CHAPTER 4. GROUNDWATER HYDROLOGY

by

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ACRONYMS, ABBREVIATIONS, AND CONVERSION FACTORS

Acronym, Abbreviation or	Definition
Ca	Calcium
Cl	Chloride
CO_3	Carbonate
CR	County Road
CV	Coefficient of variation
ECF model	East-central Florida model
EFDC	Environmental Fluid Dynamics Code surface water model
EFH	Equivalent freshwater head
HCO_3	Bicarbonate
HSPF	Hydraulic Simulation Program–Fortran
K	Potassium
LFA	Lower Floridan aquifer
MFLs	Minimum flows and levels
Mg	Magnesium
MODFLOW	USGS modular three-dimensional finite groundwater flow
MRE	Median relative error
Na	Sodium
NCF model	North-central Florida model
NGVD29	National Geodetic Vertical Datum of 1929
NRC	National Research Council
PSS78	The Practical Salinity Scale 1978
RSE	Relative standard error
SFWMD	South Florida Water Management District
SJRWMD	St. Johns River Water Management District
SO_4	Sulfate
SR	State Road
SR46H	Gauging station at SR 46, location H
SWFWMD	South West Florida Water Management District
USGS	U.S. Geological Survey
WSIS	Water Supply Impact Study

1 INTRODUCTION

The Water Supply Impact Study (WSIS) uses groundwater modeling to address several questions related to the potential impacts of water withdrawals on the St. Johns River. Groundwater modeling focuses on portions of the river where surface waters are influenced most heavily by groundwater. These areas primarily fall within the middle St. Johns River, although one analysis includes a portion of the lower Ocklawaha River (Figure 1-1), a major tributary to the St. Johns River. The St. Johns River flows north, and the middle St. Johns River is defined as the reach from location SR46H near Lake Harney (upstream; SR46H is a location where State Road 46 crosses the St. Johns River near Lake Harney) to State Road (SR) 40 at Astor (downstream). The section of the lower Ocklawaha River analyzed in this chapter extends from SR 40 near the confluence of the Silver River to County Road (CR) 316. The lower St. Johns is downstream of SR 40 at Astor, and the upper St. Johns River is upstream of SR46H.

The middle St. Johns River overlies an area of groundwater discharge from the regionally extensive Floridan aquifer system. Groundwater enters the river through numerous springs and as diffuse upward leakage into the river. Groundwater discharge is large enough to be an important component to river base flow, and this flow has been previously quantified using groundwater flow models (McGurk and Presley 2002, Motz and Dogan 2004). These groundwater flow models have also been used to examine the impacts to groundwater flows from proposed increases in regional pumping for water supply and irrigation (Agyei, Munch and Burger 2005). These same groundwater flow models are used here to examine the impacts that surface water withdrawals would have on groundwater discharge to the river due to decreasing river stage. Decreased river stage would increase groundwater discharge to the rivers, and increased base flow is not perceived as a threat to surface waters. In many areas, however, groundwater discharge has higher chloride concentration than surface waters, and a possible increase in chloride load to the river due to increased groundwater discharge is a perceived threat to surface waters. Our results show that the expected change to both groundwater discharge and chloride load from water withdrawals is insignificant relative to river conditions. This result is found because the hydraulic head in the Floridan aquifer system, a measure of the pressure within the confined aquifer underlying the river, which drives the groundwater discharge, is much larger (3 to 6 m above the river surface) than the expected reduction in river stage due to a withdrawal (1 to 5.5 cm).

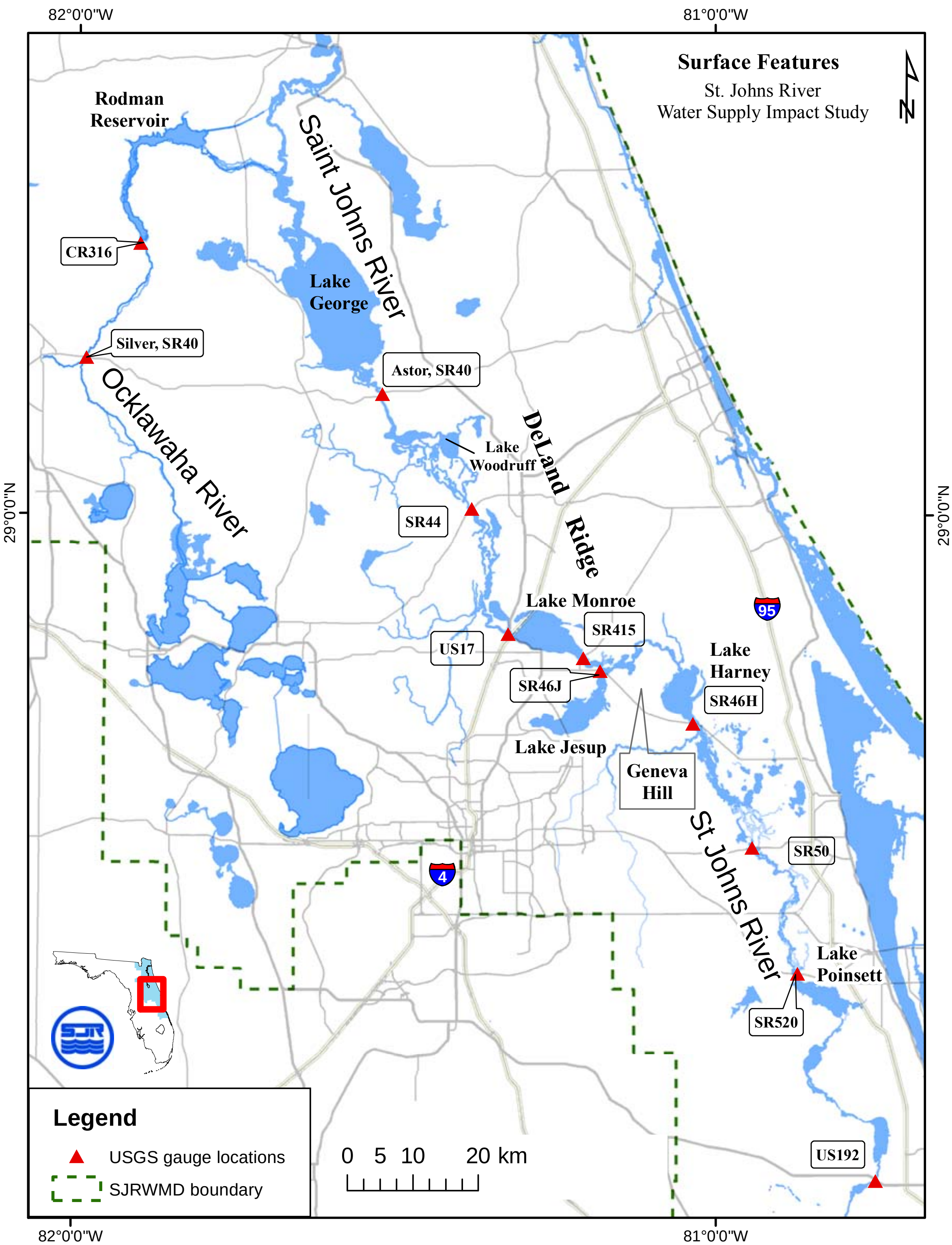


Figure 1-1. Map of the study area for groundwater analyses discussed in this chapter.

The groundwater flow models are also used to develop discharge and salinity boundary conditions for surface water modeling. Surface water modeling is used in support of the WSIS to predict how proposed surface water withdrawals would affect various physical characteristics of the river. A previous modeling study by Robison (2004) calculated stage and discharge reductions in the middle St. Johns River for a range of water withdrawal scenarios. That study concluded that river stage at DeLand would meet the requirements of the established minimum flows and levels (MFLs) for a continuous discharge reduction of up to 155 mgd. Here we extend the Robison study to examine salinity and water age (equivalent to hydraulic residence time) in addition to river stage and discharge. These additional variables are more useful than stage and discharge for assessing water quality impacts from water withdrawals. Addition of these variables required use of a hydrodynamic model to account for transport of salt and mixing processes within the river and its interconnected lakes. This chapter of WSIS provides the first hydrodynamic model application within the middle St. Johns River.

Although the existing groundwater flow models provide estimates of groundwater discharge for use in the surface water model, they had not previously been used to estimate the required chloride loads needed for salinity modeling. This chapter of WSIS develops a methodology for using the groundwater models for this purpose. The groundwater models are used to estimate chloride loads to different segments of the middle St. Johns River because chloride is a proxy for total dissolved salts. With a good estimate of chloride loads, the surface water model can then use chloride (or the associated salinity) as a conservative tracer to confirm the simulation of surface water mixing processes. In addition, the surface water model can be used to simulate the effects on river salinity from both water withdrawals and proposed desalinization processes.

Chloride load to the middle and upper St. Johns River, estimated using the groundwater models, totals 937 t d^{-1} from diffuse groundwater sources alone (not including springs). In the middle St. Johns River upstream of U.S. 17, diffuse groundwater supplies 75% of the chloride load to the river. Estimated chloride load compares well with observed load. Estimated and observed chloride loads are within 12% over a large portion of the middle St. Johns River between SR46H and SR 44.

The existing groundwater flow models do not simulate chloride transport, so they cannot simulate chloride load. We show, however, that because both chloride concentration and salt composition remain constant over time, then chloride load can be estimated externally to the groundwater flow model as a simple product of spatially varying chloride concentration and discharge. The temporal stability of groundwater chloride in this area was noted previously (Tibbals 1990).However, our analysis serves as a check that this condition was met during the period (1995 to 2005) used for surface water modeling. Salt composition of groundwater was examined in nearby wells (G. Phelps 2001).The analysis here is a comprehensive examination of salt composition in the Upper Floridan aquifer, near the middle St. Johns River. The chloride and salt analyses presented here are important in calculating the chloride load to the river from groundwater and in converting the chloride loads to salinity as required for boundary conditions to the surface water model.

The lack of chloride transport within the existing groundwater models also means that the dynamic links between chloride, density, pressure differences, and flow are not represented in the groundwater models. A groundwater model calibrated to hydraulic head in the Upper Floridan aquifer could underestimate pressure gradients, and consequently groundwater discharge, if density differences (resulting from large chloride differences) are large enough (Post, Kooi and Simmons 2007). By means of the method of equivalent freshwater head (EFH), we show that the inclusion of observed vertical density differences only increases simulated vertical discharge 3 to 13% in wells with the greatest stratification. Over the larger WSIS study area, the possible error in vertical discharge from ignoring density differences is likely less than 3%. The existing groundwater models, then, are adequate for the purposes of this groundwater study since (a) chloride transport is not needed to calculate chloride loads and (b) ignoring chloride-derived density differences within the model does not appreciably affect vertical flow calculations.

A second limitation of the existing groundwater flow models is that they simulate only steady-state groundwater flows. At the time of the WSIS, there were no calibrated transient groundwater models, although this is an area of active, cooperative research (SFWMD, SWFWMD, SJRWMD 2006). An additional question addressed here is how the use of steady-state, rather than transient groundwater discharge affects simulation of surface water variables. This question is addressed by comparing surface water model results obtained using the steady-state discharges, derived from the groundwater flow models, with transient results estimated from the temporal variability of observed hydraulic head in wells close to the river. The surface water model is shown to be insensitive to the choice of steady-state versus transient discharge, showing that steady-state discharge results from the groundwater models can be used directly for surface water modeling.

2 HYDROGEOLOGY OF THE ST. JOHNS RIVER

Aquifer Systems Underlying the St. Johns River

The groundwater system underlying the St. Johns River Basin comprises two primary aquifer systems: a surficial aquifer and the Floridan aquifer system. The surficial aquifer is generally unconfined; its upper surface, called the water table, is able to rise and fall freely and comprises Pliocene and Holocene sediments. The Floridan aquifer system is a regionally extensive aquifer system of Eocene age and one of the most productive in the world (Miller 1990). The Hawthorn Formation, an intermediate confining unit of Miocene age, separates it from the surficial aquifer (Spechler 1994).

The Floridan aquifer system is subdivided into upper and lower permeable zones: the Upper Floridan aquifer and the Lower Floridan aquifer. The Upper and Lower Floridan aquifers are separated by a low permeability unit called the middle semiconfining unit. The base of the Lower Floridan aquifer occurs at the top of low permeability anhydrite beds within the Cedar Keys Formation. This base layer is of Paleocene age and is called the sub-Floridan confining unit. The hydrogeologic units that form the underlying aquifer systems in the St. Johns River Basin are shown in Table 2-1.

Table 2-1. Generalized hydrogeologic units of the St. Johns River Basin.

<table>
<tr><th>Geologic Age</th><th>Stratigraphic Unit</th><th colspan="2">Hydrogeologic Unit</th><th>Lithologic Description</th></tr>
<tr><td>Pleistocene and Holocene (5.3 Ma to present)</td><td>Pleistocene and Holocene deposits</td><td colspan="2">Surficial aquifer system</td><td>Sand, shell, and clay lenses</td></tr>
<tr><td>Pliocene (5.3 to 2.6 Ma)</td><td>Late Miocene or Pliocene deposits</td><td colspan="2" rowspan="2">Intermediate aquifer system (intermediate confining unit and intermediate aquifer)</td><td rowspan="2">Clay, marl, and discontinuous beds of sand, shell, dolomite, and limestone</td></tr>
<tr><td>Miocene (23 to 5.3 Ma)</td><td>Hawthorn Formation</td></tr>
<tr><td rowspan="3">Eocene (56 to 34 Ma)</td><td>Ocala and Suwannee Limestone</td><td rowspan="3">Floridan aquifer system</td><td>Upper Floridan aquifer (upper and lower zones)</td><td>Very porous limestone</td></tr>
<tr><td>Avon Park Formation</td><td>Middle semiconfining unit</td><td>Leaky, low permeability limestone and dolomite</td></tr>
<tr><td>Oldsmar Formation</td><td>Lower Floridan aquifer</td><td>Porous limestone</td></tr>
<tr><td>Paleocene (65 to 56 Ma)</td><td>Cedar Keys Formation</td><td colspan="2">Sub-Floridan confining unit</td><td>Low permeability anhydrite beds</td></tr>
</table>

The Hawthorn Formation

The Hawthorn Formation is a confining layer that separates the surficial aquifer from the Upper Floridan aquifer throughout much of the St. Johns River Basin. It varies in thickness across the region because of alterations that have occurred over time. In most of the middle portion of the St. Johns River Valley, the Hawthorn Formation has been eroded following episodes of faulting and warping. As a result, the Hawthorn Formation is thin or absent in these uplifted areas (Scott 1988).

The variations in thickness of the Hawthorn Formation determine the extent of interaction between groundwater in the Upper Floridan aquifer and surface waters of the St. Johns River. Groundwater interactions are least where the Hawthorn Formation is thickest—in the lower St. Johns River and upper portions of the upper St. Johns River. In the lower St Johns River, the Hawthorn Formation thickens to a maximum of 500 ft downstream of SR 40 (Scott 1983). In the upper St. Johns River, the Hawthorn Formation thickens to 400 ft upstream of SR 520. The thick Hawthorn Formation contributes to the lack of springs and overall lack of diffuse groundwater discharge to the river in these areas (Spechler 1994, Conner and Belanger 1981). The Hawthorn Formation is thin or absent in the middle St. Johns River and lower portion of the upper St. Johns

River, from about SR 520 to SR 40, ranging in thickness from 0 to 50 ft (Johnson 1982). These areas contain numerous springs and diffuse groundwater discharge of Upper Floridan aquifer water to the river is appreciable.

