[3]δ_a from direct sampling (average for entire sampling period, rather than non steady-state period).
[4]Ground-water inflow categories using $\delta^{18}O$ steady-state results: low, G_i less than 25% of total inflows to lake; medium, G_i 25-50% of inflows; high, G_i greater than 50% of inflows.
[5]Negative ground-water inflow calculated; value set to 0.1 in/yr (0.2% of inflows) for statistical analyses.

The above analysis was based on assumptions of steady-state conditions. During the period of time when all of the lakes were sampled for this study (summer 1999 and winter 2000), lake stage and isotopic composition were relatively uniform, supporting short-term steady-state conditions. However, this is not necessarily an indicator of long-term steady-state conditions. Hydraulic steady state relies on consistent precipitation from year to year, whereas Florida's climate is characterized by multiple years of drought or excess rainfall. As a result of extremes in rainfall, Florida seepage lakes can have a wide range in stage and volume over these short-term cycles. These changes in stage should have a significant effect on the isotopic composition of lakes. At sites with long periods of record (more than 20 years), lake-stage data cycle around longer-term average conditions (provided there are no anthropogenic influences, such as large ground-water withdrawals). Similarly, it is reasonable to assume that the isotopic composition of Florida lakes cycles around long-term, steady-state values. Because long-term isotopic data are not available for the study lakes, the validity of steady-state isotopic conditions remains unresolved. Steady-state assumptions were probably not valid for some of the study lakes. In particular, shallow lakes in the coastal lowlands can have considerable variability in isotopic composition, both seasonally and between years (fig. 17). The following section computes ground-water inflow values using the transient formulation of equation 4; these values are then compared with those obtained using steady-state assumptions.

Transient Results

Ground-water inflow was also computed using a transient approach to the isotope mass-balance, whereby changes in lake volume and lake isotopic composition between samplings were considered. Similar to the steady-state approach, comparisons were made for Lake Starr between ground-water inflow derived from water-budget and isotope mass-balance approaches. Transient results of ground-water inflow were higher than water-budget results for both δD and $\delta^{18}O$ data, but δD results agreed more closely with the Lake Starr water budget (table 6) than did the $\delta^{18}O$ results. These ground-water inflow results were calculated using δ_a values averaged for the transient period July 20, 1999, to January 11, 2000 (equivalent to the time between the summer 1999 and winter 2000 lake samplings). Because δ_a values are the average of point samples, the true mean δ_a value for the entire period is not known and may be somewhat different from this average. When the average δ_a value for δD for the entire 12-month data-collection period was used (-83.0 per mil; app. B) rather than the average for the transient period (-80.8 per mil), ground-water inflow results using δD data were more similar to those calculated from water-budget data (table 6). Although the ground-water inflow value (27 in/yr) was higher than the water-budget derived ground-water inflow value (5-25 in/yr), small uncertainties in the isotopic composition of the lake (within analytical uncertainty) can bring results well within the limits of water-budget results. Effects of uncertainties in the lake isotopic composition on computed ground-water inflow will be discussed further in the following section. Because of better agreement with the independent water-budget results, the average δ_a value for the entire period was used to calculate ground-water inflow values for the rest of the study lakes, when using δD data. For $\delta^{18}O$, a back-calculated value for δ_a was used.

Transient ground-water inflow results using δD data had a similar range as the steady-state results (0 to 278 in/yr, or 0 to 86 percent of inflows; table 7), and they were highly correlated to the steady-state values of ground-water inflow ($r^2 = 0.94$; fig. 19). Median ground-water inflow was 30 in/yr, or 40 percent of inflows. Calculated ground-water inflow was more often negative for transient results than for steady-state results (6 out of 79 lakes for transient, compared to 2

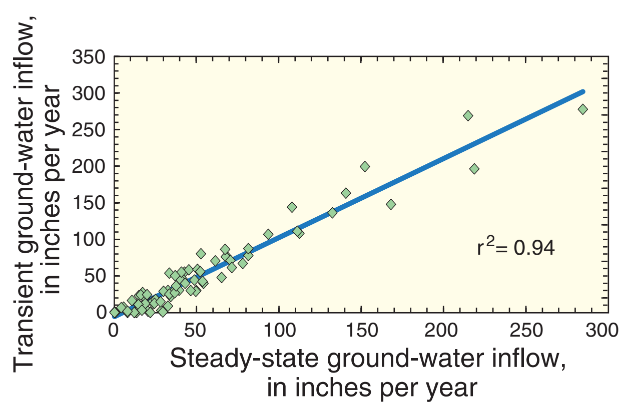


Figure 19. Relation between transient and steady-state ground-water inflow results computed from the isotope mass balance approach using δD.

out of 81 lakes for steady state using δD data). Two lakes (Lakes Isis and Olivia) were not included in the transient analysis because they were not sampled in the summer 1999. The greater number of negative values is because small uncertainties in terms for transient calculations can cause large changes in ground-water inflow. Results using $\delta^{18}O$ data and δ_a values back-calculated from the Lake Starr water budget were more problematic, with negative ground-water inflow computed for almost half of the lakes (36 out of the 79 lakes).

Sensitivity Analysis

Uncertainties in terms in the isotope mass-balance equation affect the degree of uncertainty in the calculated ground-water inflow results. Sensitivity analyses were conducted for Lake Starr using both δD and $\delta^{18}O$ data, and for steady-state and transient calculations. To compare these results to lakes with different amounts of ground-water inflow, sensitivity analyses were also examined for a lake with higher ground-water inflow (105 in/yr) and a lake with lower ground-water inflow (8 in/yr) than Lake Starr (29 in/yr). Variables evaluated in the steady-state sensitivity analysis were P, E, δ_L, δ_{Gi}, δ_P, and δ_a, relative humidity (RH), lake-surface temperature (T_o), and air temperature (T_a); for the transient analysis, δ_{L1}, δ_{L2}, V_1, and V_2 (initial and final lake isotopic compositions and initial and final lake volumes, respectively) were also evaluated. Initially, the hydrologic variables P, E, V_1, and V_2 were changed ± 10 percent of the original values, and the isotope variables were changed ± the analytical uncertainty for that isotope ratio (95 percent confidence interval; 2 per mil for δD and 0.2 per mil for $\delta^{18}O$); for the climatic variables, RH was changed ± 1 percent relative humidity, and temperature was changed ± 0.3 °C. Results for Lake Starr are discussed first, followed by results for the lakes with higher and lower ground-water inflow. Lastly, the effect of uncertainties in surface-water and stormwater inflow on uncertainties in ground-water inflow is examined for a subset of 10 lakes.

Of the hydrologic variables, ground-water inflow (G_i) was most sensitive to changes in E (table 8). For steady-state calculations, a 10 percent change in E caused computed G_i to change by 8 in/yr (or 27 percent of the original value; table 8). Results for P were slightly less sensitive, with a 10 percent change in P causing G_i to change by 5 in/yr (or 17 percent of the original value). These results point to the need for accurate estimates of lake evaporation when using the isotope mass-balance approach. In this study, evaporation was estimated using the energy-budget method, which is considered to be one of the most accurate methods (Winter, 1981). However, uncertainty in energy-budget evaporation is still assumed to be 10 percent on an annual basis (Swancar and others, 2000), and it may be larger when extrapolating results to the larger population of lakes.

Transient ground-water inflow was similarly sensitive to changes in E and P (table 8). However, the error as a fraction of the original value was higher for the transient analysis because transient ground-water inflow was much lower than steady-state inflow at Lake Starr (15 in/yr compared to 29 in/yr). For example, a 10 percent change in E caused a 5 to 6 in/yr change in G_i, but this resulted in a change of 31 to 41 percent from its original value. Transient G_i values were much less sensitive to uncertainties in initial or final lake volumes compared to uncertainties in the other hydrologic variables. A 10 percent change in lake volume caused a 3 percent change in G_i (about 0.5 in/yr). Lake volume was not well defined for many of the study lakes (see methods). The lack of sensitivity of lake volume on G_i results indicates that an estimate of lake volume, rather than a precise quantification, is acceptable.

Of the isotopic variables, G_i was most sensitive to changes in δ_L values (table 8). This is most pronounced for δD, which had poorer analytical precision than $\delta^{18}O$. For the steady-state calculations using δD data, changing δ_L values within analytical uncertainty caused about 43 percent change in G_i from the original value (or a change of about 13 in/yr). For $\delta^{18}O$ data, the sensitivity of G_i to changes in δ_L values was half that for δD data (table 8), reflecting its better analytical precision. For the steady-state analysis, changes in δ_L values influence G_i values primarily through the δ_E term, in which δ_L is a variable (eq. 5), rather than through δ_L directly in equation 4. Between 85 and 90 percent of the change in G_i is from the change in δ_E, rather than from δ_L in equation 4. For the transient analysis, computed G_i was even more sensitive to changes in δ_{L1} and δ_{L2}. Again, this effect was about twice as large when using δD data compared to using $\delta^{18}O$ data. For δD data, a 2 per mil change in either δ_{L1} or δ_{L2} caused more than a 200 percent change in G_i. In contrast to the steady-state analysis, most of the change (70-80 percent) in G_i is from the change in δ_{L1} or δ_{L2} directly in equation 4, and not from changes in the δ_E term.

Table 8. Summary of sensitivity analysis of ground-water inflow computed from the isotope mass-balance approach for Lake Starr

[δD, delta deuterium; $\delta^{18}O$, delta oxygen-18; G_i, ground-water inflow; in/yr, inches per year; %, percent; --, not in calculation; analy, analytical; C, Celsius]

Change in term	δD steady state		$\delta^{18}O$ steady state		δD transient		$\delta^{18}O$ transient	
	change in G_i (in/yr)	change in G_i (%)	change in G_i (in/yr)	change in G_i (%)	change in G_i (in/yr)	change in G_i (%)	change in G_i (in/yr)	change in G_i (%)
(Original G_i values)[1]:	29	--	29	--	15	--	15	--
***Precipitation* (P)**								
+10%	-5	-17%	-5	-17%	-4	-24%	-4	-27%
-10%	5	17%	5	17%	4	24%	4	27%
***Evaporation* (E)**								
+10%	8	27%	8	27%	6	41%	5	31%
-10%	-8	-27%	-8	-27%	-6	-41%	-5	-31%
***Initial lake volume* (V_1)**								
+10%	--	--	--	--	-0.5	-4%	0.5	3%
-10%	--	--	--	--	0.5	4%	-0.5	-3%
***Final lake volume* (V_2)**								
+10%	--	--	--	--	-0.5	-3%	0.4	3%
-10%	--	--	--	--	0.5	3%	-0.4	-3%
***Isotopic composition of lake water* (δ_L)**								
+ analy uncertainty[2]	-12	-40%	-6	-21%	--	--	--	--
- analy uncertainty	13	46%	7	23%	--	--	--	--
***Initial isotopic composition of lake water* (δ_{L1})**								
+ analy uncertainty	--	--	--	--	20	131%	10	65%
- analy uncertainty	--	--	--	--	-21	-140%	-10	-67%
***Final isotopic composition of lake water* (δ_{L2})**								
+ analy uncertainty	--	--	--	--	-31	-205%	-16	-105%
- analy uncertainty	--	--	--	--	33	220%	16	109%
***Isotopic composition of atmospheric moisture* (δ_a)**								
+ analy uncertainty	8	27%	4	13%	9	58%	4	29%
- analy uncertainty	-8	-27%	-4	-13%	-9	-58%	-4	-29%
***Isotopic composition of ground-water inflow* (δ_{Gi})**								
+ analy uncertainty	2	7%	1	3%	1	7%	0.5	3%
- analy uncertainty	-2	-6%	-1	-3%	-1	-6%	-0.5	-3%
***Isotopic composition of rainwater* (δ_P)**								
+ analy uncertainty	3	12%	2	6%	3	18%	1	9%
- analy uncertainty	-3	-12%	-2	-6%	-3	-18%	-1	-9%
***Relative humidity* (RH)**								
+1% humidity units	-3	-12%	-5	-18%	-4	-26%	-7	-44%
-1 % humidity units	3	11%	5	17%	4	25%	6	42%
***Lake-surface temperature* (T_o)**								
+0.3 degrees C	3	10%	7	23%	4	26%	9	57%
-0.3 degrees C	-3	-11%	-7	-25%	-4	-28%	-10	-64%
***Air temperature* (T_a)**								
+0.3 degrees C	-5	-18%	-8	-28%	-6	-41%	-11	-70%
-0.3 degrees C	5	16%	8	25%	6	38%	10	63%

[1]Using δ_a back calculated from the Lake Starr water budget for both isotopes, so that computed ground-water inflow was the same for consistent comparisons.

[2]Analytical uncertainty (95% confidence interval) for δD is 2 per mil and for $\delta^{18}O$ is 0.2 per mil.

The sensitivity of computed G_i to uncertainty in δ_L illustrates the need to accurately define the lake isotopic composition, and also points out a limitation in the transient approach. Fortunately, sampling lakes in Florida is relatively simple because they are typically well mixed, and, therefore, "grab" samples are representative of whole-lake isotopic composition. In addition, the analytical uncertainty reported for the isotopes is the "worst case" scenario, and analytical precision was typically much better than this (T.B. Coplen, U.S. Geological Survey, written commun., 2001). For example, duplicate samples from this study were all within 1.2 and 0.07 per mil of each other for δD and $\delta^{18}O$, respectively. The better analytical precision for $\delta^{18}O$ makes this isotope preferable over δD in the isotope mass-balance approach because of the sensitivity of G_i to δ_L. However, this study found difficulties using $\delta^{18}O$ data to estimate another term in the isotope mass balance, δ_a, which was not the case for δD data. G_i was overestimated when using $\delta^{18}O$ data from direct sampling of δ_a. To overcome this problem, δ_a was back-calculated from the independent estimate of ground-water inflow from Lake Starr's water budget, and, as a result, $\delta^{18}O$ data could be used more accurately in the isotope mass balance.

Of the other isotopic variables (δ_{Gi}, δ_P, and δ_a), G_i was most sensitive to changes in δ_a, which is used to compute δ_E (eq. 5). Similar to δ_L results, fractional changes were greater for the transient than for the steady-state calculations and were greater when using δD data than $\delta^{18}O$ data (table 8). An accurate estimate of δ_a is difficult because of uncertainties in extrapolating point measurements over the entire study period and because little data exist on factors controlling δ_a. Continuous sampling of δ_a is expensive and time consuming, and rarely has been done (Schoch-Fischer and others, 1984; Gibson and others, 1999). In addition, reasons remain unresolved as to why direct sampling of δ_a for oxygen isotopes did not provide estimates of G_i that were comparable to water-budget estimates. Thus, it is clear that an independent water budget, used in conjunction with the isotope mass-balance approach, greatly improves the confidence in the value of δ_a and subsequent estimates of G_i.

Computed G_i was less sensitive to changes in δ_{Gi} and δ_P, than δ_a (table 8). Changing δ_{Gi} within analytical uncertainty caused less than a 10 percent change in G_i. Because of the spatial nature of ground-water inflow, δ_{Gi} has to be defined from point measurements. In addition, δ_{Gi} can vary spatially around a lake due to ground-water flow reversals and isotopic variability in recharge. A subset of the study lakes was sampled to define δ_{Gi}, with only a few samples collected from most of those lakes. However, results indicate that average δ_{Gi} is well-constrained in the study area and is similar to δ_P. Sensitivity analyses indicate that the sampling of δ_{Gi} for this study was sufficient because computed G_i was not very sensitive to changes in δ_{Gi}. In fact, VWM δ_P could have been used in place of δ_{Gi} with less than a 3 percent change in calculated G_i. Changing δ_P within analytical uncertainty resulted in about twice the change in G_i, compared to δ_{Gi}, but these changes are still low compared to effects of changes in other variables (table 8). VWM δ_P can be readily obtained by sampling rainwater in a sampler that is closed to evaporation. The similarity of annual VWM δ_P at both Lake Starr and Halfmoon Lake indicates that δ_P was adequately defined for the study area.

G_i was relatively sensitive to changes in climatic variables (RH, T_o, and T_a) (table 8). These variables were used indirectly to compute δ_E in the calculations of h (relative humidity normalized to water-surface temperature), α^*, and $\Delta\varepsilon$ (eq. 5). Sensitivity of G_i to these variables was about twice as high when using $\delta^{18}O$ data compared to δD data. For example, a 0.3 °C change in T_o caused a 3 in/yr (or about 10 percent) change in G_i for δD and a 7 in/yr (about 24 percent) change for $\delta^{18}O$. This greater sensitivity is related to the greater importance of kinetic fractionation ($\Delta\varepsilon$, which is a function of h) to total fractionation during evaporation for oxygen isotopes than for hydrogen isotopes. Thus, although $\delta^{18}O$ data are preferable to δD data because of better analytical precision, δD data are preferable in defining isotopic fractionation from climatic data. Hence, use of either isotope ratio has its advantages and disadvantages. The sensitivity of G_i to climatic variables illustrates the need for accurate measurements in computing δ_E. For this study, RH, T_o, and T_a were measured continuously at Lake Starr and Halfmoon Lake, and routine site inspections increased the confidence in their measurements. Uncertainty in these climatic variables, as well as in δ_L and δ_a, increases the uncertainty in δ_E, which, in turn, affects the uncertainty of computed G_i. Results from this sensitivity analysis illustrate the importance of δ_E in the isotope mass balance of Florida lakes and also illustrates how uncertainties in δ_E, which cannot be measured directly, greatly affect uncertainties in G_i.

For lakes with lower and higher ground-water inflow than Lake Starr, the actual change in G_i (in in/yr) for a given change in one of the variables was very similar to Lake Starr (table 8). The lake with higher G_i typically had slightly greater changes in computed G_i

(in in/yr) than the lake with lower G_i. However, the *fractional* change from the original G_i value was less for the high ground-water inflow lake and was much greater for the low ground-water inflow lake, than Lake Starr (fig. 20). These differences are due to differences in the magnitude of change from the original G_i value. Thus, uncertainty in G_i was much less for the “high” ground-water inflow lake (less than

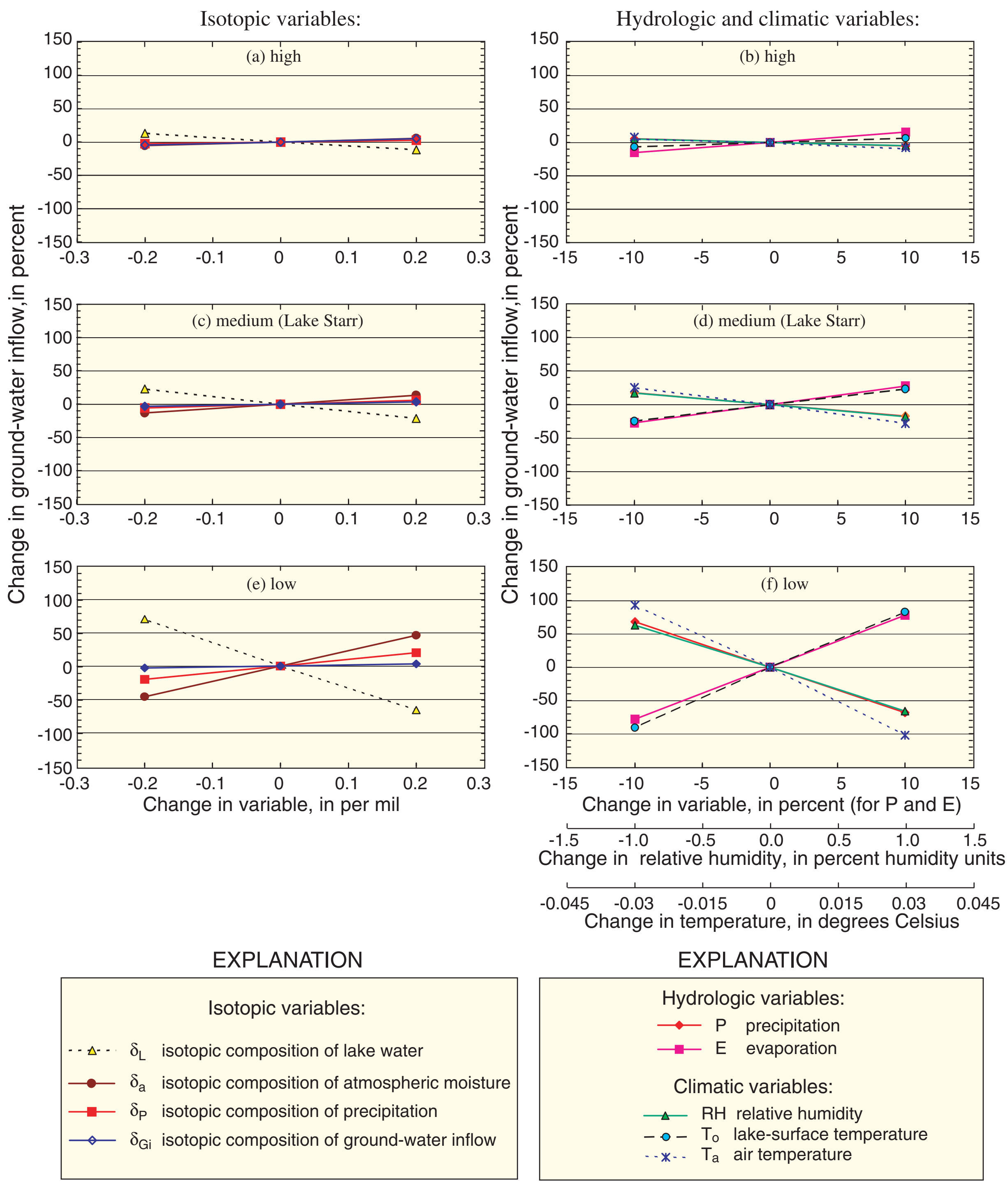


Figure 20. Steady-state sensitivity analysis of change in computed ground-water inflow to change in isotopic, hydrologic, and climatic variables used to compute ground-water inflow using $\delta^{18}O$, for (a-b) high, (c-d) medium (Lake Starr), and (e-f) low ground-water inflow lakes.