Pleistocene Sea Level Variability

The rise and fall of sea level during the Pleistocene epoch has caused repeated inundation and exposure of the St. Johns River Valley. The St. Johns River Valley has been inundated to a depth of at least 7.5 m (25 ft) several times during the last two million years (Alt and Brooks 1965). The sea level during the last glacial maximum, dropped as much as 100 m (300 ft) below present level (Zuener 1938). Transgression (sea level rise) and regression (sea level fall) due to glacial cycles (Flint 1947) occurred at least eight times during the Pleistocene (Vail Jr. and Mitchum 1978). Transgressions have exceeded present sea level at least four times (Zelmmer 1979).

During high sea level stands, lagoons formed between uplands to the west and topographically high features, such as the DeLand Ridge and Geneva Hill, to the east (White 1970). The lagoons, formed by inundation of the St. Johns River Valley, were connected to the ocean by widely spaced inlets similar to present-day coastal barrier islands. High features above the 7.5 m (25 ft) contour formed coastal barrier islands that are recharge areas on the east side of the St. Johns River (Figure 2-1). The St. Johns River Valley below the 7.5 m (25 ft) contour was inundated to sufficient depth to force Pleistocene seawater into the Eocene substrates of the Floridan aquifer system (Stringfield and Cooper 1951). Since the last low sea level stand, Pleistocene seawater has been flushing out of the Floridan aquifer system, to varying degrees, by active groundwater circulation (Stringfield and Cooper 1951).

Relict seawater is the correct term for the trapped Pleistocene seawater. Although it is sometimes called connate (from the Latin word for “born together”) that term implies waters of Eocene age or older, because that is the age of the geologic formation containing the Floridan aquifer system (Bouwer 1978). Relict seawater, the remnant portion of Pleistocene seawater, and its associated salinity now enter the St. Johns River by diffuse groundwater and spring discharges.

Portions of the St. Johns River bed were incised 30 m (100 ft) below the riverbed’s present level during low sea level stands (Stringfield and Cooper 1951). The incision cut through the Hawthorn Formation in areas where the overburden and Hawthorn Formation were already thin, particularly in the middle St. Johns River. As sea level rose, sediment was deposited in the incision because the low slope of the St. Johns River had insufficient energy to carry sediment out of the river and into the Atlantic Ocean (Linsley 1982). Although the Pleistocene incision is now filled with sediment (Kindinger, Davis and Flocks 2000), the incised feature still enhances the connection between the Floridan aquifer system and the St. Johns River (V. Stringfield 1936) and allows flow of relict seawater into the river.

Present-Day Recharge and Discharge

The previous sections identified the middle St. Johns River and lower part of the upper St. Johns River as areas of direct interaction between the Upper Floridan aquifer and the river. These are areas where the Hawthorn Formation is thin and the riverbed was incised. These areas also lie within discharge areas (Figure 2-1), defined generically as any area where the potentiometric surface of the Upper Floridan aquifer is above the water table of the surficial aquifer (or river surface). The potentiometric surface of the Upper Floridan aquifer is generally 3 to 6 m (10 to 20 ft) above the surface of the St. Johns River, even accounting for seasonal and interannual variability in response to rainfall and pumping. The positive head difference of 3 to 6 m makes flow from the Upper Floridan aquifer to the river possible; the positive head difference combined with the direct interaction between the Upper Floridan aquifer and river allow groundwater underlying the river to discharge to the river.

Most discharge from the Upper Floridan aquifer occurs through springs and by diffuse groundwater discharge along the streams and rivers in areas where conditions are suitable. Springs are generally located along the edge between recharge and discharge areas (Figure 2-1). Discharge is not quantified in Figure 2-1 because of the difficulty in determining these values. Estimation of discharge from the Upper Floridan aquifer to the middle St. Johns River is one purpose for the groundwater modeling described in this chapter.

Discharge from the Upper Floridan aquifer to rivers and streams is supplied by recharge to the Upper Floridan aquifer in areas distant from stream and river valleys (Figure 2-1). Recharge rates are calculated by the difference of known rates of precipitation, runoff, evaporation, and evapotranspiration (Phelps 1982). Recharge rates vary widely with high rates of more than 20 in. yr^{-1}. On the east side of the St. Johns River, high recharge rates along the DeLand Ridge produce an underlying dome of water that causes discharge along its base of Upper Floridan aquifer waters to the river. High recharge areas west of the river support a positive head difference that drives diffuse groundwater discharge through the river bottom (Phelps and Rohrer 1987) and allows for many springs.

Springs are an important source of groundwater to the St. Johns River (Figure 2-1). The Ocklawaha River, middle St. Johns River, and Lake George basins contain most of the springs entering the St. Johns River. Spring discharges are relatively constant compared to tributaries that are dependent on surface water runoff. Therefore, springs are important contributors to base flow of the St. Johns River (Ferguson 1947). Silver Springs is the largest spring in the St. Johns River Water Management District (SJRWMD) , with a mean discharge of about 500 mgd. Other large springs of note are Blue (Volusia County), Silver Glen, Alexander, and Salt with mean discharges of 100, 69, 66, and 52 mgd, respectively. Most discharge from the western and southern side of the DeLand Ridge is subsurface, entering the river through four springs—Green, Gemini, Blue, and Ponce DeLeon—in addition to diffuse groundwater flow. A few small springs occur in the lower St. Johns River, but these originate predominantly from the surficial aquifer. The upper St. Johns River is essentially devoid of springs.

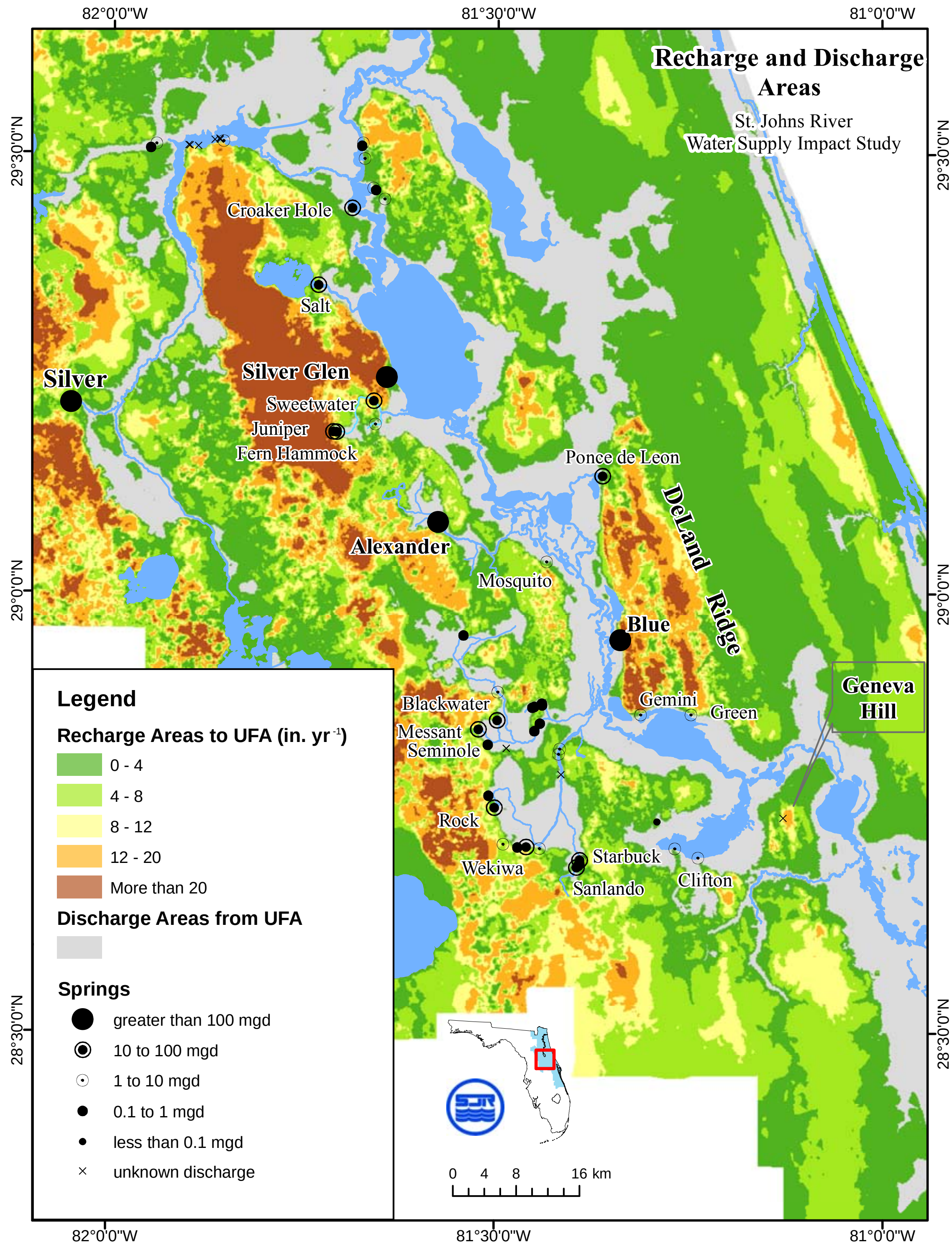


Figure 2-1. Areas of recharge to and discharge from the Upper Floridan aquifer in and near the groundwater study area along with springs that discharge from the Upper Floridan aquifer.

Relict Seawater and Chloride

Relict seawater from the Floridan aquifer system is the primary source of chloride to the upper and middle St. Johns River and produces oligohaline conditions in the river far upstream of any ocean influence. The unusually high level of chloride in the middle St. Johns River, in comparison with typical freshwater systems, results in a unique water quality that supports a wide distribution of marine fauna (Odum 1953). The areas of highest groundwater discharge, described previously, also have the highest chloride loading to the river (Tibbals 1990). High chloride loading can occur where either groundwater discharge is high and/or the chloride concentration of the inflowing groundwater is high. In the St. Johns River, areas of high groundwater discharge coincide with areas of high chloride concentration (Barraclough 1962). Chloride concentrations in the Upper Floridan aquifer exceed 1,000 mg L^{-1} along most of the course of the upper and middle St. Johns River (Figure 2-2).

Spring discharges containing relict seawater are also an important source of chloride to the St. Johns River. Not all large springs contribute a large chloride load because chloride concentrations vary considerably among springs. Blue Spring (100 mgd) has a greater chloride load than Silver Springs (500 mgd) because chloride in Blue Spring is 400 mg L^{-1}, whereas chloride in Silver Springs is only 10 mg L^{-1}. The source of chloride from Silver Springs is not relict seawater, but sea spray that is transported inland as an aerosol mixed with precipitation (Upchurch and Randazzo 1997). Salt Springs, a second-magnitude spring group entering the northeast corner of Lake George, is the single largest point source of chloride to the St. Johns River, with a mean discharge of 52 mgd and chloride of over 1500 mg L^{-1} (see Chapter 5).

Areas of high groundwater discharge have a high underlying chloride concentration because the same hydrogeologic factors, now favoring discharge from the Upper Floridan aquifer, also favored penetration of seawater into the underlying Eocene sediments during Pleistocene transgressions. Seawater more easily penetrated Eocene sediments in areas where the Hawthorn Formation is thin and incised. Along the entire St. Johns River, the highest concentration of Upper Floridan aquifer chloride occurs between lakes Poinsett and Harney, where the overburden has a minimum thickness of less than 9 m (30 ft) (Brown 1962). Areas that were not inundated during the Pleistocene (e.g., Silver Springs) now have low underlying chloride (Figure 2-2). In general, areas above 7.5 m (25 ft) elevation have chloride levels less than 50 mg L^{-1}. A second reason for the positive correlation between groundwater discharge and chloride is groundwater circulation patterns.

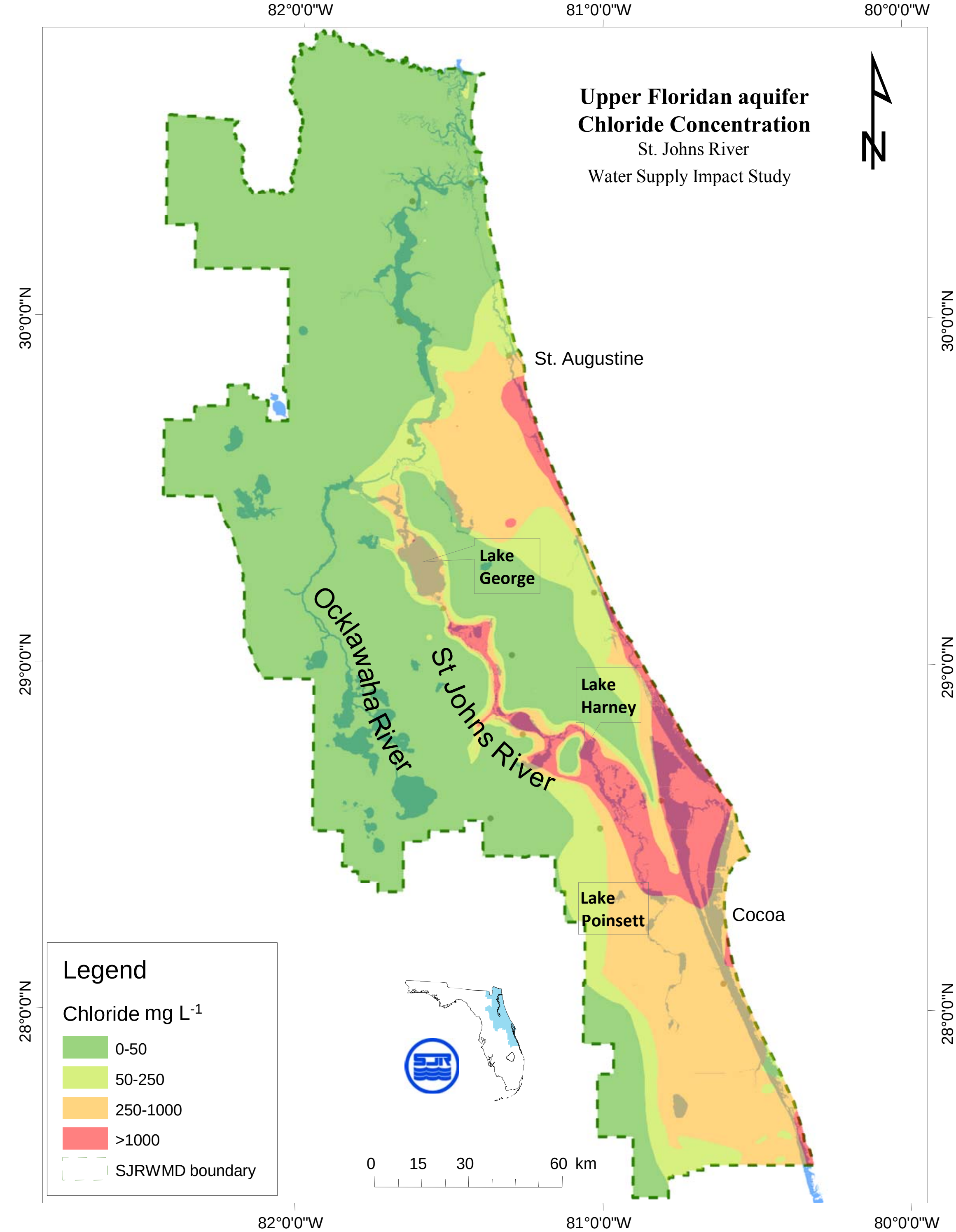


Figure 2-2. Chloride concentration of the Upper Floridan aquifer within SJRWMD.

Groundwater Circulation Patterns

The spatial distribution of chloride underneath the St. Johns River is primarily a function of two factors: the location and amount of relict seawater that entered the Floridan aquifer system during high sea level stands (discussed above), and groundwater circulation patterns that flush relict seawater from the aquifer. High groundwater flows to the middle St. Johns River and lower portion of the upper St. Johns River pick up relict seawater from below the river basin and transport it into the river. The relationship between groundwater circulation and transport of relict seawater to the river is illustrated by two hydrogeologic cross sections (Figure 2-3). Both cross sections extend from southwest to northeast along a transect perpendicular to the general axis of peninsular Florida and the St. Johns River. The north cross section intersects the middle St. Johns River in an area of high groundwater discharge to the river, while the south cross section intersects the upper St. Johns River in an area of low groundwater discharge.