30 percent) than for the “low” ground-water inflow lake, where uncertainty may be greater than 100 percent (fig. 20). For a lake such as Lake Starr (which can be considered a “medium” ground-water inflow lake), uncertainty in G_i was probably less than 50 percent.

The isotope mass-balance approach is more successful in quantifying G_i for Florida lakes with high ground-water inflow than with low ground-water inflow. Even though uncertainty in G_i is high for lakes with low ground-water inflow, the method is sensitive enough to definitively distinguish between lakes with low and high ground-water inflow. For example, a lake with G_i of 8 in/yr and 100 percent uncertainty still has low G_i compared to a lake with G_i of 105 in/yr and 30 percent uncertainty. The better resolution of G_i at the higher range is also relevant to the role of ground water in the lake’s water budget. For example, it may be more important to quantify G_i for a lake when it is an important part of the lake’s water budget, whereas it may be sufficient to merely categorize G_i as low for a lake when it is not an important part of the water budget.

Finally, the sensitivity of G_i to uncertainty in surface-water or stormwater inflow (S_i) was examined for the 10 lakes that had S_i incorporated into the lake’s isotope mass balance. The magnitude of these flows varied from 4 to 37 percent of rainfall, and uncertainty is probably on the order of 100 percent for methods used in this study. Changes in S_i affected calculated G_i differently, depending on the magnitude of both S_i and G_i. For lakes with surface-water inflow (Lotela, Clinch, and Crooked), a 100 percent change in S_i caused G_i to change by 14, 18, and 56 percent, respectively for the $\delta^{18}O$ steady-state calculations. Uncertainty was highest for Crooked Lake because it had the lowest G_i of the three lakes and had relatively high surface-water inflow (table 3). For lakes with stormwater inflow, a 100 percent change in S_i caused G_i to change between 4 and 64 percent. Lakes with the largest sensitivity to changes in stormwater inflow were Carroll, George, and Egypt (all in the coastal lowlands), which had 40, 46, and 64 percent changes in G_i, respectively. The other three lakes had less than a 15 percent change in G_i. Therefore, consideration needs to be given to the magnitude of S_i and G_i when assessing methods to determine surface-water flows to lakes for the isotope mass-balance approach. In order to minimize uncertainties in G_i, surface-water inflow should be accurately quantified for lakes with high S_i or low G_i. This sensitivity analysis illustrates the importance of understanding the uncertainties inherent in the isotope mass-balance approach when interpreting ground-water inflow results.

Categories of Ground-Water Inflow

Although the isotope mass-balance approach quantifies ground-water inflow, the sensitivity analysis described previously illustrates the high range of uncertainty in ground-water inflow results. Therefore, the method is better used to distinguish whether ground-water inflow quantities fall within certain ranges of values. The study lakes were grouped into three categories, corresponding to low ground-water inflow (less than 25 percent of total water inflows), medium ground-water inflow (25-50 percent of total inflows), and high ground-water inflow (greater than 50 percent of total inflows). These categories were chosen to encompass the range of computed ground-water inflow conditions. Steady-state ground-water inflow results using $\delta^{18}O$ data were used to define categories because the better precision in the $\delta^{18}O$ analysis improved the accuracy of ground-water inflow results.

The distribution of ground-water inflow varied according to geographic and geomorphic areas (fig. 21; table 7). Lakes in the coastal lowlands tended to have low ground-water inflow (76 percent of lakes were in the low ground-water inflow category), and only 3 percent of lakes were in the high ground-water inflow category. Coastal lowlands lakes were in Hillsborough and Pasco Counties north of Tampa Bay. In contrast, the majority (67 percent) of central highlands lakes in the Lake Wales Ridge/Intraridge Valley in Polk and Highlands Counties were in the high ground-water inflow category, and no lakes were grouped in the low ground-water inflow category (fig. 21). Lakes in other parts of the central highlands (Winter Haven Ridge, Lake Henry Ridge, Brooksville Ridge, and Polk Upland) had 50 percent of lakes in the medium ground-water inflow range, and lesser amounts of lakes in the low and high ground-water inflow categories (fig. 21).

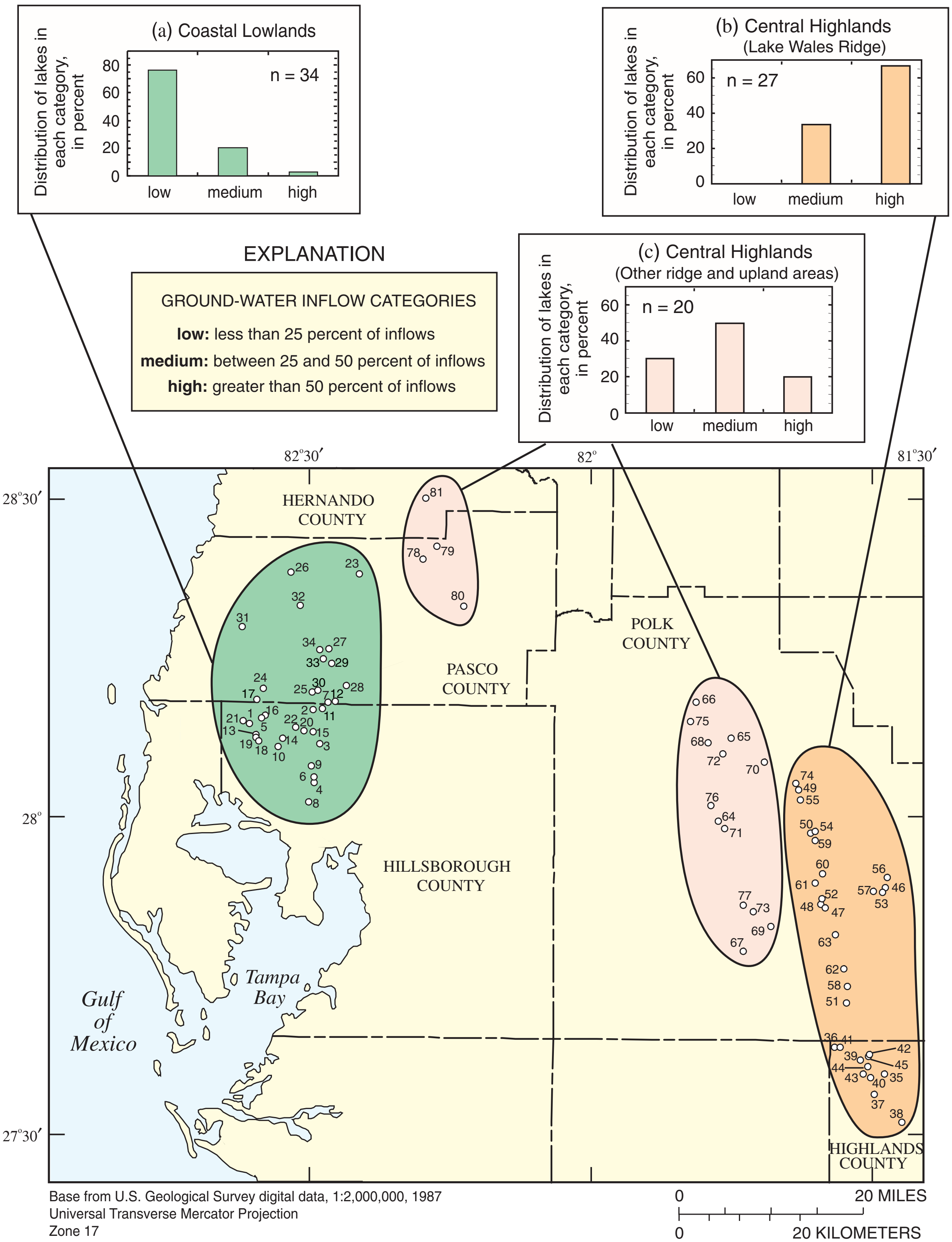


Figure 21. Fraction of lakes in each ground-water inflow category by geographic region: (a) coastal lowlands, (b) Lake Wales Ridge and Intraridge Valley, and (c) other ridge and upland areas.

STATISTICAL ANALYSIS BETWEEN GROUND-WATER INFLOW AND BASIN CHARACTERISTICS

One of the goals of computing ground-water inflow to the study lakes was to evaluate whether readily measurable basin characteristics can be used to predict ground-water inflow. If they can, then ground-water inflow can be estimated for lakes that were not sampled. Based on the range and geographic distribution of computed ground-water inflow, both regional and local basin characteristics appear to be important in predicting ground-water inflow at individual lakes. Because of the large uncertainties in ground-water inflow computed using the isotope mass-balance approach, results from multiple linear regression models should be used only to group a lake into one of the three ground-water inflow categories described previously, rather than to specifically quantify ground-water inflow for that lake.

Model Building

Basin characteristics that had statistically significant ($p<0.05$) correlation coefficients with ground-water inflow were used as potential explanatory variables in regression models (table 9). To evaluate possible differences between geographic regions, correlation matrices were also constructed separately for lakes in the three geographic areas: coastal lowlands, central highlands in the Lake Wales Ridge and Intraridge Valley, and central highlands in other ridge and upland areas. For these separate correlation matrices, far fewer variables were significantly correlated with ground-water inflow than when the entire data set was considered together.

Models were built with varying combinations of all potential explanatory variables, and the models were ranked according to the Mallow's Cp value. Models including water-quality variables were examined for a subset of the lakes because of the lower number of observations (n=47 to 54, based on the variable) compared to the entire data set (n=81). Several explanatory variables were log transformed after examining partial residuals; Mallow's Cp values were also reduced after the transformations. Any variable that did not have a statistically significant ($\alpha = 0.05$) slope was omitted from the model. Explanatory variables that were highly cross correlated were not considered in the same model. For final models, the absolute value of the variance inflation factor (VIF) for all variable was less than 2.5, indicating minimal problems with multicollinearity between variables (Helsel and Hirsch, 1992).

Table 9. Potential explanatory variables with statistically significant correlation coefficients ($p<0.05$) with ground-water inflow

[Correlations with steady-state ground-water inflow using oxygen-18 (see table 7); A soils, high infiltration rate; B/D soils, moderate infiltration rate when water table is low, and very slow infiltration rate when water table is high; m, meters]

Variable	Correlation coefficient	Number of observations
Nitrate concentration in lake	0.66	54
Maximum lake depth	0.63	81
Total nitrogen concentration in lake	0.63	53
Fraction of hydrologic group A soils in basin	0.58	81
Fraction of hydrologic group A soils within 100 m of lake	0.56	81
Fraction of hydrologic group B/D soils in basin	-0.55	81
Depth to Upper Floridan aquifer	0.54	81
Secchi depth (lake clarity)	0.52	81
Thickness of intermediate confining unit	0.51	81
Average land slope within 50 m of lake	0.50	81
Thickness of surficial aquifer	0.50	81
Sodium concentration in lake	-0.48	48
Fraction of B/D soils within 100 m of lake	-0.48	81
Basin area/lake surface area	0.48	81
Fraction of wetlands in basin[1]	-0.45	81
Iron concentration in lake	-0.42	47
Average lake stage (1992-2000)	0.38	75
Lake region	0.39	81
Thickness of surficial deposits beneath lake	0.38	81
pH	0.37	34
Fraction of wetlands in basin[2]	-0.37	81
Chloride concentration in lake	-0.35	48
Fraction of wetlands within 100 m of lake[1]	-0.35	81
Color	-0.35	53
Orthophosphate concentration in lake	-0.33	47
Average land slope in basin	0.32	81
Bicarbonate concentration in lake	0.31	47
Magnesium concentration in lake	0.30	48
Head in Upper Floridan aquifer	0.30	81
Geographic group (see fig. 21)	0.26	81
Geomorphic division (Brooks, 1981)	0.23	81
Head difference between lake and Upper Floridan aquifer	0.22	81

[1]From land-use data.
[2]From National Wetlands Inventory data.