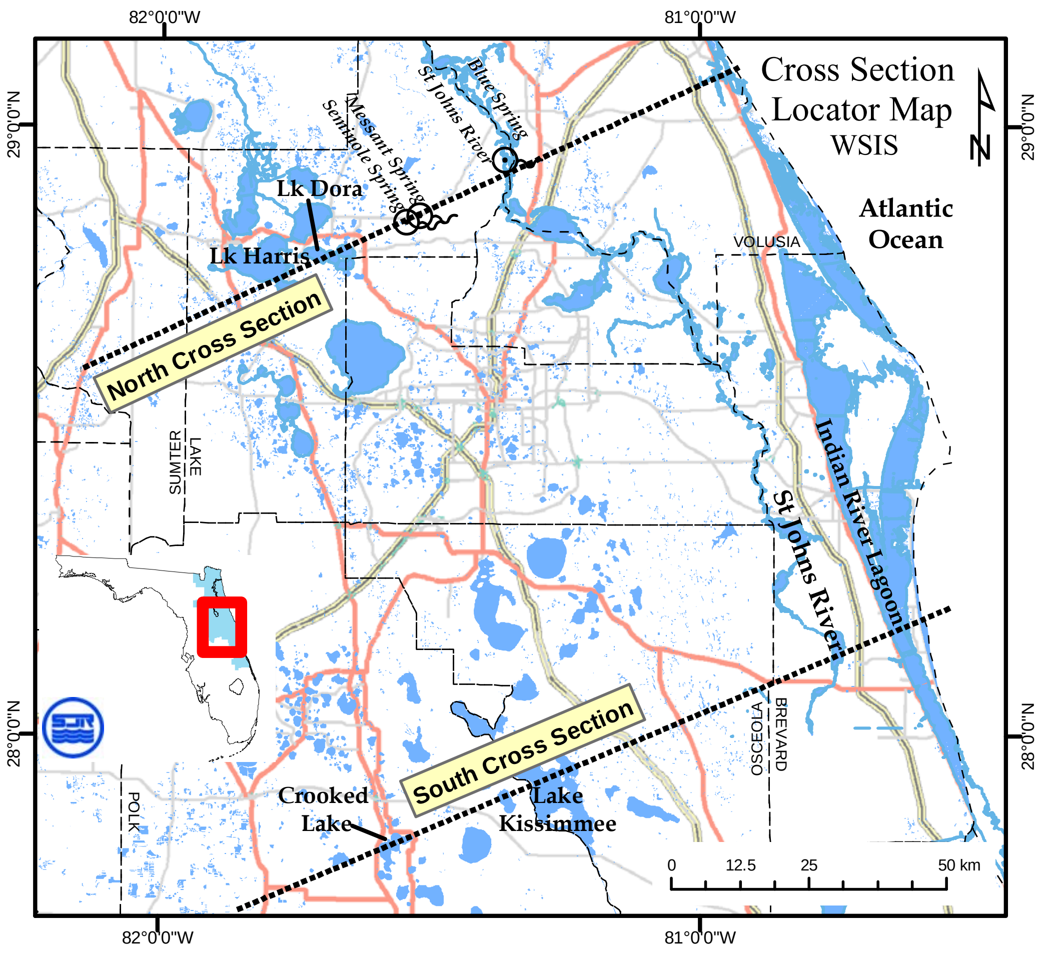


Figure 2-3. Location of hydrogeologic cross sections intersecting areas of high (north) and low (south) groundwater discharge to the St. Johns River.

In the north cross section, the upper confining unit thins, and the Upper Floridan aquifer approaches the surface near the St. Johns River (Figure 2-4). The potentiometric surface of the

Upper Floridan aquifer (upper dashed line) is above the level of the river, which is almost at sea level. Upper Floridan aquifer flow is directed toward the river from both east and west so that the river is a zone of convergence. The depiction of a large discharge coming from the mound on the eastern side of the river is recharge water from the DeLand Ridge. The discontinuity in subsurface features immediately underneath the St. Johns River is called the Volusia Fault. The Volusia Fault provides a face for lateral flow into erosional sediments that filled in the Pleistocene incision beneath the St. Johns River (Wyrick 1960) and provides a conduit for transport of relict seawater. Recharge along the DeLand Ridge produces a convex bowl of freshwater surrounded by higher salinity that augments the transport of relict seawater from the Lower Floridan aquifer.

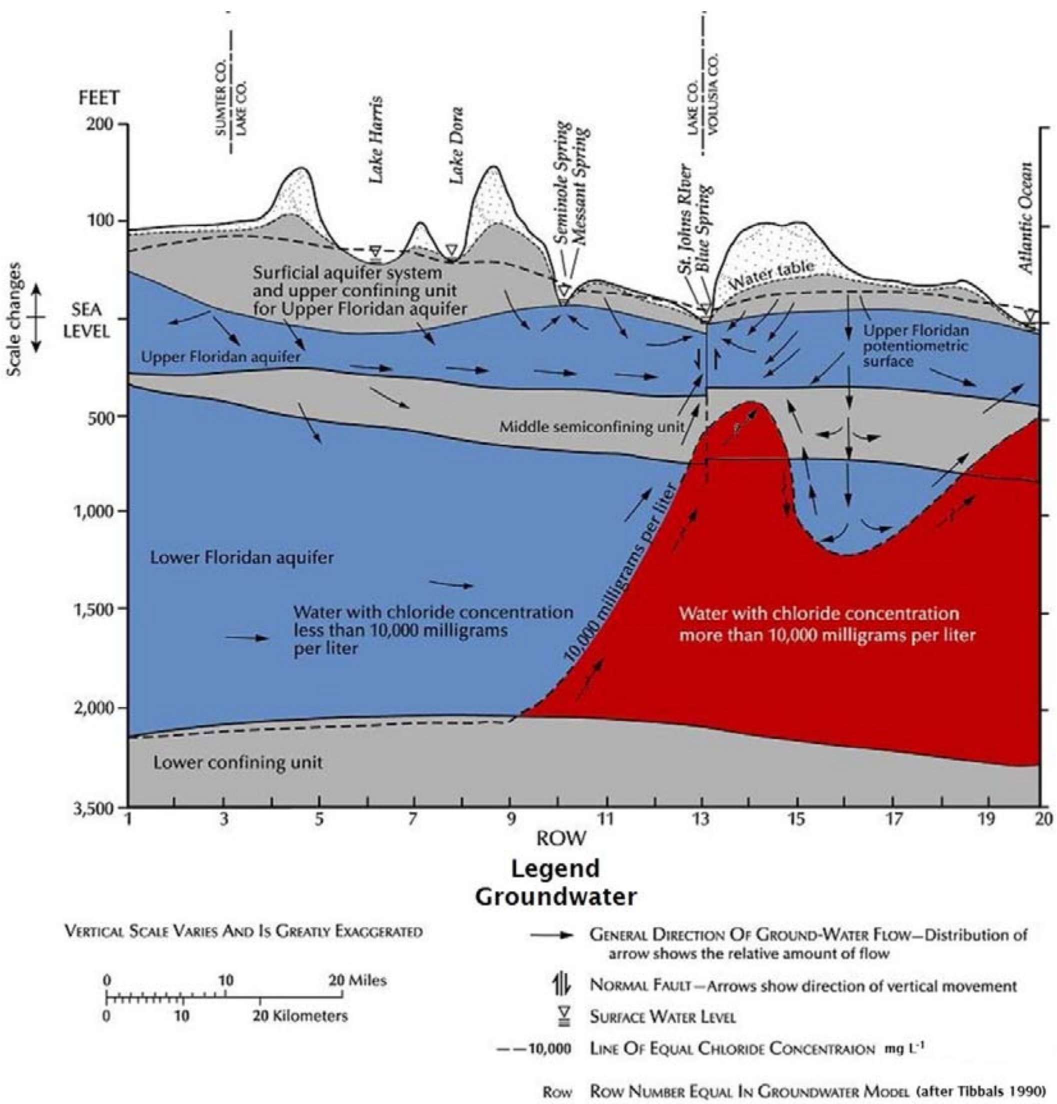


Figure 2-4. North hydrogeologic cross section of the St. Johns River Valley from Sumter County to Volusia County crossing the St. Johns River in an area of high groundwater discharge (Tibbals 1990).

In the south cross section, the potentiometric surface of the Upper Floridan aquifer is still above the level of the river, indicating a region of potential discharge and absence of recharge (Figure 2-5). However, several factors prevent large discharge of groundwater to the river. First, the extent of recharge areas to the west of the St. Johns River is less compared with the north cross section. (Recharge areas occur where the potentiometric surface of the Upper Floridan aquifer is below the water table.) There are no recharge areas directly east of the St. Johns River along this cross section. The small uphill recharge results in minimal discharge to the river. Second, the overburden beneath the river is thicker in the south than in the north cross section, placing the Upper Floridan aquifer farther from the surface, and the thickness of the overburden inhibits transport of Upper Floridan aquifer waters to the surface.

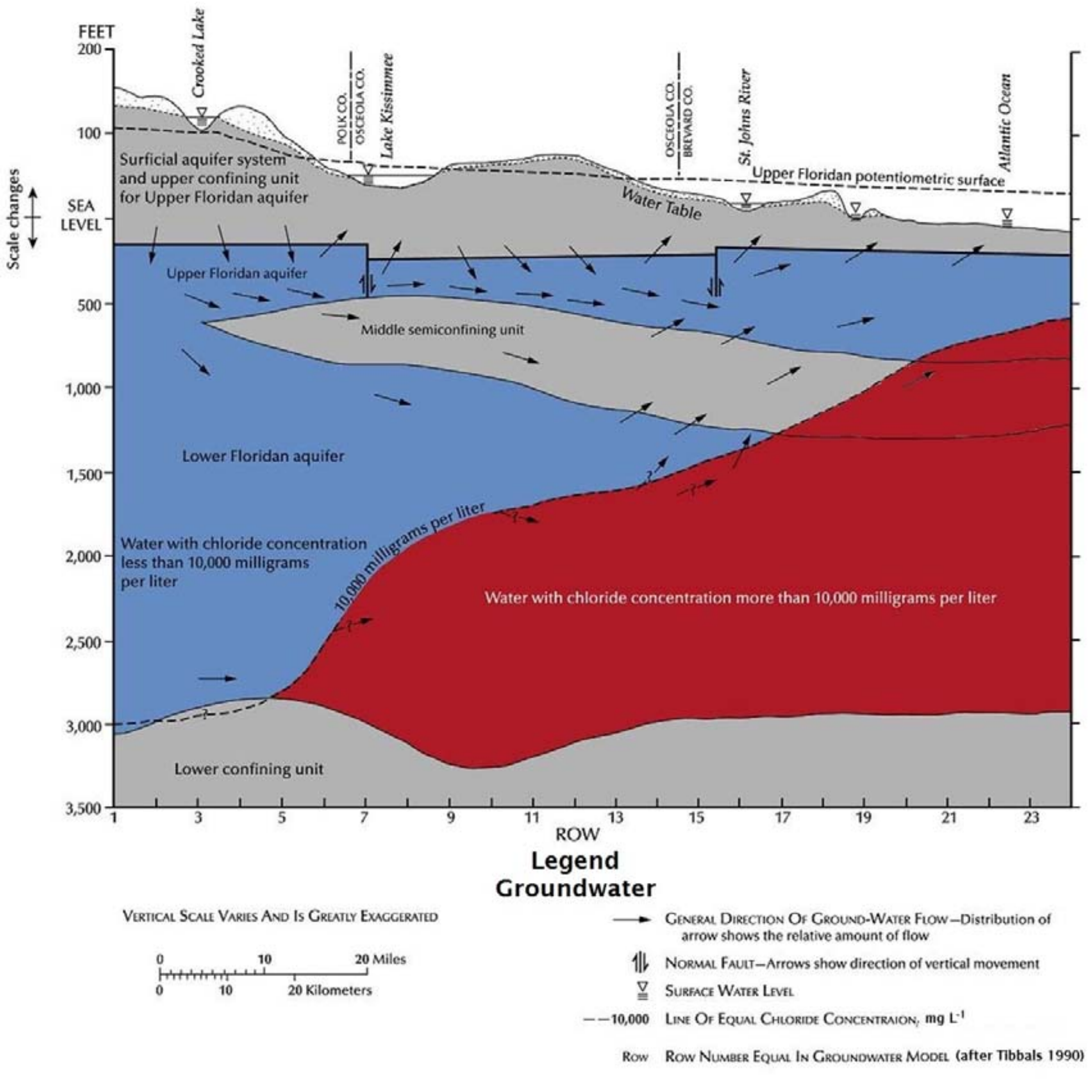


Figure 2-5. South hydrogeologic cross section of the St. Johns River Valley from Polk County to Brevard County crossing the St. Johns River in an area of low groundwater discharge (Tibbals 1990).

Summary

The Floridan aquifer system is a large, regionally extensive aquifer system beneath the St. Johns River. The potentiometric surface of the Floridan aquifer system is above the level of the river everywhere, making discharge from the aquifer to the river possible. In some areas the Hawthorn Formation, a confining layer, inhibits interaction between the river and aquifer. However, in the middle St. Johns River and lower portion of the upper St. Johns River, the Hawthorn Formation is thin or absent, and the river in these areas has direct connection to the aquifer. A Pleistocene incision beneath the river channel, now filled with sediment, further enhances connection between the river and aquifer in the middle St. Johns River. As a result, the river receives appreciable groundwater flows from numerous springs and diffuse groundwater discharges that are important contributors to base flow.

Chloride concentrations are high in groundwater underlying the middle St. Johns River. The chloride is derived from relict seawater that was trapped in the underlying sediments during Pleistocene transgressions. Relict seawater is found in large areas of the St. Johns River Valley below the 7.5 m (25 ft) contour. Active groundwater circulation flushes relict seawater from the aquifer, transporting chlorides to the river. Chloride from groundwater discharge is responsible for the unique oligohaline character of broad areas of the St. Johns River.

3 GROUNDWATER MODEL FRAMEWORK

3.1 INTRODUCTION

A model framework describes the level of detail included in a model that is appropriate to the model's purpose. The model framework leads to selection of an appropriate model and modeling approach that includes relevant processes while avoiding inclusion of unnecessarily complex processes. An overly complex model is undesirable because it can produce uncertain results, may require input data that cannot be accurately obtained from field measurements, and can have unnecessarily excessive modeling costs (National Research Council 1989). Knowledge of the groundwater study area and understanding of the state-variables to be simulated is central for selecting a suitable and defensible numerical model and modeling approach. This section describes analyses that lead to simplifying assumptions for the following three model components:

- Temporally constant chloride concentrations
- Constant chloride-salinity relationship for diffuse groundwater flow
- Constant density of groundwater for vertical flow calculations

A steady-state assumption for groundwater flows is discussed separately in Section 5.

Groundwater modeling is used to quantify chloride loading to the river for application to a surface water model. The estimation of chloride load is made simpler by assuming temporally constant chloride concentrations. Under this assumption, chloride loading is estimated from the product of simulated groundwater flows and observed chloride concentrations. A complex solute transport model is then not required.

Groundwater chloride concentrations are subsequently converted to salinity for use in the surface water model. This conversion is simplified by assuming stability of salt composition, which makes the chloride-salinity relationship constant at any given location. A basic understanding of sources of salt-to-groundwater entering the river further establishes a single, constant chloride-salinity relationship, equivalent to that of seawater, for all diffuse groundwater sources entering the middle St. Johns River.