Ground-water inflow defined as a fraction of total inflows ($G_i/(G_i+P+S_i)$) was used as the dependent variable because residuals were more normally distributed than ground-water inflow in linear units or any of its transformations. Steady-state ground-water inflow results using both $\delta^{18}O$ and δD data, as well as transient results using δD data, were used in initial regression models. Models using the $\delta^{18}O$-derived ground-water inflow (table 7) had better r^2 values and lower standard errors than δD models and, thus, were used in the final models. The models using $\delta^{18}O$ data were better because of the better analytical precision of the $\delta^{18}O$ analysis of the lake water, which resulted in more accurate ground-water inflow estimates (see table 8). The explanatory variables in final regression models, however, were very similar for the $\delta^{18}O$ and δD models.

Model Results

Models using all lakes (n=81) and the subset of lakes with water-quality data (n=47) are summarized in table 10. Both these models had three variables in common: maximum lake depth, depth to the Upper Floridan aquifer, and the ratio of basin area to lake

Table 10. Multiple linear regression models to predict ground-water inflow

[Beta, standardized slope; r^2, coefficient of determination; std, standard; %, percent; n, number of observations; max, maximum; VIF, variance inflation factor; --, not applicable; UFA, Upper Floridan aquifer; ft, feet; <, less than; mg/L, milligrams per liter; N, nitrogen; µg/L, micrograms per liter; m, meter; btw, between; G_i, ground-water inflow as fraction of total inflows to lake; β_i, least-squares regression coefficient; x_i, selected independent variables. Model form is $G_i = \beta_0 + \beta_1 x_1 + ... + \beta_k x_k$, where k is the number of independent variables in the model]

	Term in model	Value	Standard error of value	Beta	Standard error of beta	Independent variable range
	Model using all lakes: r^2 = 0.71; adjusted r^2 = 0.69; std error of estimate = 12.9% of total inflows to lake; n = 81; max VIF = -1.7					
β_0	Intercept	-0.442	0.096	--	--	--
β_1	Basin area/lake surface area[1]	0.421	0.059	0.460	0.065	2.3 - 31.6
β_2	Depth to UFA[1]	0.210	0.049	0.348	0.081	19 - 470 ft
β_3	Maximum lake depth	4.41×10^{-3}	1.29×10^{-3}	0.267	0.078	2 - 78 ft
β_4	Fraction of wetlands in basin	-0.485	0.152	-0.237	0.074	0 - 47%
	Model using lakes with water-quality data: r^2 = 0.82; adjusted r^2 = 0.79; std error of estimate = 10.3% of total inflows to lake; n = 47; max VIF = -2.2					
β_0	Intercept	0.229	0.213	--	--	--
β_1	Depth to UFA[1]	0.138	0.053	0.258	0.099	19 - 470 ft
β_2	Nitrate concentration in lake[1]	0.0526	0.0220	0.230	0.097	< 0.001 - 3.2 mg/L as N
β_3	Sodium concentration in lake[1]	-0.282	0.103	-0.229	0.084	3.6 - 18.7 mg/L
β_4	Maximum lake depth	3.11×10^{-3}	1.25×10^{-3}	0.224	0.090	2 - 78 ft
β_5	Iron concentration in lake	-1.68×10^{-3}	6.09×10^{-4}	-0.216	0.078	5 - 148 µg/L
β_6	Basin area/lake surface area[1]	0.217	0.083	0.214	0.082	2.3 - 16.4
	Model using lakes in Lake Wales Ridge and Intraridge Valley with water-quality data: r^2 = 0.95; adjusted r^2 = 0.93; standard error of estimate = 3.6% of total inflows to lake; n = 15; max VIF = -1.5					
β_0	Intercept	0.906	0.109	--	--	--
β_1	Standard deviation of lake stage[1]	-1.01	0.126	-0.616	0.077	1.2 - 2.6 ft
β_2	Secchi depth	3.02×10^{-2}	6.61×10^{-3}	0.366	0.080	0.5 - 6.9 m
β_3	Sodium concentration in lake[1]	-0.396	0.107	-0.330	0.089	3.6 - 8.6 mg/L
β_4	Head difference btw lake and UFA	4.14×10^{-3}	1.11×10^{-3}	0.313	0.084	4.4 - 37.9 ft

[1]Variable log transformed in model.

surface area. The model using all lakes also included fraction of wetlands in the basin. The model for lakes with water-quality data contained nitrate, sodium, and iron concentrations and was a somewhat better model than the model using all lakes (adjusted $r^2 = 0.79$ compared to 0.69; standard error of 10.3 percent compared to 12.9 percent of total inflows). The relation between predicted and observed ground-water inflow was also better for the model using lakes with water-quality data (fig. 22a and b). Both models, however, had a bias toward low estimates for lakes with high ground-water inflow and high estimates for lakes with low ground-water inflow, illustrated by the deviation between the 1:1 line and regression line (fig. 22). Basin characteristics used in the final regression models are shown for each lake in appendix F.

Models were also constructed for the three geographic areas to see whether there were geographic differences in factors important in predicting ground-water inflow. Explanatory variables considered were those variables in the final models for the entire geographic area, as well as those basin characteristics that had statistically significant correlations with ground-water inflow for the geographic subset. Geographic-specific models typically were worse than the model using all the lakes together. The better model including all the lakes is the result of the wider range in ground-water inflow and explanatory variables, compared to the individual geographic areas. Models using lakes with water-quality data also were generally worse by geographic region than for the entire area considered together. The only exception was the model for lakes in the Lake Wales Ridge/Intraridge Valley with water-quality data. This model had an adjusted r^2 of 0.93 and a standard error of 3.6 percent of inflows (table 10) and successfully predicted ground-water inflow (fig. 22c). However, the model does contain a large number of explanatory variables (4) for the small number of data points (n = 15). One of the four variables (sodium concentration) is the same as in the water-quality-subset model for the entire study area. Three additional variables appear to be important in predicting ground-water inflow to lakes in the Lake Wales Ridge/Intraridge Valley: lake stage standard deviation, head difference between the lake and the Upper Floridan aquifer, and secchi depth (water clarity).

Predicted ground-water inflow from the regression models was used to group the lakes into the three previously defined categories based on the amount of ground-water inflow: low ground-water inflow (<25 percent of inflows), medium ground-water inflow (25-50 percent of inflows), or high ground-water inflow (>50 percent of inflows). For the model using all the lakes, 68 percent of the lakes grouped into the same ground-water inflow category as they did from the isotope mass-balance estimates. When the standard error of the predicted ground-water inflow was considered, 81 percent of the lakes grouped into the correct category. For the model using lakes with water-quality data, 85 percent of the lakes initially

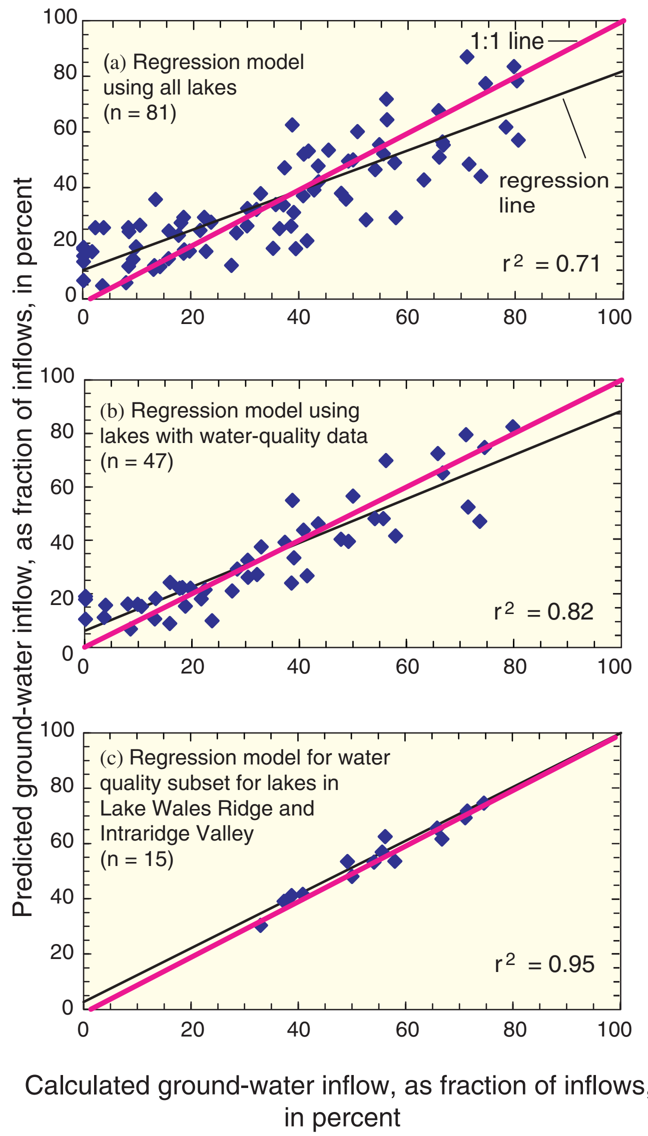


Figure 22. Relation between ground-water inflow predicted by the regression models and ground-water inflow calculated from the isotope mass-balance approach, (a) for the entire data set, (b) for the water-quality subset, and (c) for the water-quality subset for lakes in the Lake Wales Ridge and Intraridge Valley.

grouped into the correct category, and 91 percent grouped correctly when the model's standard error was considered. For the model using lakes in the Lake Wales Ridge/Intraridge Valley with water-quality data, 87 percent of the lakes initially grouped into the correct category, which increased to 93 percent when the model's standard error was considered.

Explanatory Variables Used to Predict Ground-Water Inflow

Each of the explanatory variables in the regression models represents a physical or chemical process that can explain its association with ground-water inflow. The variable with the most significant slope in the model using all the lakes was the ratio of the topographically defined lake basin to lake surface area (table 10). The positive slope indicates that as this ratio increases, predicted ground-water inflow also increases. For two lakes of similar size, the lake in the larger basin has a potentially larger area contributing ground-water inflow to the lake. In modeling hypothetical lake basins, Lee (2002) found that extending the basin divide farther from the lake caused an increase in the size of the ground-water contributing area, resulting in more ground-water inflow. It is important to note, however, that the topographically defined basin is not equivalent to the ground-water contributing area (watershed), particularly because of the irregular nature of Florida's karst terrain. Localized breaches in the confining unit can cause complex flow patterns near lakes (Sacks and others, 1992; Lee, 1996; Swancar and others, 2000). Nevertheless, topographic basin area and ground-water contributing area are probably highly correlated.