Vertical groundwater flows are simulated by the groundwater model under the assumption of constant density, that is, no vertical stratification of groundwater salinity. This assumption eliminates the need for detailed knowledge of the vertical structure of the groundwater-salinity field and the need for a solute transport model and is justified by the comparison of density-independent and density-dependent vertical flow calculations at three well locations.

3.2 SJRWMD Monitoring Wells

Analyses for this section use data collected as part of the SJRWMD groundwater-quality monitoring network (Boniol 2002). The network measures physical and chemical properties of groundwater within monitoring wells placed throughout SJRWMD. Most regional groundwater monitoring wells tap the Upper Floridan aquifer because groundwater from the Upper Floridan aquifer is the chief source of water supply within SJRWMD (SJRWMD 2010). For this study, 12 monitoring wells were selected for their proximity to the course of the middle St. Johns River (Figure 3-1).

These wells measure the hydraulic head of the aquifer, specific conductivity, temperature, alkalinity, and a suite of dissolved inorganic ions, including sodium (Na^+), magnesium (Mg^{2+}), sulfur (as SO_4^{2-}), chloride (Cl^-), potassium (K^+), and calcium (Ca^{2+}). These water-quality data are used in the analyses below to examine temporal and spatial variability of both salt composition and chloride concentrations of the Upper Floridan aquifer near the middle St. Johns River. Aquifer hydraulic head is used to estimate the relative importance of vertical density gradients on vertical groundwater flows. Aquifer hydraulic head over the period 1995 to 2005 is also used in Section 5 to examine the sensitivity of a surface water model to transient groundwater flows. The five wells in Figure 3-1 noted with a star have the most complete observed hydraulic head for this period.

Among all wells, the average period of record is 16 years (1992 to 2008), with a mean sample frequency of about two samples per year. A sample drawn from a well is collected when temperature, pH, and specific conductivity do not change by more than 10% between two successive well volumes. Toth (1999) provided a detailed description of the sampling methodology in a comparative analysis of water quality characteristics and water age within springs and nearby Upper Floridan aquifer wells. (That study showed that water age within Upper Floridan aquifer wells is a mixture of waters of different ages, ranging from new sources of less than 50 years to relict sources thousands of years old.) The specific periods of record and sample counts (for most, but not all, parameters) for each well are shown in Table 3-1.

Table 3-1. Period of record and sample count for 12 Upper Floridan aquifer monitoring wells near the middle St. Johns River.

Well ID	Period of Record	Sample Count
BR-1526	October 1996 to April 2009	16
S-0025	September 1987 to July 2008	31
S-0034	October 1987 to January 2010	37
V-0165	October 1987 to December 2005	24
V-0818	December 1996 to July 2008	22
V-0240	October 1986 to June 2009	27
S-1397	November 1999 to November 2008	38
L-0032	June 1990 to December 2008	43
V-1091	August 2000 to August 2008	17
V-0083	October 1990 to February 2008	37
L-0059	October 1991 to September 2004	50
L-0455	May 1992 to December 2008	42

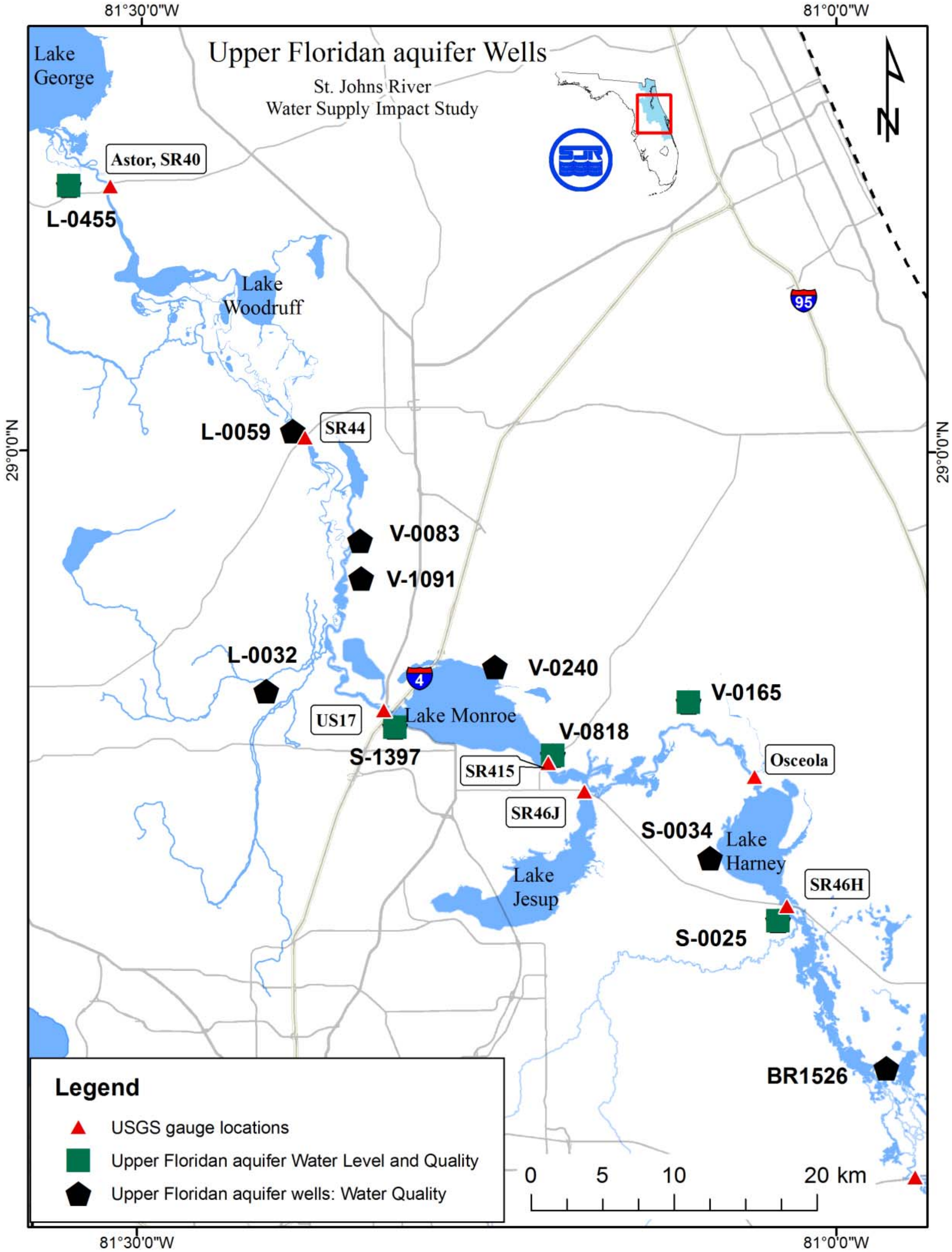


Figure 3-1. SJRWMD groundwater quality monitoring wells near the middle St. Johns River. The five wells, indicated by a square, measured Upper Floridan aquifer hydraulic head during the surface water model simulation period (1995 to 2005).

3.3 Stability of Chloride

Introduction

Flow through aquifers is very slow, measured in days versus seconds for surface waters (Bouwer 1978). Groundwater chloride in the Upper Floridan aquifer, then, is expected to exhibit relatively modest temporal gradients compared to spatial gradients, which are quite steep in a direction lateral to the riverbed. Spatial variability of chloride in the confined Upper Floridan aquifer likely results from circuitous flow paths and the amount of relict seawater within a given unit. Temporal variability depends on the replacement rates of recharge water and associated dissolved salts (Tibbals 1990).

The stability (i.e., lack of temporal variability) of chloride in groundwater underlying the river is shown by examining observed chloride within observation wells in the following four ways:

- Calculation of relative standard error (*RSE*)
- Visualization of chloride time series
- Visualization of ranked chloride observations
- Comparison of chloride with rainfall indicators

The four analyses provide evidence that chloride throughout the WSIS study area is very stable and can be assumed constant for estimating chloride loads to the river.

Relative Standard Error of Chloride Observations

Evidence for stability is shown first by a statistical summary of chloride in the 12 monitoring wells. The minimum and maximum chloride values fall close to the mean within each well (Table 3-2). *RSE* are generally less than 10%, indicating low temporal variability of chloride concentration about the mean within each well. *RSE* is a normalization of the standard deviation using the mean and is useful for comparison of datasets with widely varying mean values (Schefler 1969).

Table 3-2. Upper Floridan aquifer monitoring wells and associated chloride descriptive statistics used in this analysis.

Well ID	Period of Record	Chloride (mg L^{-1})				
		Min	*Mean*	*Max*	*RSE* (%)*	Count
BR-1526	Oct 1996 to Apr 2009	1580	1674	1810	1.05	16
S-0025	Sep 1987 to Jul 2008	4620	5144	5798	0.97	31
S-0034	Oct 1987 to Jan 2010	9	10	12	1.27	37
V-0165	Oct 1987 to Dec 2005	24	27	29	0.87	24
V-0818	Dec 1996 to Jul 2008	629	658	692	0.63	22
V-0240	Oct 1986 to Jun 2009	478	604	706	2.32	27
S-1397	Nov 1999 to Nov2008	39	43	50	0.86	38
L-0032	Jun 1990 to Dec 2008	685	751	800	0.60	43
V-1091	Aug 2000 to Aug 2008	626	669	693	0.72	17
V-0083	Oct 1990 to Feb 2008	2750	3165	3757	1.26	37
L-0059	Oct 1991 to Sep 2004	148	176	195	1.02	50
L-0455	May 1992 to Dec 2008	8	9	11	1.35	42

* *RSE*: relative standard error

Visualization of Chloride Time Series

Time series for chloride concentration are plotted for five wells (Figure 3-2). The chloride concentrations vary widely between wells with mean concentrations varying from 9 to 5,144 mg L^{-1}. Only five wells are selected for plotting to prevent the wide spatial range of chloride between individual wells from obscuring the smaller temporal variability within an individual well. The wells are classified into three chloride types: NaCl, mixed, and Ca-Mg-HCO_3. The U.S. Geological Survey (USGS) classification system for Florida waters is discussed in Section 3.4. The y-axis scale changes for each of the three plots to show the temporal variability observed within each group of wells. The plots show small variability of chloride within wells and no interannual trends over the 16 years of observations (except for well L-0059 where chloride is trending down slightly). The sampling frequency of two points per year, on average, is probably too infrequent for observing temporal variability related to seasonal–meteorological variability, however. An analysis of chloride variability based on rainfall indicators is used to test for the effects of seasonal variability below.

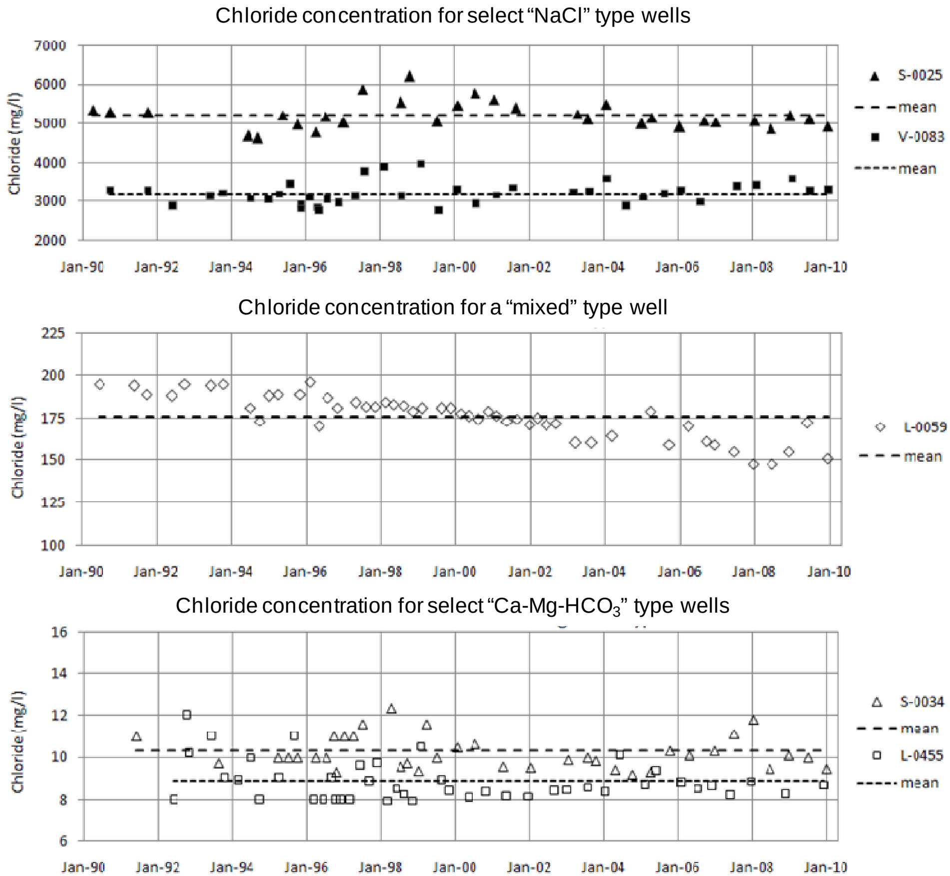


Figure 3-2. Time series of chloride concentration in selected Upper Floridan aquifer wells sorted by chloride classification.

Visualization of Ranked Chloride

The relative stability of chloride compared with the spatial variability across the groundwater modeling study area is shown by plotting sorted chloride values for each well on a logarithmic scale (Figure 3-3). Chloride observations for all wells are sorted from minimum to maximum value. The log scale allows all wells to be plotted simultaneously, and shows that spatial variability of chloride is much greater than temporal variability. Chloride varies over four orders of magnitude among wells while ranked-chloride values within a well are nearly flat, indicating that temporal variability of chloride is insignificant compared with spatial variability.

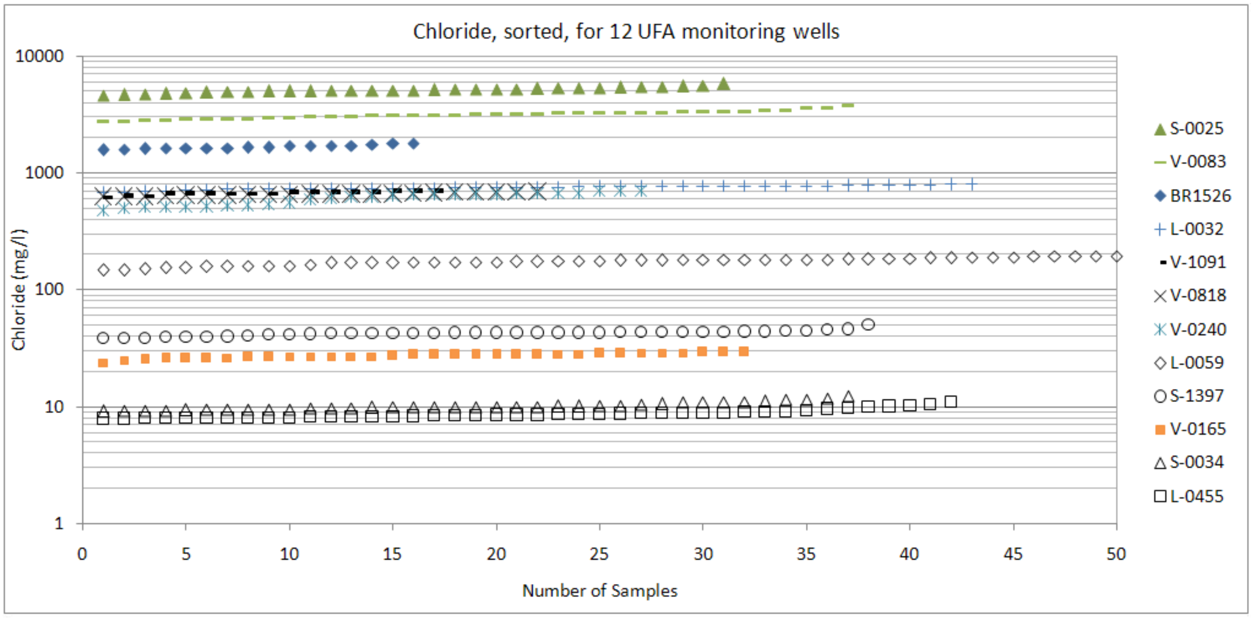


Figure 3-3. Ranked chloride observations in 12 Upper Floridan aquifer wells.