The depth to the Upper Floridan aquifer also had a positive slope in the regression equations. Thus, lakes with a greater depth to the Upper Floridan aquifer potentially receive more ground-water inflow compared to lakes with a shallow depth to the Upper Floridan aquifer. This variable includes the thickness of both the surficial aquifer system and the intermediate confining unit, and it can also be described as the thickness of unconsolidated "mantle" deposits overlying the Upper Floridan aquifer. In the study area, flow in the surficial aquifer system is generally downward to the Upper Floridan aquifer, unless intercepted by a lake or other surface-water body. Downward flow is controlled by the degree of confinement or separation between the aquifers, as well as the head difference between aquifers. The thicker the mantle deposits, the more hydraulically separated the lake is from the Upper Floridan aquifer, and, as a consequence, the influence of downward flow may become more subdued. For example, a lake with thicker mantle deposits may have more lateral ground-water flow toward the lake, which would otherwise have moved downward if mantle deposits were thinner. More water also is stored in thicker mantle deposits, compared to thinner deposits, which may affect the amount of ground water available to discharge into a lake. The intermediate confining unit typically is thicker where mantle deposits are thicker. Thickness of the intermediate confining unit was also significant in regression models, although those models had higher Mallow's Cp values than models containing depth to the Upper Floridan aquifer. Models including both variables were not considered because thickness of the intermediate confining unit is cross correlated with depth to the Upper Floridan aquifer.

Lake depth also had a positive slope in the regression equations, indicating that predicted ground-water inflow increases with increasing lake depth. A deep lake has the potential to intercept more, and possibly deeper, ground-water flow lines in the surficial aquifer system, increasing the amount of ground-water inflow. In contrast, a shallow lake would have less of an influence on ground-water flow patterns in the surficial aquifer system. In comparing two lakes that were the focus of detailed basin-scale studies, the deep lake (Lake Five-O) had considerably more ground-water inflow that the shallow lake (Lake Barco) (Grubbs, 1995; Lee, 1996). In modeling hypothetical basins, Lee (2002) found that ground-water inflow increased with lake depth, but only if mantle deposits increased concurrently.

The fraction of wetlands in the basin had a negative slope in the regression equation. It was the least important variable in the regression model using all the lakes, although it was still statistically significant ($p = 0.02$). The negative slope indicates that wetlands may decrease ground-water inflow to a lake by decreasing the size of the lake's potential ground-water contributing area. Wetlands in most of the lake basins were contiguous with the lake shore, although more lakes in the coastal lowlands had isolated wetlands in their basins compared to the central highlands. When the wetlands are contiguous with the lake shore, evapotranspiration can reduce the water table below the wetlands and cause lateral ground-water outflow from the lake, rather than ground-water inflow to the lake (Metz and Sacks, 2002). Isolated wetlands can diminish ground-water inflow to a lake by causing ground-water flow lines to deviate toward the wetlands, thereby reducing the area contributing ground water to the lake.

For the model using lakes with water-quality data, lake nitrate concentration had a positive slope in the equation. Nitrate concentration also had the highest individual correlation coefficient with ground-water inflow of all basin characteristics (table 9). The high positive correlation between ground-water inflow and nitrate concentration indicates that ground water is a significant source of nitrate to the study lakes. High nitrate concentrations (>10 mg/L as N) have been observed in ground-water inflow and in shallow ground water near lakes in central Florida, primarily associated with citrus agriculture (Fellows and Brezonik, 1981; German, 1996; Tihansky and Sacks, 1997). When ground-water inflow rates are high and phosphorous loading is limited, this nitrate can remain in the lake water column. This high nitrate loading has important implications because, if land use changes and phosphorus loading to the lake increases, this excess nitrate may then be available for algae, resulting in algal blooms and subsequent degradation in water quality (Kolasa and others, 2001).

Sodium concentration had a negative slope in the water-quality regression model. In freshwater lakes, sodium typically acts as a conservative ion without significant sources or sinks. Evaporation increases (concentrates) the sodium concentration in lake water. In contrast, ground-water inflow dilutes the sodium concentration in lake water because ground-water inflow is typically low in sodium. Thus, lakes with low ground-water inflow tend to have higher sodium concentrations than lakes with high ground-water inflow. The correlation between ground-water inflow and sodium concentration indicates that a simpler water-quality variable may be able to be used to estimate ground-water inflow, rather than the more complex isotope mass-balance approach. Of the major-ion tracers examined by Sacks and others (1998) to estimate ground-water inflow to Florida lakes, sodium was better than other potentially conservative tracers, such as chloride or magnesium, which had very large ranges in concentration in ground water near lakes. However, using sodium as a tracer is complicated because background sodium concentrations can be enriched in the surficial aquifer system, primarily due to septic-tank leachate (Alhajjar and others, 1990; Sacks and others, 1998).

Iron concentration also had a negative slope in the model for lakes with water-quality data. However, processes affecting iron concentrations are not the same as those affecting sodium. Iron does not act conservatively in the environment and is strongly influenced by the oxidation state of the water and biologically mediated reactions. Iron also can complex with organic acids in lake water (Wetzel, 1975). For the study lakes, iron concentrations (total) were significantly correlated with dissolved organic carbon concentrations (only available for selected lakes in Hillsborough and Pasco Counties), color, pH, and fraction of wetlands in the basin. Lakes with high amounts of ground-water inflow are flushed with ground water low in organic carbon. In contrast, reactions with organic matter become more important in a lake with low ground-water inflow, and iron complexes with the additional organic carbon in the lake water. An increase in biological reactions may also affect the oxidation state of the water, particularly near lake-bottom sediments, and a lower oxidation state favors the more mobile reduced iron (Fe^{2+}). Reactions involved in the iron cycle in lakes are undoubtedly complicated, but it is interesting to note its association with ground-water inflow.

Three other variables were included in the model for lakes in the Lake Wales Ridge/Intraridge Valley with water-quality data: head difference between the lake and Upper Floridan aquifer, lake-stage variability, and secchi depth (table 10). Head difference between the lake and Upper Floridan aquifer had a positive slope in the regression equation. This variable is another measure of the degree of hydraulic separation between the lake and the Upper Floridan aquifer, similar to depth to the Upper Floridan aquifer in the other models. Lakes with a greater degree of confinement have a greater head difference between the lake and the Upper Floridan aquifer than lakes that are in a more poorly confined setting.

Secchi depth (water clarity) also had a positive slope in the regression equation, which is the result of a subset of lakes in the Lake Wales Ridge that have high clarity and high ground-water inflow. High ground-water flow rates through excessively drained and well-leached sands of the Lake Wales Ridge can result in low phosphorous concentrations in shallow ground water. Low phosphorous concentrations in ground-water inflow can cause phosphorous to be a limiting lake nutrient, resulting in low algal productivity and high clarity. For lakes in the Lake Wales Ridge, water clarity was negatively correlated to total phosphorous and chlorophyll *a* concentrations and positively correlated to nitrate concentration.

Lake-stage variability (represented by the standard deviation of the lake-stage record) had a negative slope in the regression equation. This negative relation with ground-water inflow illustrates how ground-water inflow moderates lake stage in the Lake Wales Ridge. It is not clear why this variable is not significant in the other geographic areas. Lakes in the Lake Wales Ridge area have higher ground-water inflow (fig. 22), compared to other parts of the study area, and so the influence of ground water on lake stage may be more apparent. Ground-water inflow may also be more constant from year to year, making its influence discernible in the longer lake-stage record. Lakes in areas of lower topographic relief tend to be more affected by transient water-table mounds during and following periods of high recharge (Lee, 2002). As a result, ground-water inflow can increase significantly during wet years compared to that computed from the isotope mass-balance approach. Thus, lake stage variability in these areas may not be a good indicator of the "steady-state" ground-water inflow rate used here.

SUMMARY AND CONCLUSIONS

The isotope mass-balance approach was used to estimate ground-water inflow to 81 lakes in the central highlands and coastal lowlands of central Florida. The study area is characterized by a subtropical climate and numerous lakes in a mantled karst terrain. Values of δD and $\delta^{18}O$ were determined for rainwater, atmospheric moisture, ground-water and surface-water inflow, and lake water. Ground-water inflow was computed using both steady-state and transient formulations of the isotope mass-balance equation.

More detailed climatic, hydrologic, and isotopic data were collected from two study lakes, which were in different physiographic settings in the study area. Lake Starr is in the central highlands, and Halfmoon Lake is about 60 miles to the west in the coastal lowlands. For Lake Starr, ground-water inflow was independently computed from a water-budget study, and evaporation was computed using the energy-budget method. Climatic data (relative humidity, air temperature, and lake-surface temperature), which were necessary to compute the isotopic composition of lake evaporate, were not significantly different at the two lakes. In addition, the local meteoric water line and the average isotopic composition of rainwater, ground-water inflow, and atmospheric moisture were not significantly different at the two lakes. Isotopic data from all of the study lakes plotted on an evaporation trend line, which had a slope very similar to that theoretically computed for Lake Starr. This similarity suggests that data collected from the detailed study lakes can be extrapolated to the rest of the study area.

For Lake Starr, ground-water inflow computed from the isotope mass balance using δD data, steady-state assumptions, and a 4-year average for hydrologic variables was similar to the independent estimate of ground-water inflow averaged over 4 years. However, computed ground-water inflow was considerably higher than water-budget results using $\delta^{18}O$ data and sampled atmospheric moisture. Ground-water inflow from $\delta^{18}O$ data was also computed using a value for the isotopic composition of atmospheric moisture (δ_a) that was back-calculated from the independent estimate of ground-water inflow from the Lake Starr water budget. Ground-water inflow for the rest of the study lakes was computed using sampled δ_a for δD data and using back-calculated δ_a for $\delta^{18}O$ data.

Ground-water inflow ranged from 0 to more than 250 in/yr (or 0 to more than 80 percent of total inflows) for the 81 study lakes. For the various methods, median ground-water inflow was between 30 and 37 in/yr (or between 36 and 41 percent of total inflows). Steady-state and transient results were very similar. Of the hydrologic variables in the calculations, ground-water inflow results were most sensitive to changes in lake evaporation. Of the isotopic variables, results were most sensitive to the isotopic composition of lake water. Sensitivity using δD data was about twice that using $\delta^{18}O$ data because of the greater analytical uncertainty for δD measurements. The transient formulation of the isotope mass-balance equation is extremely sensitive to lake isotopic composition. Ground-water inflow was also very sensitive to other variables used to compute the isotopic composition of lake evaporate, including δ_a and climatic variables of relative humidity, air temperature, and lake-surface temperature. The fractional uncertainty in ground-water inflow results is considerably less for lakes with higher ground-water inflow than for lakes with lower ground-water inflow.

Because of large uncertainty in results, ground-water inflow calculated using the isotope mass-balance approach provides information on the general magnitude of ground-water inflow, rather than precise quantification of this inflow. The lakes were grouped into three categories based on their range of ground-water inflow: low (less than 25 percent of total inflows), medium (25-50 percent of inflows), and high (greater than 50 percent of inflows). The majority of lakes in the coastal lowlands had low ground-water inflow, whereas the majority of highland lakes in the Lake Wales Ridge and Intraridge Valley had high ground-water inflow. The majority of highland lakes in other ridge and upland areas grouped in the medium ground-water inflow range.

Lake and basin characteristics were used in multiple linear regression models to predict ground-water inflow. Explanatory variables in the final regression model using all the lakes included: ratio of basin area to lake surface area, depth to the Upper Floridan aquifer, maximum lake depth, and fraction of wetlands in the basin. For a model using lakes with water-quality data, nitrate, sodium, and iron concentrations were also included in the model, but fraction of wetlands was not included. Geographically specific models were generally poorer than regression models for the entire geographic area. The only exception is a model for lakes in the Lake Wales Ridge/Intraridge Valley with water-quality data; this model included lake-stage variability, secchi depth, sodium concentration, and head difference between the lake and the Upper Floridan aquifer. Ground-water inflow results predicted from regression models were grouped into the three ground-water inflow categories, and 81 to 93 percent of the lakes grouped into the previously determined category (after considering the standard error of the estimate). Regression models should not be used for precise quantification of ground-water inflow for individual lakes; rather, model results can be used to group lakes into one of the three ground-water inflow categories.

The isotope mass-balance approach was particularly useful for comparing the range of ground-water inflow to numerous lakes in Florida. Although the uncertainty in exact values is high, no better method exists for estimating ground-water inflow to a large number of lakes. Results from this study also indicated how more simplified regression models, based on basin characteristics, could be used to estimate ground-water inflow. In addition, these models helped to identify which basin characteristics are important in controlling ground-water inflow to Florida lakes. Models were improved by including lake water-quality data, illustrating the link between ground-water geochemistry and lake chemistry. In order to use the isotope mass-balance approach for multiple lakes in Florida, the lakes should be sampled during the same time period; detailed isotopic, hydrologic, and climatic data should also be collected over this same time period. Isotopic data for Florida lakes can change over time, both seasonally and interannually, because of differences in net precipitation and ground-water exchange. Thus, ground-water inflow from the current study does not necessarily indicate historical inflow, particularly if lowered ground-water levels have caused reductions in ground-water inflow to lakes. The isotope mass-balance approach was most successful for lakes in the central highlands, where lakes typically are deeper, undergo less isotopic variability, and ground-water inflow is higher than lakes in the coastal lowlands. Results from this study illustrate the large range of ground-water inflow to Florida lakes and underscore how important ground water is in the water budget of many of Florida's lakes.

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APPENDIXES

Appendix A. Isotopic composition of rainwater at Lake Starr and Halfmoon Lake

[Units in δ per mil, unless otherwise noted; δD, delta deuterium, $\delta^{18}O$, delta oxygen-18; n/a, not available (insufficient volume)]

Begin date	End date	δD	$\delta^{18}O$	Rainfall total (inches)
Lake Starr rainwater:				
1/21/99	2/1/99	-13.5	-3.08	2.43
2/1/99	3/1/99	6.9	-1.20	0.44
3/1/99	4/1/99	-3.5	-1.74	0.68
4/1/99	5/1/99	-11.8	-3.21	1.74
5/1/99	5/18/99	-12.2	-2.97	2.35
5/18/99	6/1/99	-2.5	-1.62	1.61
6/1/99	7/1/99	-29.0	-5.11	13.38
7/1/99	7/20/99	-8.4	-2.10	1.20
7/20/99	8/17/99	2.5	-1.20	5.34
8/17/99	9/28/99	-25.3	-4.43	3.87
9/28/99	10/20/99	-19.4	-3.83	3.90
10/20/99	11/23/99	-26.3	-4.93	3.36
11/23/99	12/15/99	7.5	-0.45	0.14
12/15/99	1/11/00	-15.8	-4.72	2.60
1/11/00	2/23/00	-10.8	-3.14	1.15
2/23/00	3/22/00	8.4	-0.58	0.56
3/22/00	4/18/00	-3.4	-1.85	1.34
4/18/00	5/10/00	n/a	n/a	0.02
5/10/00	6/13/00	-14.9	-3.30	1.98
6/13/00	7/11/00	-7.5	-2.23	5.30

Begin date	End date	δD	$\delta^{18}O$	Rainfall total (inches)
Halfmoon Lake rainwater:				
1/8/99	2/1/99	-17.4	-3.79	2.11
2/1/99	3/1/99	2.3	-1.44	0.08
3/1/99	4/1/99	-0.1	-2.00	1.66
4/1/99	5/1/99	-15.9	-4.07	1.00
5/1/99	5/17/99	-4.8	-1.59	1.13
5/17/99	6/1/99	-3.0	-2.00	2.65
6/1/99	7/1/99	-21.0	-4.13	11.05
7/1/99	7/19/99	-12.0	-2.77	4.58
7/19/99	8/16/99	-2.5	-1.94	3.82
8/16/99	9/29/99	-29.4	-5.24	12.58
9/29/99	10/19/99	-26.4	-4.86	2.55
10/19/99	11/22/99	-43.3	-6.94	1.51
11/22/99	12/16/99	8.4	-0.70	0.35
12/16/99	1/12/00	-10.2	-3.11	1.32
1/12/00	2/24/00	-15.6	-3.48	1.41
2/24/00	3/23/00	10.4	-0.15	0.13
3/23/00	4/19/00	0.6	-1.00	0.61
4/19/00	5/11/00	9.7	-0.41	0.22
5/11/00	6/16/00	-20.2	-3.60	2.69
6/16/00	7/13/00	-5.6	-2.21	11.16

Appendix B. Isotopic composition of atmospheric moisture at Lake Starr and Halfmoon Lake --(Continued)

[δD, delta deuterium; $\delta^{18}O$, delta oxygen-18; ºC, degrees Celsius; --, not applicable]

Date	δD (per mil)	$\delta^{18}O$ (per mil)	Air temperature (ºC)[1]	Relative humidity (percent)[1]
Lake Starr atmospheric moisture:				
5/18/99	-79.9	-12.07	28.8	59.5
6/10/99	-84.5	-12.77	30.4	70.5
7/20/99	-77.0	-12.02	33.3	60.0
8/17/99	-71.8	-10.91	30.7	71.5
8/17/99[2]	-68.6	-10.87	30.7	71.5
9/28/99	-76.7	-11.42	30.4	68.3
10/20/99	-92.0	-13.18	29.3	75.0
12/15/99	-86.4	-14.44	20.4	59.5
1/11/00	-80.4	-12.08	20.6	60.0
1/11/00[3]	-80.3	-12.26	23.3	53.0
3/22/00	-71.3	-11.20	24.8	59.0
4/18/00	-74.7	-11.21	26.8	57.0

Date	δD (per mil)	$\delta^{18}O$ (per mil)	Air temperature (ºC)[1]	Relative humidity (percent)[1]
Halfmoon Lake atmospheric moisture:				
5/17/99	-89.1	-13.91	26.3	57.5
6/9/99	-89.8	-12.58	28.4	70.0
7/19/99	-75.7	-11.76	31.3	63.5
8/16/99	-67.7	-10.99	31.7	63.5
9/29/99	-74.5	-11.14	28.8	74.1
10/19/99	-95.4	-14.59	28.1	66.8
11/22/99	-83.0	-12.66	22.8	73.0
12/1/99	-92.5	-16.18	13.1	48.0
12/16/99	-83.0	-13.32	16.8	69.5
1/12/00	-72.5	-12.34	19.6	64.0
1/12/00[4]	-75.9	-12.42	21.2	61.3
2/24/00	-80.3	-12.66	21.9	68.0
3/23/00	-112.3	-16.78	22.4	50.0
4/19/00	-106.8	-16.30	24.2	48.0
4/19/00[2]	-107.4	-16.38	24.2	48.0
Average of monthly data from both lakes:				
--	-83.0	-12.80	25.8	64.1

[1]Average of data from beginning and ending of sampling period, typically 2-3 hours.
[2]Duplicate.
[3]Sample from north shore.
[4]Sample from southeast shore.

Appendix C. Isotopic composition of ground water at selected study lakes

[Samples from minipiezometer unless specified otherwise; δD, delta deuterium, $\delta^{18}O$, delta oxygen-18; S, south; E, east; W, west; N, north]

Lake name	General location of sample	Sampling date	δD (per mil)	$\delta^{18}O$ (per mil)	Head difference (feet)[1]	Distance onshore (+) or offshore (-) (feet)	Depth below land surface or lake bottom (feet)
Lakes in the coastal lowlands (fall 1999 sampling):							
Calm	SE shore	10/6/99	-3.9	-0.90	0.39	5.5	2.1
Calm	SE shore	10/6/99	-5.8	-1.18	0.45	-6.5	2.4
Crews	SW shore	10/5/99	13.3	2.80	0.09	13.0	1.1
Crews	SW shore	10/5/99	7.1	1.75	0.005	-1.0	1.1
Deer	NE shore	10/6/99	-18.9	-3.84	0.32	6.0	1.7
Deer	NE shore	10/6/99	-16.5	-3.48	0.18	-7.5	1.6
Halfmoon[2]	SE shore	10/4/99	-20.7	-3.87	1.28	30.0	[3]12
Halfmoon[2]	SE shore	10/4/99	-17.7	-3.43	0.18	5.7	[3]5
Halfmoon	SE shore	10/4/99	-19.2	-3.40	0.09	-3.0	2.0
Halfmoon	SE shore	10/4/99	-17.0	-3.31	0.33	-7.5	2.5
Halfmoon	SE shore	10/4/99	-17.9	-3.59	0.20	10.2	2.4
Jessamine	S shore	10/6/99	-20.1	-3.81	0.14	6.5	2.7
Jessamine	S shore	10/6/99	-18.9	-3.78	0.03	-1.5	2.0
Moon	E shore	10/5/99	-29.0	-4.97	0.02	-1.0	1.6
Moon	E shore	10/5/99	-24.0	-4.23	0.12	4.0	1.7
Starvation	E shore	10/4/99	-12.6	-3.05	0.09	8.0	0.9
Starvation	E shore	10/4/99	1.0	0.40	0.02	-8.0	0.7
Lakes in the coastal lowlands (spring 2000 sampling):							
Calm	SE shore	4/4/00	-13.7	-2.82	0.140	8.0	2.0
Calm	SE shore	4/4/00	-8.3	-2.31	0.025	0.0	1.6
Deer	NE shore	4/3/00	-16.4	-3.53	0.080	8.0	1.2
Deer	NE shore	4/3/00	-19.0	-3.88	0.080	-6.0	1.2
Halfmoon	SE shore	4/4/00	-17.3	-3.54	0.170	4.0	2.2
Halfmoon	SE shore	4/4/00	-17.0	-3.51	0.260	-3.5	3.1
Jessamine	S shore	4/6/00	-18.3	-3.64	0.110	.5	2.5
Jessamine	S shore	4/6/00	-4.4	-0.60	0.035	-4.5	2.3
Starvation	E shore	4/6/00	-11.5	-2.11	0.060	21.0	1.5
Lakes in the central highlands (fall 1999 sampling):							
Crooked	NW shore	10/12/99	1.2	-0.19	0.195	6.5	1.7
Crooked	NW shore	10/12/99	-11.3	-2.73	0.21	6.5	1.7
Eagle	E shore	10/7/99	-17.3	-3.67	0.36	10.0	1.3
Eagle	E shore	10/7/99	-21.3	-4.00	0.32	-6.0	2.1
Helene	S shore	10/12/99	-32.9	-5.46	0.195	7.3	1.2