Comparison of Chloride to Rainfall Indicators

The low sampling frequency (two observations per year) of well observations could obscure seasonal variability. Two tests were conducted to determine if the chloride concentrations in the Upper Floridan aquifer are influenced by seasonal variability of rainfall by using observed river stage and Upper Floridan aquifer hydraulic head as indicators of rainfall variability. The first test correlated paired values of chloride concentration with river stage. River stage responds rapidly to rainfall and changes in river stage could have a direct influence on groundwater flow patterns, perhaps even forcing river water into the aquifer. The second test correlated paired values of chloride concentration with Upper Floridan aquifer hydraulic head. Variability in hydraulic head directly alters the magnitude of groundwater discharge to a well, but also could alter the relative proportions of flow coming from different aquifers. Spring discharge can vary in this way because it is often a mix of water from the Upper Floridan aquifer and a local surficial aquifer. Blue Spring exhibits this trait, with chloride being diluted by increasing groundwater discharge due to increased surficial influence (Osburn, Toth and Boniol 2006).

For the first test, the chloride concentration in two wells (BR-1526 and S-0025) is correlated to river stage at SR46H above Lake Harney. These wells were selected based on their location in an area of the St. Johns River that experiences the greatest variability in stage. River stage and chloride concentration are uncorrelated for either well (r^2 = 0.032 for BR-1526, r^2 = 0.021 for S-0025).

For the second test, the chloride concentration in two wells (S-1397, and L-0455) is compared to Upper Floridan aquifer hydraulic head. These wells were selected for completeness of paired chloride and Upper Floridan aquifer hydraulic head observations. Chloride and hydraulic head are uncorrelated in both wells (r^2 = 0.08 for S-1397; r^2 = 0.001 for V-0455). These tests indicate that rainfall variability likely does not cause seasonal variability of chloride in the Upper Floridan aquifer within the WSIS study area.

Summary

The four analyses indicate that chloride throughout the WSIS study area is stable. The stability of chloride exists over a wide range of mean chloride concentrations. Wells with low chloride concentrations have no greater percentage of variation than wells with high chloride concentrations. The stability of chloride means that estimates of chloride loads to the river can be simplified by matching observed aquifer chloride concentration with estimates of groundwater discharge.

3.4 Stability of Salt Composition

Introduction

In the preceding section, we show that chloride concentration in the Upper Floridan aquifer is stable across the study area. This section addresses the stability of the salt composition. Although absolute chloride does not change in time, the possibility exists that the proportion of chloride relative to other major ions could change in time. This question is relevant for converting chloride concentration to salinity for use in the surface water model. When salt composition is stable, chloride-salinity relationships are stable.

The relative abundance of the eight main ionic species found in water samples—bicarbonate (HCO_3^-), carbonate (CO_3^{2-}), sodium (Na^+), magnesium (Mg^{2+}), sulfate (SO_4^{2-}), chloride (Cl^-), potassium (K^+), and calcium (Ca^{2+})—is assessed to evaluate the stability of salt composition in Upper Floridan aquifer wells. Hereafter the charge (i.e., +, -) is omitted. Relative abundance is the ratio or percentage of each constituent in solution. These eight ionic species constitute the majority of salt in ocean water (>99%), with the remainder being made up of trace elements (Stumm and Morgan 1981). The term salt is used here in the chemical sense to mean any ionic compounds that can result from the neutralization of an acid and a base.

Three important tools are used to examine the stability of salt composition: normalization of ionic concentrations, USGS chemical classification, and Maucha diagrams. The general application of these tools is first described for three wells with greatly differing mean chloride.

Salt composition stability analysis is completed by calculating the RSE for normalized chloride within all observation wells with sufficient record length. Results indicate that salt composition, like chloride concentration, is stable in groundwater underlying the middle St. Johns River.

Estimation of Bicarbonate and Carbonate from Alkalinity

Observed concentrations of the eight main ionic species are obtained for each of the 12 monitoring wells (see Figure 3-1). CO_3 and HCO_3 concentrations are calculated from observed pH and alkalinity (Sawyer, McCarthy and Parkin 1994). HCO_3 constituted 99.6 % of the carbonate alkalinity—averaged for all samples in this analysis. HCO_3 is the dominant component of alkalinity in the range of pH typical to these subsurface waters.

Normalization of Ionic Concentrations

Salt composition between samples is easily compared by normalizing the observed ionic concentrations. Concentrations for a set of eight ions are normalized by the sum of the concentration for each ion multiplied by the number of ions (eight). In the normalized set of ions, the average of the normalized values is equal to one and the sum is equal to eight. For example, mean concentrations of salt ions for three wells are normalized in Table 3-3. The salt composition in the wells is compared to the ocean standard abundance for the eight main ionic constituents. (The same eight ionic species that dominate salt composition in the Floridan aquifer system wells also dominate salt composition of the ocean.) The salt composition of two wells (L-0059 and S-0034) differs from the ocean standard, while the third well (S-0025) is similar.

Table 3-3. Mean concentration of eight major salt ions and normalized values for three representative Floridan aquifer system wells and comparison with the ocean standard.

Well	Observed Concentration (mg L^{-1})						Calculated Concentration (mg L^{-1})	
	Cl	SO_4	Na	K	Mg	Ca	HCO_3	CO_3
Ocean standard*	19354	2712	10770	399	1290	412.1	140.67	1.66
S-0025	5297.2	812.4	2811.1	96.8	285.7	270.5	144.52	0.18
L-0059	171.1	38.8	99.4	3.5	19.9	61.7	181.55	0.62
S-0034	10.0	4.0	7.0	1.1	1.6	72.0	187.97	0.05
Well	Normalized Concentration (PSS78)†						Normalized Concentration (PSS78)	
	Cl	SO_4	Na	K	Mg	Ca	HCO_3	CO_3
Ocean standard	4.4138	0.6185	2.4561	0.0910	0.2942	0.0940	0.0321	0.0004
S-0025	4.3606	0.6687	2.3140	0.0797	0.2352	0.2227	0.1190	0.0001
L-0059	2.3744	0.5383	1.3794	0.0484	0.2759	0.8562	2.5189	0.0086
S-0034	0.2820	0.1128	0.1974	0.0310	0.0451	2.0302	5.3002	0.0013

*Ocean standard presented as concentration for comparative purposes
†The Practical Salinity Scale 1978 (PSS78)(Lewis and Perkin 1978); (see Chapter 5)

USGS Chemical Classification

The wells in the above example were preselected to fall within, and illustrate, three primary chemical classes established by USGS for springs (Slack and Rosenau 1979). All 12 wells used in this analysis fall within the following three USGS chemical classes: NaCl (salts dominated by relict seawater), mixed (an almost equal mix of salts derived from bicarbonate dissolution and relict seawater), and Ca-Mg-HCO_3 (salts primarily derived from bicarbonate dissolution). A fourth USGS chemical classification for waters dominated by SO_4 is found in certain tributaries and springs within SJRWMD (see Chapter 5), but is not found among the 12 wells used in this analysis.

The normalized concentrations in Table 3-3 identify the USGS chemical classifications. The NaCl well (S-0025) has normalized chloride (4.3606) nearly identical to the ocean standard (4.4138). By contrast, the Ca-Mg-HCO_3 (bicarbonate) well (S-0034) has low normalized chloride (0.2820), but normalized HCO_3 (5.3002) far exceeds the ocean standard (0.0321) indicating the dominance of bicarbonate in this well. The mixed well (L-0059) has nearly equal normalized concentrations of chloride (2.3744) and HCO_3 (2.5189).

USGS chemical classification can also be identified by normalized chloride alone (Table 3-4). NaCl wells have normalized chloride above 3.0, while bicarbonate wells have normalized chloride less than 0.5. Normalized chloride is a useful proxy for salt composition because its value represents the proportion of chloride relative to the other major ionic salt species.

Maucha Diagrams

Normalized concentrations are effectively visualized by use of the Maucha diagram (Maucha 1932). The Maucha diagram, also known as an ionic polygonic diagram, is an eight-sided polygon representing the relative concentrations of the major cations and anions in a water-quality sample (Wetzel 2001). The Maucha diagram is commonly used for comparing the relative abundance of ions for locations with widely varying concentrations (Silberbauer and King 1991). Differences between the USGS chemical classes for the three wells discussed above are apparent when using Maucha diagrams (Figure 3-4). The salt composition of S-0025, with its NaCl classification, is similar to ocean water because of the presence of relict seawater.

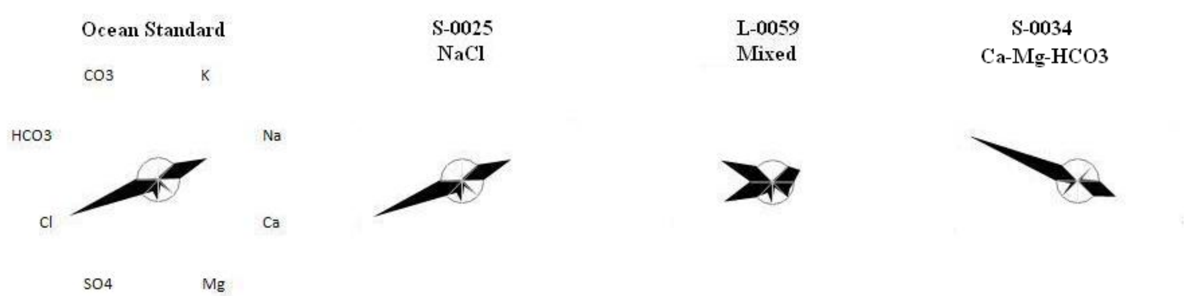


Figure 3-4. Maucha diagram comparing the eight dominant components of ocean water in three Upper Floridan aquifer wells depicting USGS chemical classification.

Relative Standard Error for Normalized Chloride

RSE of normalized chloride ranges from 0.51 to 1.36% (Table 3-4). *RSE* of normalized chloride is calculated for wells with sufficient record length. Most wells have an acceptable number of samples for which concentrations of all eight salt ions were measured; but three wells (L-0455, S-0034, and V-0165) have an insufficient number of samples. The sample counts for this dataset are less than the sample counts previously shown in Table 3-1, because fewer samples include all eight salt ions simultaneously.

The low *RSE* values for normalized chloride show that relative abundance of chloride exhibits low temporal variability. The low temporal variability of normalized chloride means that salt composition, in addition to chloride concentration, is stable in these wells. This conclusion is supported by the strong correlation between normalized chloride and USGS chemical class. Because normalized chloride is stable, then chemical class, and salt composition, is also stable.

The stability of salt composition means that chloride-salinity relationships for Upper Floridan aquifer groundwater are essentially constant at any given location. This result simplifies the conversion of chloride to salinity for use in the surface water model. The next section provides even greater simplification by showing that the chloride-salinity relationship is also spatially constant for diffuse groundwater dominated by relict seawater.

Table 3-4. Upper Floridan aquifer wells near the middle St. Johns River sorted by mean of normalized chloride, with relative standard error (*RSE*) for normalized chloride, sample count, and USGS chemical classification.

Well	Normalized Chloride		Sample Count	USGS Chemical Class
	Mean	*RSE* (%)*		
L-0455	0.26	—	1	$Ca-Mg-HCO_3$
S-0034	0.28	—	1	$Ca-Mg-HCO_3$
V-0165	0.46	—	3	$Ca-Mg-HCO_3$
S-1397	1.36	1.34	23	$Ca-Mg-HCO_3$
L-0059	2.37	1.36	18	Mixed
L-0032	3.36	1.00	14	Na-Cl
V-1091	3.45	0.53	18	Na-Cl
V-0240	3.58	1.17	12	Na-Cl
V-0818	3.74	0.70	14	Na-Cl
BR1526	3.77	0.51	16	Na-Cl
V-0083	4.26	0.51	14	Na-Cl
S-0025	4.36	0.73	14	Na-Cl

* *RSE*: relative standard error

3.5 Sources of Salts

Introduction

The source of salts being transported to the river by groundwater affects the salt composition of the inflowing groundwater. Near the St. Johns River, salts in groundwater are derived from three main sources: relict seawater, aerosols dissolved in rainwater, and dissolution of minerals. Relict seawater was previously shown to be the primary contributor of chloride to the middle and upper St. Johns River, and has a salt composition nearly identical to seawater. Sea spray is transported inland as an aerosol mixed with rainwater. Upchurch and Randazzo (1997) note that Florida rainfall is a very dilute mixture that has the ionic proportions—but not the concentrations—of seawater. As the rainfall infiltrates the soil, the chloride concentration can increase by evaporation and evapotranspiration. Although this water has the same salt composition as seawater, chloride concentrations derived from aerosols are low, generally less than 10 mg L^{-1}. Dissolution of minerals also contributes salt ions as groundwater passes through a recharge area. Charged with carbon dioxide, rainfall is naturally acidic and can dissolve minerals as it infiltrates the soil zone and into the aquifer of a recharge area (Back and Hanshaw 1971). The source of salts for dissolution depends on the minerals present in the soils through which the groundwater moves. In Florida waters, the predominant sources for dissolution are calcite ($CaCO_3$), dolomite ($CaMg\ (CO_3)_2$), and gypsum ($CaSO_4$) (Randazzo and Jones 1997).

The chloride-salinity relationship of groundwater varies depending on the source of salts. Groundwater dominated by relict seawater has a chloride-salinity relationship close to that of seawater. Groundwater dominated by dissolution, however, may have appreciable salinity but no chloride at all. Therefore, understanding the sources of salts in groundwater entering the river, in different locations, is essential to convert groundwater chloride to salinity for surface water modeling.

The relative contribution of salt sources is analyzed by comparing chloride to other salt ions. Chloride is a tracer of the extent of seawater present in water-bearing formations (Stringfield 1966) because it is always present in seawater and is nonreactive (conservative) as it is transported through an aquifer (Hem 1992). When dissolution is small, the ratios of chloride to other salt ions remain the same as seawater; when dissolution is large, these ratios decrease as other the concentrations of other salt ions increase while chloride remains constant.

Salt sources in groundwater underlying the middle St. Johns River are examined using several types of chloride ratios. First, the observed ratios of chloride with four other variables, (a) specific conductance, (b) magnesium, (c) sulfate, and (d) calcium, are compared with the theoretical ratio for seawater, called the seawater line. Comparison of chloride ratios with the seawater line is used to identify the dominant source of salt ions as either relict seawater or dissolution. A second analysis examines the relationship between chloride and the ratio of monovalent to divalent ions, the M:D ratio. This analysis shows that both M:D ratio and chloride concentration determine the USGS chemical classification of groundwater in the groundwater modeling study area. The seawater line and M:D ratio analyses identify a chloride level of 250 mg L^{-1} for delimiting groundwater dominated by relict seawater. This delimitating chloride level identifies diffuse groundwater sources to the river as relict seawater, which implies a constant chloride-salinity relationship for diffuse groundwater entering the river.

The results of this section show that chloride concentration alone is a predictor of salt composition of Upper Floridan aquifer groundwater in the groundwater modeling study area. From this result, we show that relict seawater is the dominant source of salts entering the middle St. Johns River as diffuse groundwater, which means that the chloride-salinity relationship for diffuse groundwater is constant across the groundwater modeling study area.