Appendix C. Isotopic composition of ground water at selected study lakes --(Continued)

[Samples from minipiezometer unless specified otherwise; δD, delta deuterium, $\delta^{18}O$, delta oxygen-18; S, south; E, east; W, west; N, north]

Lake name	General location of sample	Sampling date	δD (per mil)	$\delta^{18}O$ (per mil)	Head difference (feet)[1]	Distance onshore (+) or offshore (-) (feet)	Depth below land surface or lake bottom (feet)
Helene	S shore	10/12/99	-19.0	-3.79	0.12	-4.5	1.9
Menzie	W shore	10/12/99	-22.7	-3.85	0.14	4.0	1.3
Menzie	W shore	10/12/99	-26.9	-4.58	0.07	-4.5	1.5
Saint Anne	NE shore	10/7/99	-12.1	-2.49	0.17	6.5	2.7
Saint Anne	NE shore	10/7/99	-23.0	-4.19	0.02	-9.0	2.1
Starr[2]	N shore	10/7/99	-25.8	-4.82	0.34	80.0	[3]5
Starr[2]	N shore	10/7/99	-18.0	-3.73	0.50	150.0	[3]7
Starr	N shore	10/7/99	-23.8	-4.30	0.11	22.0	2.6
Starr	W shore	10/7/99	-21.3	-4.01	0.05	13.5	1.3
Starr	W shore	10/7/99	-13.5	-2.66	0.02	-9.0	2.1
Verona	W shore	10/12/99	-24.7	-4.35	0.11	5.0	1.2
Verona	W shore	10/12/99	-18.0	-3.50	0.035	-3.5	1.2
Lakes in the central highlands (spring 2000 sampling):							
Crooked	NW shore	3/27/00	-8.0	-1.44	0.140	6.0	2.1
Eagle	E shore	3/28/00	-19.3	-3.88	0.320	5.0	2.2
Eagle	E shore	3/28/00	-18.1	-3.85	0.100	-6.0	1.7
Menzie	W shore	3/28/00	-18.8	-3.83	0.135	11.0	2.7
Menzie	W shore	3/28/00	-20.4	-4.05	0.015	-6.0	3.0
Saint Anne	NE shore	3/27/00	-16.6	-3.21	0.280	9.1	2.5
Saint Anne	NE shore	3/27/00	-16.5	-3.55	0.055	-6.0	2.2
Starr	N shore	3/31/00	-19.0	-3.79	0.130	20.0	1.8
Starr	N shore	3/31/00	-14.1	-2.83	0.015	-8.5	3.1
Starr	N shore	3/31/00	-14.2	-2.82	0.060	11.5	2.2
Starr	W shore	3/31/00	-14.5	-3.52	0.310	6.0	2.5
Verona	W shore	3/27/00	-18.8	-3.97	0.060	4.0	1.5
Verona	W shore	3/27/00	-23.6	-4.17	0.030	-3.4	1.6

[1]Head difference between ground water and lake; ground-water head greater than lake.
[2]Shallow monitoring well near lake.

Appendix D. Isotopic composition of surface-water inflow

[Latitude and longitude in degrees (^{o}), minutes ('), and seconds ("); δD, delta deuterium; $\delta^{18}O$, delta oxygen-18; ft^3/s, cubic feet per second; NW, northwest; W, west; SW, southwest]

Site	Latitude	Longitude	Sampling date	δD (per mil)	$\delta^{18}O$ (per mil)	Discharge (ft^3/s)
Crooked Lake inflow:						
NW channel	27° 49' 20"	81° 35' 42"	8/13/99	-13.0	-2.43	0.47
			10/27/99	-14.5	-3.09	0.35
W channel	27° 48' 32"	81° 35' 50"	8/13/99	4.3	0.12	3.6
			10/27/99	-7.4	-1.17	1.3
SW channel	27° 47' 42"	81° 35' 50"	8/13/99	2.8	-0.26	4.2
			10/27/99	-9.2	-1.77	13
Clinch Lake inflow:						
NW channel	27° 45' 06"	81° 33' 33"	10/26/99	-15.4	-3.65	0.33
			3/1/00	-17.3	-3.45	0.14
W channel	27° 44' 40"	81° 33' 34"	8/13/99	-14.3	-3.02	0.38
			10/26/99	-16.2	-3.24	0.35
SW channel	27° 44' 08"	81° 33' 52"	8/13/99	-12.2	-2.10	0.87
			10/26/99	-15.2	-2.98	0.87
			3/1/00	-12.5	-2.26	0.041
Lake Lotela inflow:						
W channel	27° 34' 35"	81° 29' 39"	8/13/99	3.4	0.86	0.10
			10/26/99	2.6	0.37	4.3

Appendix E. Isotopic composition of lake water from summer 1999 and winter 2000 samplings

[δD, delta deuterium, $\delta^{18}O$, delta oxygen-18; n/a, not available; SW, southwest; dup, duplicate sample]

Map reference number	Lake name	Summer 1999			Winter 1999		
		Sampling Date	δD (per mil)	$\delta^{18}O$ (per mil)	Sampling Date	δD (per mil)	$\delta^{18}O$ (per mil)
Lakes in the coastal lowlands:							
1	Alice	7/1/99	13.1	2.80	1/18/00	12.7	2.80
2	Allen	7/6/99	8.3	2.08	2/4/00	10.6	2.32
3	Bird	7/6/99	8.6	2.01	1/19/00	11.0	2.90
4	Boat (SW lobe)	7/8/99	6.4	1.54	1/19/00	6.9	1.71
	Boat (north lobe)	7/8/99	5.8	1.58	1/19/00	6.8	1.66
5	Calm	8/24/99	13.0	2.75	1/27/00	12.1	2.52
6	Carroll	8/31/99	8.8	2.01	1/26/00	7.9	2.14
7	Deer	7/7/99	10.2	2.41	1/19/00	9.7	2.40
8	Egypt	7/8/99	3.7	1.09	1/19/00	6.4	1.42
9	George	7/8/99	1.5	0.92	1/19/00	2.7	0.97
10	Halfmoon (north lobe)	7/19/99	13.0	2.90	1/12/00	12.6	2.90
	Halfmoon (south lobe)	n/a	n/a	n/a	1/12/00	13.1	2.94
11	Hobbs (west lobe)	7/7/99	10.4	2.61	1/19/00	12.1	2.62
	Hobbs (east lobe)	7/7/99	10.3	2.63	n/a	n/a	n/a
12	Hog Island	7/7/99	8.8	2.14	2/4/00	9.7	2.02
13	Juanita	7/1/99	13.2	2.84	1/18/00	11.9	2.74
14	LeClare	7/8/99	13.0	2.90	1/18/00	12.2	3.00
15	Merrywater	7/6/99	8.2	2.46	1/19/00	13.6	2.87
16	Mound	7/6/99	6.9	1.76	1/18/00	8.0	2.03
17	Osceola	7/6/99	12.2	2.71	1/18/00	12.8	2.74
	Osceola (dup)	n/a	n/a	n/a	1/18/00	12.2	2.74
18	Raleigh	7/1/99	10.4	2.47	1/18/00	12.1	2.79
19	Rogers	7/1/99	15.4	3.07	1/18/00	16.0	3.27
20	Starvation	7/6/99	9.7	2.66	1/19/00	9.0	2.03
21	Taylor	8/24/99	11.6	2.63	1/27/00	10.6	2.49
22	Van Dyke	7/22/99	7.4	1.76	1/18/00	4.5	0.85
23	Big Fish	7/27/99	16.6	3.88	1/25/00	16.4	3.35
24	Black	9/9/99	8.5	2.69	1/27/00	7.7	2.67
25	Camp	7/7/99	7.5	2.51	1/25/00	8.5	2.70
26	Crews	8/24/99	12.1	2.54	1/31/00	12.1	2.50
27	Curve	7/22/99	14.5	3.41	1/24/00	12.4	3.22
	Curve (dup)	7/22/99	15.7	3.42	n/a	n/a	n/a
28	Gooseneck	8/17/99	8.9	2.19	2/1/00	6.1	1.42
29	King	7/22/99	12.7	2.83	2/1/00	14.4	3.00
30	Linda (SW lobe)	7/7/99	11.6	2.71	2/4/00	8.7	2.25
	Linda (center lobe)	n/a	n/a	n/a	2/4/00	7.0	2.36
31	Moon	7/13/99	14.8	2.93	1/24/00	13.8	3.09
32	Pierce (center lobe)	7/13/99	18.7	4.13	1/25/00	17.0	3.99
	Pierce (east lobe)	7/13/99	19.9	4.28	n/a	n/a	n/a
33	Thomas (Pasco)	7/22/99	13.5	2.94	1/25/00	15.7	3.10
34	Wistaria	8/17/99	11.8	3.33	2/1/00	12.4	3.20
Lakes in the central highlands (Lake Wales Ridge and Intraridge Valley):							
35	Angelo	7/19/99	11.3	2.27	1/31/00	7.1	1.77
36	Chilton	7/19/99	5.5	1.46	1/31/00	4.5	1.21
37	Denton	7/19/99	4.2	0.65	1/31/00	3.1	0.49
38	Dinner (Highlands)	7/19/99	9.0	1.56	1/31/00	7.5	1.14
39	Isis	n/a	n/a	n/a	1/31/00	-5.8	-1.00
40	Lotela	7/19/99	11.6	1.89	1/31/00	10.8	1.82
41	Olivia	n/a	n/a	n/a	1/31/00	0.9	0.25
42	Pioneer	7/19/99	9.9	1.94	1/31/00	8.8	1.52

Appendix E. Isotopic composition of lake water from summer 1999 and winter 2000 samplings --(Continued)

[δD, delta deuterium, $\delta^{18}O$, delta oxygen-18; n/a, not available; SW, southwest; dup, duplicate sample]