Comparison of Chloride to the Seawater Line

A seawater line shows the constant ratio of chloride to another variable for seawater. A plot of observed ratios against the seawater line shows the extent that salt composition deviates from that of pure seawater. Significant deviation of salt ions from the seawater line indicates that dissolution is an important source of salts. This method for characterizing salt sources was previously used to analyze wells in northeast Florida (G. Phelps 2001) and southwest Florida (Sacks and Tihansky 1996, Torres 2001).

Mean chloride and specific conductance are compared against the seawater line for the 12 observation wells in Figure 3-5a. Three different symbol types are used to identify each well by USGS chemical classification. The NaCl wells and the mixed well fall closely on the seawater line. Since salinity is defined by specific conductance (Lewis and Perkin 1978), this shows that the chloride-salinity relationship is nearly identical to seawater when groundwater chloride exceeds about 250 mg L^{-1}. Use of the chloride-salinity relationship for seawater would under predict salinity of the bicarbonate wells, however, for which mean chloride is less than 100 mg L^{-1}.

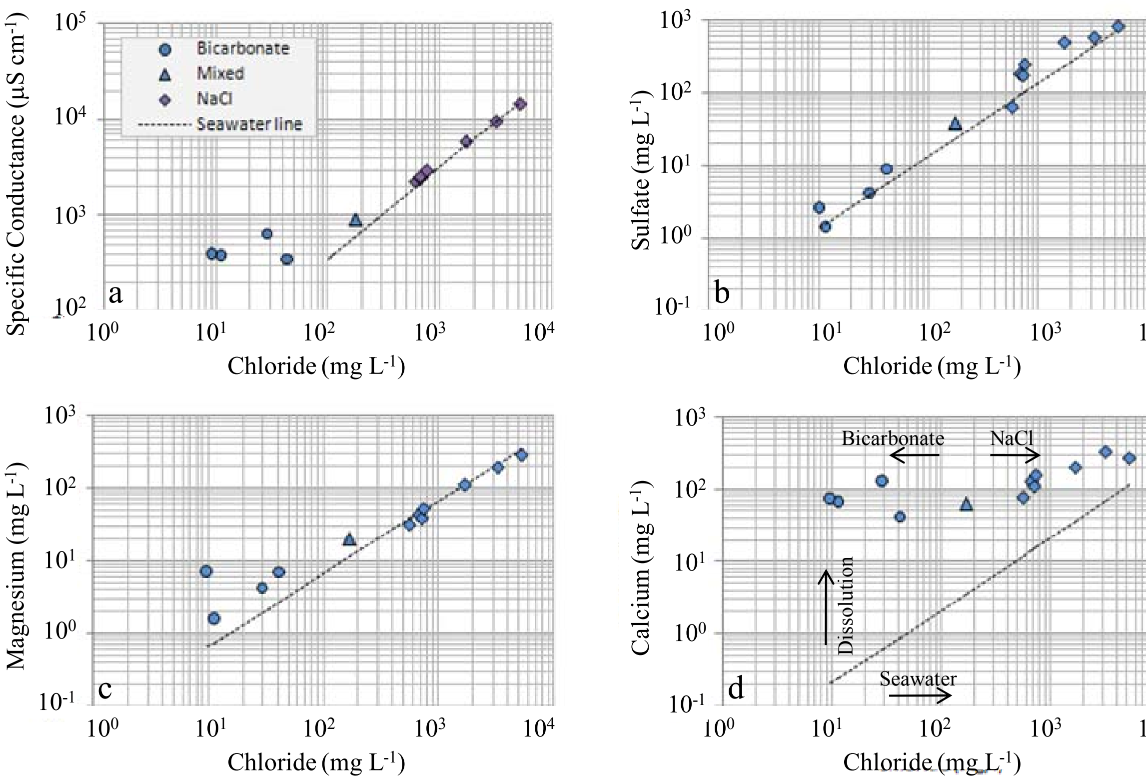


Figure 3-5. Ratios between chloride and specific conductance (a), SO_4 (b), Mg (c), and Ca (d). The dashed "seawater line" is the hypothetical concentration for seawater.

Comparison of chloride ratios against the seawater line for the divalent salt ions of sulfate (SO_4), magnesium (Mg), and calcium (Ca) indicate the source of these divalent ions. Observations falling above the seawater line indicate addition of salts from dissolution (Sacks and Tihansky 1996).

SO_4 has the highest relative abundance of any divalent species in ocean water, and it is abundant in the wells where relict seawater is the major influence (see Figure 3-5b). A point above the seawater line indicates a fraction of SO_4 is derived from dissolution, likely from deposits of gypsum ($CaSo_4$) (Torres 2001). The similarity of $Cl:SO_4$ ratios to that of seawater for the bicarbonate wells could indicate ocean aerosols as the source of SO_4 to these wells. (A thorough analysis of SO_4 sources in these wells is beyond the scope of this work.)

Mg behaves similarly to SO_4 and converges with the seawater line with increasing chloride concentration (see Figure 3-5c). Mg in the bicarbonate wells falls above the seawater line indicating dissolution, likely from dolomite ($CaMg\ (CO_3)_2$.

Ca is almost a trace element in ocean water, so observed data are positioned above the seawater line in all wells, indicating dissolution is the predominant source of calcium (see Figure 3-5d). Ca concentration remains remarkably constant and independent of chloride across wells. The constancy of Ca could mean that Ca concentrations are controlled by the degree of saturation of well water with calcite ($CaCO_3$) and dolomite.

Comparisons of Chloride to the M:D ratio

Comparing chloride to the M:D ratio illustrates that sources of salts remain stable across the groundwater modeling study area. The M:D ratio is the sum of monovalent sodium (Na) and potassium (K) ions divided by the sum of divalent calcium (Ca) and magnesium (Mg) ions (Wetzel 2001). A plot of chloride and M:D ratios for all 148 individual samples across all 12 wells shows a strong correlation between chloride and M:D ratio (Figure 3-6). Because the temporal variability of chloride within wells is low, observations in individual wells form clusters of points. The clustering of points within each well is additional evidence of the stability of salt composition discussed in the previous section. M:D ratios are readily aligned with USGS chemical classification as shown on the y-axis of the plot.

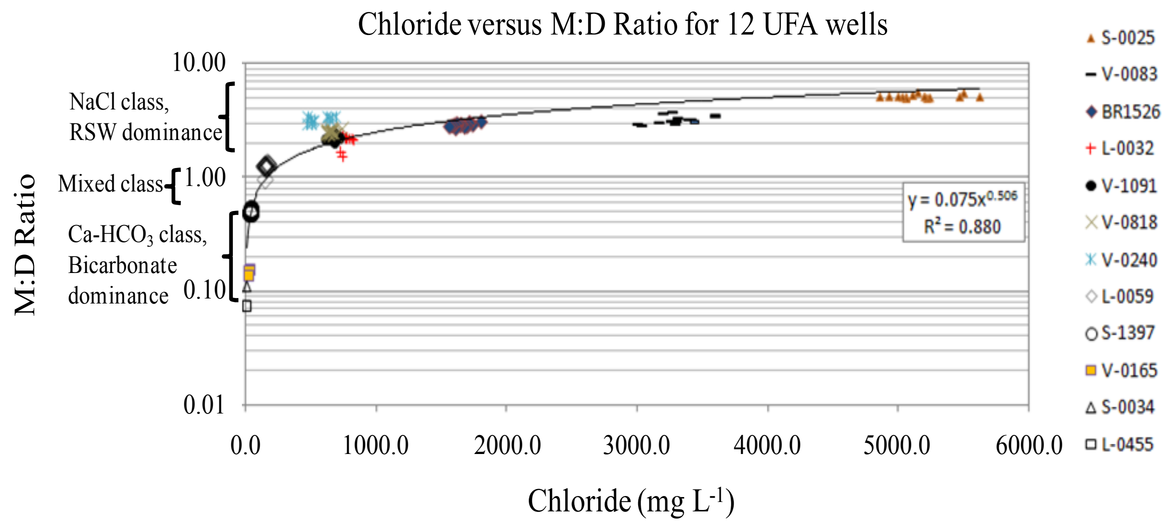


Figure 3-6. Observed chloride versus the M:D ratio for 12 Upper Floridan aquifer wells and power-fitted trend line.

The relationship between chloride and M:D ratio follows a power law and approaches the asymptotic limit of seawater for increasing chloride. The observations are described by M:D = $0.075Cl^{0.506}$, with $r^2 = 0.88$. The strong correlation between chloride and M:D ratio across wells shows that chloride concentration (even a single observation) can be used to determine the M:D ratio (a proxy for salt composition) and the USGS chemical classification. Key to this is the stability of salt composition in these wells. NaCl waters, having salts dominated by relict seawater, are found in the groundwater modeling study area wherever chloride is greater than 250 mg L^{-1}.

Chloride Concentrations in Diffuse Groundwater

Although chloride varies widely in observation wells across the groundwater modeling study area (see Table 3-2), diffuse groundwater flow to the river is dominated by relict seawater because chloride levels exceed 250 mg L^{-1} in nearly all areas directly underlying the river channel (Figure 3-7). Maucha diagrams for the 12 wells overlain on a map of chloride distribution in the Upper Floridan aquifer explain the occurrence of bicarbonate wells near the river. The chloride map is a synthesis of existing maps developed by spatial interpolation of observed Upper Floridan aquifer chloride in previous studies by Tibbals (1990) and Shampine (1975). The four bicarbonate wells all lie on the margins of the steep lateral gradient of Upper Floridan aquifer chloride underlying the river channel. Well S-0034 is close to Lake Harney but lies at the base of the Geneva Hill. Freshwater underlying Geneva Hill forms a convex bowl of freshwater in the Upper Floridan aquifer surrounded by waters of higher chloride on all sides. Most of the river channel, however, has chloride above 250 mg L^{-1}. The Maucha diagrams show that all wells falling in these areas are relict seawater-dominated waters with salt composition equivalent to seawater.

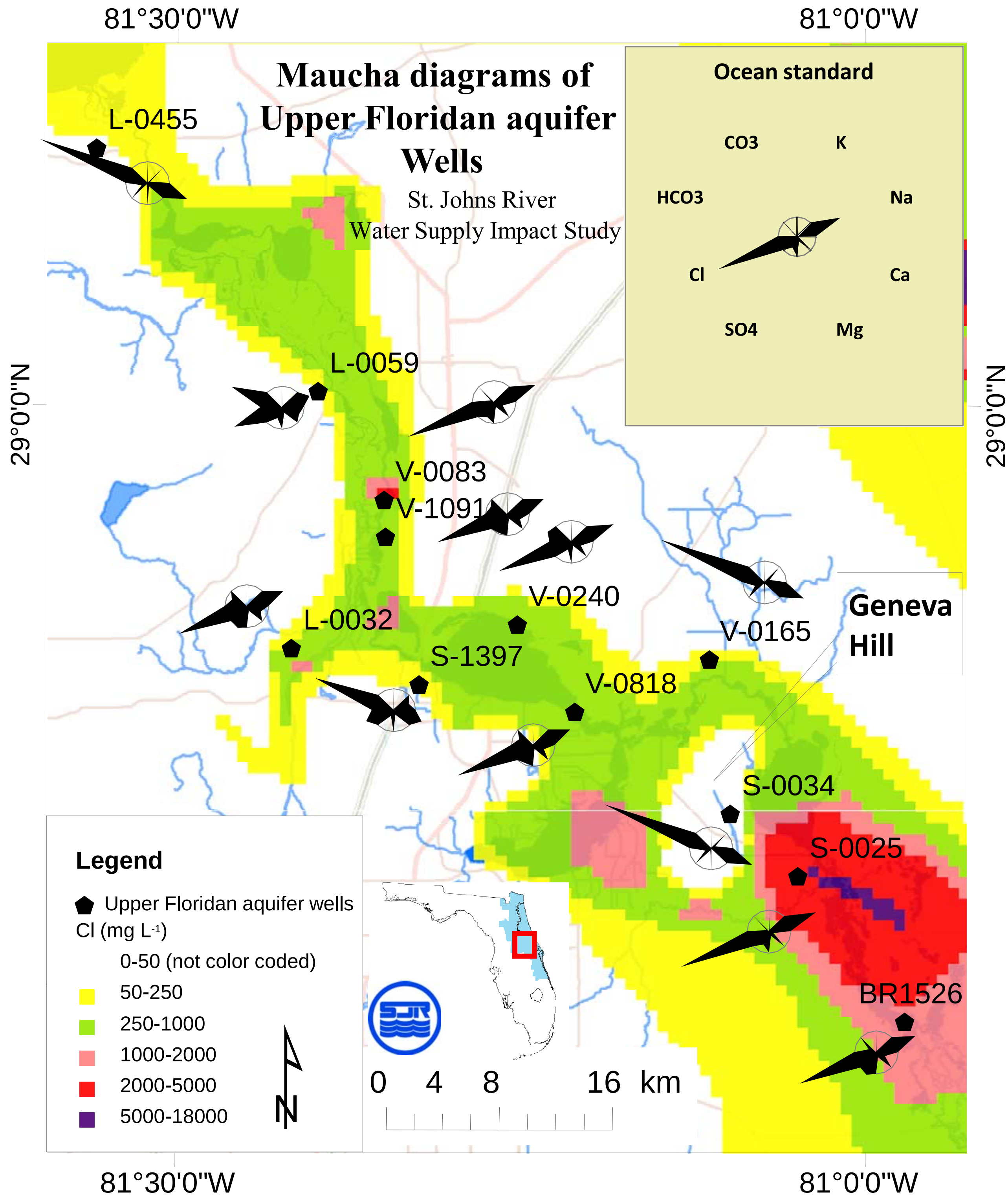


Figure 3-7. Maucha diagrams for 12 monitoring wells near the middle St. Johns River, with well locations and underlying Upper Floridan aquifer chloride concentration.

3.6 Importance of Density Gradients to Groundwater Circulation

Introduction

Groundwater modeling is often simplified by assuming a constant-density fluid. No simple guidelines exist that predetermine whether a groundwater flow simulation should be treated in a density-dependent or density-independent manner (Post, Kooi and Simmons 2007). This determination is made here by comparison of density-independent and density-dependent vertical discharge estimated from monitoring-well data. Density-independent discharge is estimated as a linear function of the difference between river stage and Upper Floridan aquifer hydraulic head. Density-dependent discharge is estimated from equivalent freshwater head (EFH) following the methodology outlined by Post et al. (2007). The small percent differences between calculated density-independent and density-dependent groundwater discharge show that the assumption of constant density for groundwater modeling is reasonable for simulating groundwater discharge to the middle St. Johns River. This conclusion primarily applies to the middle St. Johns River because the Upper Floridan aquifer hydraulic head is generally much larger than the river stage.

Selection of Monitoring Wells

Density-independent and density-dependent groundwater discharge deviate most in areas having the largest vertical density gradients. Vertical density gradients are largest where the underlying groundwater has the highest chloride concentration. The three monitoring wells with the highest chloride concentration, S-0025, V-0083, and BR-1526, were selected for this analysis. Chloride concentrations in these wells range from 2,035 to 5,220 mg L^{-1} (Table 3-5). For comparison, mean chloride within the river is 200 mg L^{-1} and ranges to about 1,000 mg L^{-1}. Chloride and density in these three wells is always higher than chloride and density of the river. The focus on these three wells for the analysis is conservative in that density effects on groundwater discharge are most significant at these locations.

Selection of Sample Well Data

For each of the three wells, coincident measurements of aquifer hydraulic head, river stage, and observed well chloride were obtained for the single day of maximum chloride. Alternatively, mean values could have been used, but use of maximum values is more conservative. The days of observations are 22 May 1995, 19 March 2002, and 17 April 1998, for wells S-0025, V-0083, and BR-1526, respectively.

Calculation of Equivalent Freshwater Head (EFH)

EFH is calculated in Equation 3-1 using observed data from each monitoring well (Table 3-5). EFH is the level to which water would rise in a piezometer if it were completely fresh instead of brackish.

$$\mathbf{EFH} = \mathbf{h_f} = \frac{\rho}{\rho_f}\mathbf{h} - \frac{\rho - \rho_f}{\rho_f}\mathbf{z} \tag{3.1}$$

In Equation 3-1, h is the observed Upper Floridan aquifer hydraulic head, z is elevation of the midpoint of the well screen (i.e., water sampling point), and ρ and ρ_f are the densities of the brackish groundwater and freshwater, respectively. All elevations are measured relative to the

National Geodetic Vertical Datum of 1929 (NGVD29). The density of the groundwater (ρ) was calculated from observed chloride concentration using a density concentration ratio (Guo and Langevin 2002). The density of river water is conservatively assumed that of pure freshwater (1,000 kg m^{-3}). Figure 3-8 is a schematic of variables used in Equations 3-1, 3-2, and 3-3.

Table 3-5. Calculation of equivalent freshwater head (EFH) in three wells near the middle St. Johns River using Equation 3-1. (Well depth is included for reference but not used in the calculation.)

Well ID	Well Depth (m)	Chloride Concentration (mg L^{-1})	ρ (kg m^{-3})*	h (m, NGVD29)	z (m, NGVD29)	h_f (m, NGVD29)
S-0025	47	5,220	1,006.53	2.27	-29.26	2.476
V-0083	132	4,047	1,005.06	2.03	-75.59	2.423
BR-1526	91	2,035	1,002.54	5.39	-78.90	5.604

*ρ_f is assumed 1,000.00 kg m^{-3}.

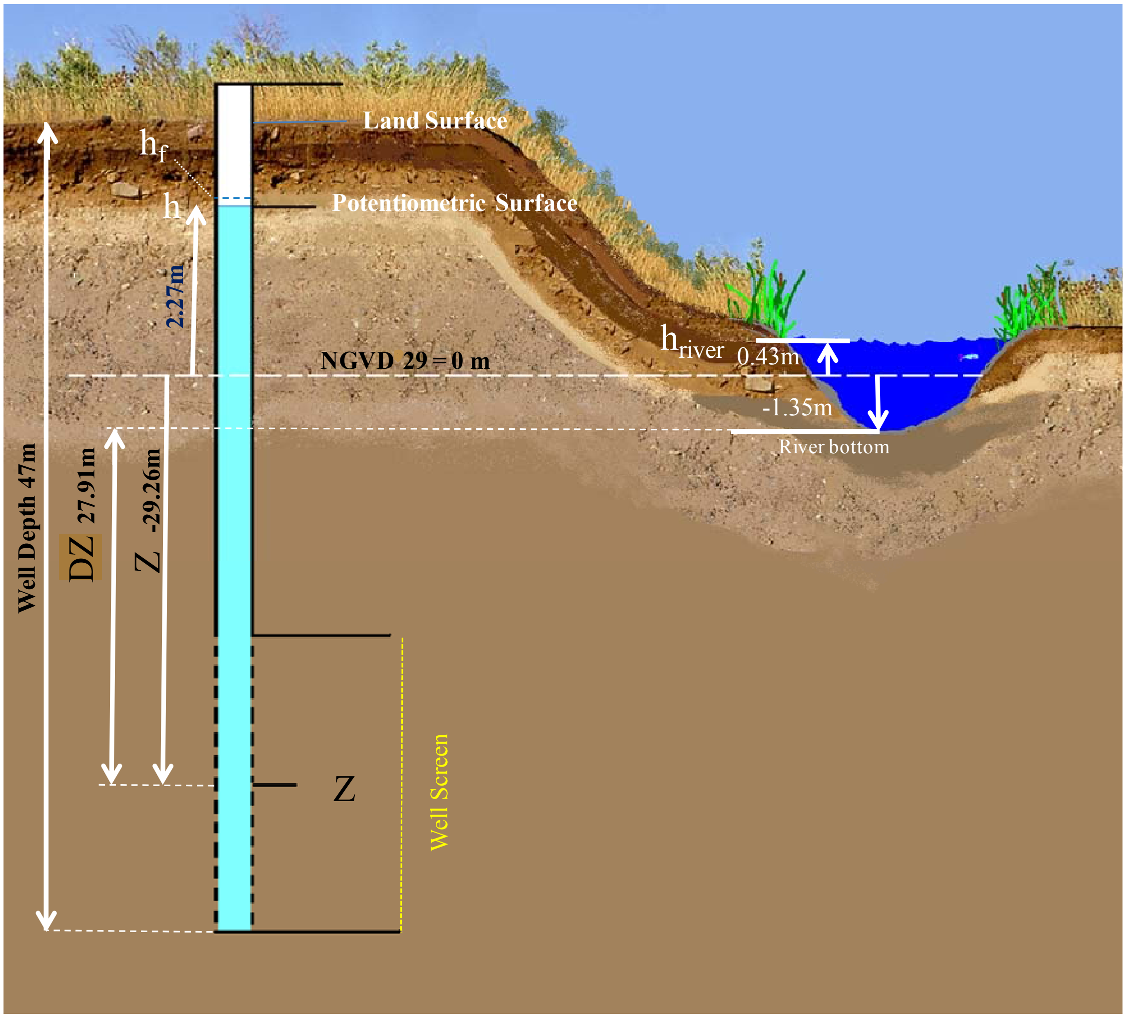


Figure 3-8. Schematic representation of the physical measurements used to estimate Equivalent Freshwater Head (EFH) and vertical groundwater discharge using Equations 3-1 to 3-3. Example values shown are for well S-0025. All elevations are in m, NGVD29.

Calculation of Density-Independent Discharge

In Equation 3-2, density-independent groundwater discharge (q_{ind}) is calculated as a linear function of the difference between river stage (h_{river}) and Upper Floridan aquifer hydraulic head (h):

$$q_{ind} = -C \cdot (h_{river} - \mathrm{h}) \tag{3.2}$$

where C is the conductance of the riverbed ($m^2\ s^{-1}$), obtained from the nearest river cell in the east-central Florida groundwater model (McGurk and Presley 2002). A positive q_{ind} indicates flow from the Upper Floridan aquifer into the river.

Calculation of Density-Dependent Discharge

Density-dependent groundwater discharge (q_{dep}) is calculated from the difference between river stage (h_{river}) and EFH (h_f), with a correction for negative buoyancy, following Post et al. (2007):

$$\boldsymbol{q_{dep} = -C \cdot \left[\left(h_{river} - h_f\right) + \Delta z\left(\frac{\rho_a - \rho_f}{\rho_f}\right)\right]} \qquad (3.3)$$

River stage (h_{river}) and conductance of the riverbed (C) are as defined for Equation 3-2. Conductance values are applicable for calculating density-dependent discharge because the east-central Florida groundwater model was calibrated against head and spring discharge. EFH (h_f) is used in place of Upper Floridan aquifer hydraulic head (h) to correct for the effect of density on the observed hydraulic head difference driving the vertical groundwater discharge. The term $\Delta z\left(\frac{\rho_a-\rho_f}{\rho_f}\right)$ is an additional gravity force (negative buoyancy) resulting from the increase in density above freshwater (Apel 1987). Δz is the distance between the well sampling point and the river bottom, and ρ_a is the average density between these points.

The average density, ρ_a, is not known for these locations because vertical profiles of groundwater density are not measured. A reasonable assumption is a linear profile between the well sampling point (ρ) and river bottom ρ_f. The average of a linear profile is simply the mean of the two end points, so $\rho_a = ½(\rho + \rho_f)$.

Comparison of Density-Independent and Density-Dependent Discharge

Density-independent (q_{ind}) and density-dependent (q_{dep}) vertical discharge for the three wells is compared in Table 3-6. The calculated discharge between wells is into groundwater model cells of varying areas and cannot be directly compared. Here we are interested in the percent difference between density-independent and density-dependent discharge at each well location, defined as $\Delta_q\% = 100 \times \frac{q_{dep}-q_{ind}}{q_{ind}}$. Δ_q% ranges from 2.8–13.0% in the three wells.

Table 3-6. Calculation and comparison of density-independent and density-dependent groundwater discharge near three monitoring wells using Equations 3-2 and 3-3.

Well ID	C ($m^2\ s^{-1}$)	Δz (m)	ρ_a (kg m^{-3})	h_{river} (m, NGVD29)	Hydraulic head (m, NGVD29)		Calculated groundwater discharge ($m^3 s^{-1}$)		Δ_q%
					h	h_f	q_{ind}	q_{dep}	
S-0025	0.081	27.91	1,003.27	0.43	2.270	2.476	0.1490	0.1583	6.3
V-0083	0.054	72.59	1,002.53	0.42	2.030	2.423	0.0869	0.0982	13.0
BR-1526	0.004	79.36	1,001.27	0.88	5.390	5.604	0.0180	0.0185	2.8

Factors Controlling Discharge Differences

An equation for Δ_q% in terms of the well variables is readily found by substituting Equations 3-2 and 3-3 in the relation $\Delta_q\% = 100 \times \frac{q_{dep}-q_{ind}}{q_{ind}}$. This substitution results in the following equation for Δ_q%:

$$\Delta_{\mathbf{q}}\% = \mathbf{100} \times \frac{\left(\frac{\boldsymbol{\rho}-\boldsymbol{\rho}_{\mathbf{f}}}{\boldsymbol{\rho}_{\mathbf{f}}}\right)\times(\mathbf{h}-\mathbf{z})-\Delta\mathbf{z}\times\left(\frac{\boldsymbol{\rho}-\boldsymbol{\rho}_{\mathbf{f}}}{2\boldsymbol{\rho}_{\mathbf{f}}}\right)}{(\mathbf{h}-\mathbf{h}_{\mathbf{river}})} \tag{3-4}$$

Equation 3-4 also uses the assumption that $\rho_a = ½(\rho + \rho_f)$. Since h is the hydraulic head at the well location, then by definition, h – z is the pressure head of the well (the length of the column of fluid inside the well).

For the middle St. Johns River, the pressure head in the well (h–z) has nearly the same magnitude as Δz. With the assumption that these quantities are equal, Equation 3-4 is simplified as follows:

$$\Delta_{\mathbf{q}}\% = \mathbf{100} \times \frac{\left(\frac{\boldsymbol{\rho}-\boldsymbol{\rho}_{\mathbf{f}}}{2\boldsymbol{\rho}_{\mathbf{f}}}\right)}{\left(\frac{\mathbf{h}-\mathbf{h}_{\mathbf{river}}}{\Delta\mathbf{z}}\right)} \tag{3-5}$$

Equation 3-5 shows the difference between the density-independent and density-dependent flow calculations does not depend on the values of conductance (C) or choice of vertical datum. Δ_q% is made large when either $(\rho–\rho_f)/2\rho_f$ is large or $(h – h_{river})/\Delta z$ is small. For typical densities of groundwater underlying the middle St. Johns River, the numerator is not larger than 0.0033. Given that Δz ranges from 30 to 80 m and is constant at any given location, then the percent error in Equation 3-5 is primarily driven by the hydraulic head difference between the aquifer and river. For maximal values of ρ (1,006 kg m^{-3}) and Δz (80 m), the hydraulic head difference would need to be less than 1.2 m for Δ_q% to be greater than 20%. In general, hydraulic head differences along the middle St. Johns River are 2.5 to 5.5 m.

Large Δ_q% can only occur when aquifer hydraulic head and river stage are close in value. This situation can occur in areas near Lake Harney where the aquifer hydraulic head exhibits a relative minimum and river stage is large during high river flow (Figure 5-1). Still, large Δ_q% only occurs in a relatively small fraction of the total groundwater modeling study area, and only then during infrequent high flow events when the contribution of groundwater discharge is relatively small. In addition, the largest percent differences occur when groundwater discharge is lowest because hydraulic head differences driving the groundwater discharge are small. For these reasons, we conclude that vertical groundwater discharge can be reasonably simulated under the assumption of constant density.

3.7 Summary

Although Upper Floridan aquifer chloride varies widely across the study area, temporal variability of chloride is extremely small. Chloride observations within 12 observation wells range from 8 to 5,798 mg L^{-1}, but the temporal variability within an individual well is less than 3%. The stability of groundwater chloride means that chloride concentrations can be assumed temporally constant for estimating chloride loads to the river. Chloride loads are then simply estimated as the product of simulated groundwater discharge, and obtained from groundwater modeling and a constant, (though spatially varying) chloride concentration.

In addition to chloride, salt composition is also stable within the Upper Floridan aquifer. This result allows the conversion of chloride to salinity for input to a surface water model, because the chloride-salinity relationship is constant over time at any given location.

Salt composition and USGS chemical class of groundwater is completely determined from chloride concentration alone in the study area. Groundwater having chloride greater than 250 mg L^{-1} falls in the NaCl class, has salts dominated by relict seawater, and has a salt composition nearly identical to that of seawater. Because chloride is above 250 mg L^{-1} in nearly all areas beneath the middle St. Johns River channel, then Upper Floridan aquifer groundwater entering the river as diffuse groundwater discharge also has a salt composition of seawater. The chloride-salinity relationship is constant for all diffuse groundwater sources to the middle St. Johns River.

Vertical density gradients have only a minor effect on calculated vertical groundwater discharge to the river, primarily because the difference between the Upper Floridan aquifer hydraulic head and river stage is relatively large. The percent difference between density-dependent and density-independent groundwater discharge ranged from 2.8 to 13.0% within three wells with the largest expected difference indicating that density differences between brackish groundwater and fresh river water do not appreciably increase groundwater discharge to the river.

Although an important use of groundwater modeling is to estimate chloride loads from diffuse groundwater to the middle St. Johns River, this model's goal does not require direct use of a transport model. One reason for neglecting chloride transport processes is that groundwater chloride is temporally stable. Chloride loads are estimated by the simple product of simulated-groundwater discharge and observed chloride concentration. A second reason is that density differences resulting from spatial chloride variability are shown to be unimportant in the simulation of vertical groundwater discharge. The use of constant-density groundwater flow models is well suited to application in this groundwater model study area.

4 GROUNDWATER MODELING

4.1 INTRODUCTION

Groundwater modeling has two main goals for the WSIS. The first goal is to provide boundary conditions for a surface water model of the middle St. Johns River. These boundary conditions are developed from simulated groundwater discharge and its associated chloride load. Groundwater discharge is difficult to measure by direct observation, so groundwater modeling is a viable alternative. The second goal is to test whether increased groundwater discharge and chloride load would appreciably alter river conditions if water levels decline due to surface water withdrawals. These tests are made within a larger study area where surface water withdrawals are proposed.

The study area for the groundwater modeling is the middle St. Johns River, lower portions of the upper St. Johns River, and the lower Ocklawaha River. The lower St. Johns River, Lake George, and upper portions of the upper St. Johns River were excluded from the study area for two reasons: first, the interaction between Upper Floridan aquifer and surface waters is weak in these areas (Motz and Sedighi 2008), and second, possible changes to water levels by water withdrawals are small.

The groundwater modeling study described in this chapter uses two existing SJRWMD regional groundwater flow models: the north-central Florida model (NCF model) (Motz and Dogan 2004) and the east-central Florida model (ECF model) (McGurk and Presley 2002). NCF model covers the lower Ocklawaha River and a small portion of the northern middle St. Johns River; the ECF model covers the northern portion of the upper St. Johns River and most of the middle St. Johns River (Figure 4-1). Both models are steady-state, constant-density groundwater flow models based on the USGS modular three-dimensional finite difference groundwater flow model (MODFLOW) (McDonald and Harbaugh 1988). The models were calibrated to average-groundwater conditions for 1995 using groundwater levels in wells, water levels in the surficial aquifer system, and spring discharges. SJRWMD developed these models to predict the potential changes to aquifer levels and spring discharges due to future increases in pumping (McGurk and Presley 2002). The models were also used to provide simulated data to a regional optimization model for evaluating future scenarios of well locations, pumping rates, and alternative water sources which prevent water-deficit problems (Agyei, Munch and Burger 2005). Williams (2006) used the ECF model to estimate predevelopment aquifer conditions and calculate changes in evapotranspiration from surficial aquifers and recharge to the Upper Floridan aquifer caused by proposed pumping changes.