Map reference number	Lake name	Summer 1999			Winter 1999		
		Sampling Date	δD (per mil)	$\delta^{18}O$ (per mil)	Sampling Date	δD (per mil)	$\delta^{18}O$ (per mil)
43	Tulane	7/19/99	0.6	0.03	1/31/00	-0.2	-0.28
44	Verona (1.5 ft)	7/19/99	-0.8	-0.29	1/31/00	-2.2	-0.74
	Verona (40-45 ft)	7/19/99	-4.6	-0.80	1/31/00	-3.0	-0.75
45	Viola	7/19/99	6.8	1.40	1/31/00	5.4	0.97
46	Aurora	7/20/99	4.3	1.03	2/1/00	1.7	0.76
47	Blue (south lobe)	7/21/99	6.5	1.30	2/1/00	6.1	1.28
48	Blue (north lobe)	7/21/99	0.5	0.29	2/1/00	0.3	0.29
49	Crystal	7/15/99	10.8	1.99	2/7/00	12.6	2.03
50	Dinner (Polk)	7/15/99	6.3	1.60	2/2/00	6.3	1.32
51	Hickory	7/20/99	7.0	1.73	1/31/00	6.2	1.22
52	Josephine	7/20/99	-6.0	-0.72	2/1/00	-5.9	-1.01
53	Little Aurora	7/20/99	-3.6	0.04	2/1/00	-3.1	-0.40
	Little Aurora (dup)	7/20/99	-2.9	-0.03	2/1/00	-2.0	-0.45
54	Mabel	7/15/99	5.2	1.49	2/2/00	8.3	1.39
	Mabel (dup)	7/15/99	5.9	1.52	n/a	n/a	n/a
55	Menzie (1.5 ft)	7/15/99	8.8	2.18	2/7/00	10.9	2.30
	Menzie (4 ft)	n/a	n/a	n/a	2/7/00	10.6	2.33
56	Saddlebag	7/20/99	7.7	1.76	2/1/00	6.2	1.28
57	Saint Anne	7/20/99	-1.7	-0.28	2/1/00	-4.7	-0.57
58	Silver	7/20/99	10.8	1.55	1/31/00	9.7	1.43
59	Starr	7/20/99	10.4	2.39	1/11/00	11.2	2.25
	Starr (dup)	n/a	n/a	n/a	1/11/00	11.8	2.29
60	Wales	7/15/99	9.9	1.83	1/31/00	12.8	2.04
61	Warren	7/21/99	4.6	2.01	2/1/00	6.0	2.05
Lakes in the central highlands (other ridge and upland areas):							
62	Clinch	7/15/99	11.0	1.92	1/31/00	10.0	1.85
63	Crooked	7/15/99	[1]13.4	[1]2.28	1/27/00	11.6	1.98
64	Eagle	7/26/99	12.9	2.45	1/26/00	14.3	2.27
65	Grassy	7/14/99	15.5	3.26	1/26/00	14.9	2.78
	Grassy (near shore)	n/a	n/a	n/a	1/26/00	15.8	2.66
66	Helene	7/14/99	12.5	2.77	1/26/00	14.7	2.96
67	Henry	7/20/99	9.6	2.00	2/1/00	10.4	2.02
68	Little Van	7/14/99	1.4	0.86	1/26/00	2.5	0.44
69	Lizzie	7/20/99	10.9	2.34	2/1/00	8.7	1.59
70	Lucerne	7/14/99	15.1	3.14	1/26/00	16.3	3.28
71	McLeod	7/15/99	11.3	2.07	2/2/00	13.2	2.13
	McLeod (near shore)	n/a	n/a	n/a	2/2/00	14.7	1.92
72	Medora	7/14/99	14.0	2.76	1/26/00	14.9	2.98
73	Polecat	7/20/99	2.6	1.08	2/2/00	2.2	0.18
74	Sara	7/15/99	11.1	2.50	2/7/00	12.6	2.51
75	Tennessee	7/28/99	9.5	2.05	2/1/00	9.6	1.77
76	Thomas (Polk)	7/21/99	11.2	2.32	1/26/00	9.4	2.15
77	Walker	7/20/99	9.6	1.78	2/2/00	6.8	1.40
78	Iola	7/13/99	10.0	2.20	1/25/00	12.5	2.54
79	Jessamine	8/24/99	-1.5	-0.05	1/31/00	-1.1	0.03
80	Pasadena	7/13/99	15.9	2.94	1/25/00	16.1	3.15
81	Spring	7/13/99	8.5	1.82	1/25/00	7.9	1.61

[1]Average of large lobe (δD = 13.7, $\delta^{18}O$ = 2.31) and small southern lobe (Little Crooked) (δD = 13.1, $\delta^{18}O$ = 2.25).

Appendix F. Basin characteristics used in final multiple regression models for the entire study area

[UFA, Upper Floridan aquifer; N, nitrogen; mg/L, milligrams per liter; µg/L, micrograms per liter; n/a, not available; <, less than]

Map reference number	Lake name	Ratio of basin area to lake surface area (unitless)	Maximum lake depth (ft)	Depth to UFA (feet below lake stage)	Wetlands in basin (percent of area)	Lake water: Nitrate (mg/L as N)	Lake water: Sodium (mg/L)	Lake water: Iron (µg/L)
Lakes in the coastal lowlands:								
1	Alice	3.1	24	48	20	0.004	11.5	20
2	Allen	5.9	25	49	13	0.003	17.6	72
3	Bird	4.8	14	37	11	n/a	n/a	n/a
4	Boat	3.6	17	34	1	n/a	n/a	n/a
5	Calm	2.9	25	57	1	0.019	11.1	25
6	Carroll	3.4	18	34	1	0.010	12.7	25
7	Deer	3.5	31	54	14	0.29	17.0	50
8	Egypt	4.9	32	45	1	0.017	9.6	44
9	George	3.8	24	44	0	n/a	n/a	n/a
10	Halfmoon	4.3	25	41	11	0.030	17.7	46
11	Hobbs	3.0	18	63	24	n/a	n/a	n/a
12	Hog Island	10.9	19	55	46	0.013	7.5	69
13	Juanita	4.3	20	38	18	0.008	15.3	25
14	LeClare	3.9	16	49	14	n/a	n/a	n/a
15	Merrywater	5.8	8	53	36	0.004	9.2	114
16	Mound	3.1	24	49	14	n/a	n/a	n/a
17	Osceola	3.1	17	44	22	n/a	n/a	n/a
18	Raleigh	3.0	17	35	19	< 0.001	9.7	34
19	Rogers	2.7	16	34	24	< 0.001	6.1	50
20	Starvation	6.8	11	38	33	0.002	3.6	87
21	Taylor	4.9	20	46	7	0.010	13.4	40
22	Van Dyke	16.8	4	43	47	n/a	n/a	n/a
23	Big Fish	14.4	8	32	50	n/a	n/a	n/a
24	Black	8.3	11	37	18	0.010	5.3	75
25	Camp	26.7	7	38	46	n/a	n/a	n/a
26	Crews	7.7	2	29	26	0.014	5.4	63
27	Curve	2.4	16	54	0	n/a	n/a	n/a
28	Gooseneck	4.8	13	41	12	0.012	6.0	30
29	King	4.2	9	50	37	0.023	9.1	65
30	Linda	3.7	11	43	18	n/a	n/a	n/a
31	Moon	3.9	21	48	37	0.001	13.2	45
32	Pierce	3.3	15	59	28	< 0.001	4.5	58
33	Thomas (Pasco)	2.5	11	52	28	0.043	18.7	27
34	Wistaria	2.9	14	51	22	0.010	9.0	25
Lakes in the central highlands (Lake Wales Ridge and Intraridge Valley):								
35	Angelo	11.7	15	416	0	0.35	7.0	35
36	Chilton	6.7	21	339	0	0.015	5.4	23
37	Denton	7.1	48	420	0	2.80	5.1	14
38	Dinner (Highlands)	4.0	31	470	2	0.107	8.3	70
39	Isis	12.2	63	379	0	3.20	5.1	10
40	Lotela	2.3	27	415	6	0.074	8.6	33
41	Olivia	3.3	46	353	0	0.049	6.0	15
42	Pioneer	4.7	35	370	1	n/a	n/a	n/a
43	Tulane	6.1	78	394	0	0.25	5.2	12

Appendix F. Basin characteristics used in final multiple regression models for the entire study area --(Continued)

[UFA, Upper Floridan aquifer; N, nitrogen; mg/L, milligrams per liter; µg/L, micrograms per liter; n/a, not available; <, less than]

Map reference number	Lake name	Ratio of basin area to lake surface area (unitless)	Maximum lake depth (ft)	Depth to UFA (feet below lake stage)	Wetlands in basin (percent of area)	Lake water		
						Nitrate (mg/L as N)	Sodium (mg/L)	Iron (µg/L)
44	Verona	13.2	68	392	1	0.066	3.6	17
45	Viola	13.3	33	376	0	0.45	5.0	9
46	Aurora	4.8	37	245	1	2.66	n/a	n/a
47	Blue (south lobe)	3.5	58	186	1	0.40	7.4	5
48	Blue (north lobe)	5.5	51	182	1	2.32	5.2	15
49	Crystal	6.0	20	120	18	0.001	n/a	n/a
50	Dinner (Polk)	14.3	13	140	0	n/a	n/a	n/a
51	Hickory	2.8	13	345	13	0.61	8.2	19
52	Josephine	31.6	28	172	0	n/a	n/a	n/a
53	Little Aurora	9.4	34	235	0	7.55	n/a	n/a
54	Mabel	8.8	17	135	8	0.14	7.0	10
55	Menzie	8.1	21	99	5	0.030	n/a	n/a
56	Saddlebag	8.9	46	242	3	n/a	n/a	n/a
57	Saint Anne	8.1	33	202	0	5.21	n/a	n/a
58	Silver	4.5	28	342	0	n/a	n/a	n/a
59	Starr	6.3	31	132	6	0.008	8.5	15
60	Wales	4.5	19	156	8	0.20	5.7	25
61	Warren	9.3	20	166	0	n/a	n/a	n/a
Lakes in the central highlands (other ridge and upland areas):								
62	Clinch	5.6	29	304	10	0.21	9.8	28
63	Crooked	3.7	25	207	20	0.20	7.7	148
64	Eagle	3.6	27	178	6	0.010	14.2	25
65	Grassy	3.2	14	100	22	n/a	n/a	n/a
66	Helene	3.5	27	93	13	n/a	n/a	n/a
67	Henry	9.3	21	247	5	n/a	n/a	n/a
68	Little Van	11.1	9	107	11	n/a	n/a	n/a
69	Lizzie	5.1	23	215	12	n/a	n/a	n/a
70	Lucerne	6.0	16	105	8	0.010	17.5	n/a
71	McLeod	3.7	27	189	16	0.005	8.8	15
72	Medora	2.5	23	146	0	n/a	n/a	n/a
73	Polecat	9.5	25	201	7	n/a	n/a	n/a
74	Sara	4.5	15	119	9	n/a	n/a	n/a
75	Tennessee	3.4	17	91	2	0.010	9.4	28
76	Thomas (Polk)	4.0	25	159	8	0.026	n/a	n/a
77	Walker	10.0	31	202	0	n/a	n/a	n/a
78	Iola	3.2	41	42	0	0.089	8.8	26
79	Jessamine	16.4	26	26	5	0.20	5.0	38
80	Pasadena	13.9	12	19	18	< 0.001	6.2	85
81	Spring	4.0	39	101	4	0.025	6.3	25

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