The groundwater models are used here to calculate steady-state groundwater discharge and chloride loads to the middle St. Johns River. Steady-state discharges are constant and unchanging in time. Actual groundwater discharge varies seasonally in response to wet and dry periods; the use of steady-state values assumes that the seasonal variability of discharge is of secondary importance compared with the average discharge. We test the reasonableness of this assumption in Section 5.

Because diffuse groundwater discharge is the dominant source of chloride to the middle St. Johns River, the groundwater modeling results are needed for defining boundary conditions for the application of the Environmental Fluid Dynamics Code model (EFDC model) to the middle St. Johns River (Hamrick 1992). Estimates of discharge and chloride load are developed for 16 subregions of the EFDC model from the groundwater modeling results.

Increases in discharge and chloride load due to water withdrawals are shown to be insignificant relative to the volume and chloride budgets of these systems.

Methods

Chloride load is estimated by multiplying simulated groundwater discharge with observed chloride concentration. Chloride concentrations are assigned to each model cell from existing contour maps interpolated from points of observed Upper Floridan aquifer chloride (Figure 2-2). This simple method of estimating chloride load is possible because of the stability of chloride underlying the river in the groundwater modeling study area.

For simplification, groundwater model results are aggregated into nine river segments (Table 4-1, Figure 4-1). The St. Johns River is divided into eight segments (B to I), while the lower Ocklawaha River contains a single segment (A). The three large middle St. Johns River lakes are found in segments D (Lake Monroe), E (Lake Jesup), and F (Lake Harney). The segment boundaries are split at locations of USGS gauges. This segmentation allows us to specify surface water profiles along each segment from observed river stage.

Table 4-1. River segments used for aggregation of groundwater modeling results. Segment boundaries are split at USGS gauging stations.

Segment ID	USGS Station Location*	USGS Gauge ID*	Groundwater Model Used
A	CR 316–SR 40†	02240000–02240500	NCF
B	SR 40‡–SR 44	02236125–02236000	NCF
C	SR 44–U.S. 17	02236000–02234500	ECF
D	U.S. 17–SR46J	02234500–02234435	ECF
E	Lake Jesup (SR46J) §	02234435	ECF
F	SR46J–SR46H††	02234435–02234000	ECF
G	SR 46–SR 50	02234000–02232500	ECF
H	SR 50–SR 520	02232500–02232400	ECF
I	SR 520–U.S. 192	02232400–02232000	ECF

* Location and gauge ID are ordered from downstream to upstream
†SR 40 at the Ocklawaha River
‡ SR 40 at the St. Johns River§**Segment E represents Lake Jesup where water level is represented by a single gauge at location SR46J
††SR 46 passes over the mouth of Lake Jesup (at location SR46J) and just south of Lake Harney (at location SR46H)

The groundwater models simulate discharge from springs throughout the WSIS study area. Here we focus on only six named springs: Mosquito and Ponce De Leon (segment B), Blue (segment C), Gemini and Green (segment D), and Clifton (segment E). These springs are close to the St. Johns River where the river can influence their pool elevation. They are included here so that changes in spring discharge caused by reduction of river stage by water withdrawals can be estimated. Several major springs are excluded from this analysis because they are located in tributaries that are unaffected by downstream-water withdrawals. Major springs excluded from the analysis (although they are included within existing groundwater models) include Wekiva, Blackwater, and Alexander springs in the middle St. Johns River and Silver Springs in the lower Ocklawaha River.

This section uses metric units with the exception of discharge, which is reported in million gallons per day (mgd) for consistency with other SJRWMD documents.

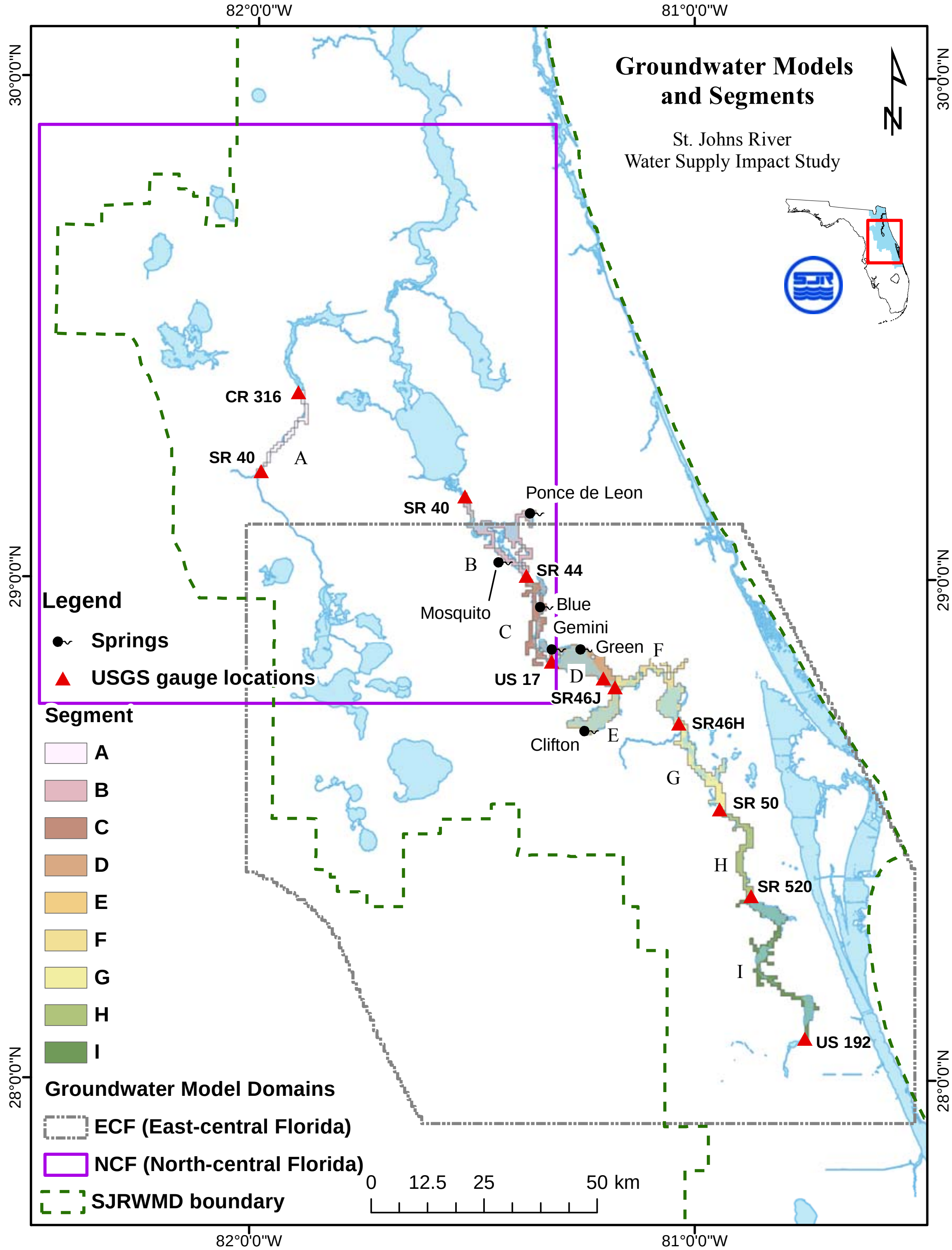


Figure 4-1. Spatial extent of the NCF and ECF groundwater flow models. Lettered segments are used to aggregate groundwater model results.

4.2 RESULTS

4.2.1 STEADY-STATE GROUNDWATER DISCHARGE AND CHLORIDE LOAD

Groundwater Discharge by River Segment

Simulated discharges range from 2.7 to 128.7 mgd over the nine river segments (Table 4-2). Diffuse groundwater discharge alone contributes a total of 211.1 mgd, primarily to the St. Johns River. The total spring discharge of 118.2 mgd is dominated by Blue Spring entering segment C, but recall that spring discharge reported here does not include all spring sources. Diffuse groundwater discharge declines in the upper St. Johns River south of SR 50 (H and I) as overburden over the Upper Floridan aquifer thickens. The lower Ocklawaha River (A) has the lowest diffuse groundwater discharge (2.7 mgd).

Table 4-2. Simulated discharge from springs and diffuse groundwater discharge, and total discharge of Upper Floridan aquifer water entering each river segment.

Discharge Source	Discharge to River Segments (mgd)*									Total Discharge
	A	B	C	D	E	F	G	H	I	
Springs	0.0	13.2	97.7	6.2	1.1	0.0	0.0	0.0	0.0	118.2
Diffuse groundwater	2.7	42.3	31.0	22.4	27.9	30.6	37.6	13.2	3.4	211.1
Total	2.7	55.5	128.7	28.6	29.0	30.6	37.6	13.2	3.4	329.3

* Segment A is in the lower Ocklawaha River; Segments B to I are in the St. Johns River.

We can easily compare the importance of diffuse groundwater discharge among segments by normalizing the absolute discharge by segment area. The normalized values have units of velocity and represent the segment-averaged, net upward velocity of diffuse groundwater (see Table 4-3). The smallest net upward velocity occurs in segments A and I. Net upward velocity is fairly uniform in the remaining segments (B to H), ranging from 0.13 to 0.31 cm d^{-1} with a mean of 0.18 cm d^{-1}.

Table 4-3. Diffuse groundwater discharge (mgd), segment area (km^2), and net upward velocity (cm d^{-1}) for each river segment.

River Segment*	A	B	C	D	E	F	G	H	I	Total
Diffuse groundwater discharge	2.7	42.3	31.0	22.4	27.9	30.6	37.6	13.2	3.4	211.1
Area	20.4	61.6	49.4	63.9	42.4	62.7	46.4	32.0	74.3	453.1
Velocity	0.05	0.26	0.24	0.13	0.25	0.19	0.31	0.16	0.02	0.18

* Segment A is in the lower Ocklawaha River; Segments B to I are in the St. Johns River.

Chloride Load by River Segment

The simulated groundwater discharges from the previous section are used to calculate chloride load. Chloride load from both diffuse groundwater and springs is calculated as the product of discharge and chloride concentration and is reported as metric tons per day (t d^{-1}). (A metric ton is 10% greater in weight than an English ton.) Chloride loads are calculated and reported separately for springs and diffuse groundwater discharge.

Chloride loads totaled 170 t d^{-1} for the six springs included in the groundwater modeling study (Table 4-4). Of these springs, Blue Spring dominates chloride load to the river because of its large discharge and moderate chloride concentration. For the smaller springs, chloride concentration is important to the spring's relative contribution of chloride load. Chloride load is five times larger for Green Spring than Clifton Spring, for example, even though their discharge is nearly identical.

Table 4-4. Discharge (mgd), chloride concentration (mg L^{-1}), and resultant chloride load ($t\ d^{-1}$) for the six springs included in the groundwater modeling study.

Spring	Mean Discharge	Mean Chloride	Chloride Load
Mosquito	1.1	7.9	0.03
Ponce DeLeon	12.1	128.5	5.9
Blue	97.7	401.9	148.6
Gemini	5.1	636.4	12.2
Green	1.1	606.8	2.6
Clifton	1.1	123.9	0.53
Totals	118.2	380.0*	170.0

*Discharge-weighted concentration of the six springs

Chloride load from diffuse groundwater discharge is calculated as the product of discharge (see Table 4-3) and segment-averaged chloride concentration (Table 4-5). Chloride load is normalized by segment area to produce a segment-averaged chloride flux. Chloride flux is useful for comparing the relative rate of chloride input to the river across segments.

Chloride flux is lowest in segments A and I, which also have the lowest net upward velocity of diffuse groundwater. Chloride flux is notably large (8.40 $g\ m^{-2}\ d^{-1}$) in segment G, between SR46H and SR 50. This segment has both the highest net upward velocity and largest chloride concentration among segments. Chloride flux is fairly uniform throughout the middle St. Johns River (B to F), ranging from 1.13 to 2.30 $g\ m^{-2}\ d^{-1}$.

Within the St. Johns River (B to I), chloride concentrations for diffuse groundwater exceed spring concentrations because some portion of the discharge from springs includes low-chloride waters from recharge areas distant from the high-chloride groundwater beneath the river (Toth 1999).

Table 4-5. Chloride concentration (mg L^{-1}), chloride load (t d^{-1}), and chloride flux (g m^{-2} d^{-1}) for diffuse groundwater discharge by river segment.

Chloride from Diffuse Groundwater Discharge	River Segment									Total[a]
	A	B	C	D	E	F	G	H	I	
Chloride concentration	26.3	709.1	731.1	853.5	682.7	1,246.6	2,751.4	1,002.1	716.0	1,173.0
Chloride load	0.27	113.3	85.9	72.3	72.0	144.3	389.9	50.2	9.3	937.4
Chloride flux	0.01	1.84	1.74	1.13	1.70	2.30	8.40	1.57	0.13	2.07

[a] Segment A is in the lower Ocklawaha River; Segments B to I are in the St. Johns River.

Comparison of Observed and Simulated Chloride Load for the Middle St. Johns River

The simulated chloride loads developed in the previous section are reasonable by comparison to observed chloride load. For comparison to simulated loads, observed chloride load was calculated for two river reaches: Reach 1, containing segment C, and Reach 2, containing segments D to F. These reaches had sufficient monitoring at upstream and downstream boundaries to estimate the chloride load entering and exiting each reach for the period of October 2000 through September 2002. Reach 1 is bounded by monitoring stations at SR 44 and U.S. 17, and Reach 2 is bounded by monitoring stations at U.S. 17 and SR46H. Chloride concentration was estimated from observed conductivity, and then chloride load from the product of chloride concentration and discharge. A chloride budget for each reach is completed by using previous estimates of chloride loads from springs and surface tributaries developed for the set-up of the EFDC model (see Chapter 5).

Observed total chloride load entering a reach from all local sources—diffuse groundwater, springs, and tributaries—is the difference between the chloride loads entering and exiting a reach across its upstream and downstream boundaries. The observed total chloride load was 366 t d^{-1} for Reach 1 and 360 t d^{-1} for Reach 2 over the observation period (Table 4-6). For comparison, simulated total chloride load is estimated by combining diffuse chloride load (see Table 4-5) with estimates of chloride load from all springs and tributaries entering the reach. Simulated total chloride load is about 17% greater than observed for Reach 1, and 7.5% greater for Reach 2. These percent differences are remarkably modest given the uncertainties in estimating the chloride budget components, including: (a) uncertainties in estimating discharge from ungauged tributaries, (b) lack of chloride observations in ungauged tributaries, (c) incomplete chloride observations in gauged tributaries, (d) temporal averaging of spring discharges, and (e) mismatch of time periods between observed total chloride load and the groundwater models.

Table 4-6. Comparison of simulated and observed chloride load to two middle St. Johns River reaches. Numbers in parentheses are the percent contribution of the major source of chloride to the total simulated chloride load.

Reach	Boundaries	Simulated Chloride Load (t d^{-1})[a]				Observed Chloride Load (t d^{-1})		
		Diffuse	Springs	Tributaries	Total Simulated	Inflow	Outflow	Total Observed
1	SR 44–U.S. 17	86 (20%)	289 (67.5%)	53 (12.5%)	428	1,799	1,433	366
2	U.S. 17–SR46H	289 (75%)	15 (4%)	83 (21%)	387	1,427	1,067	360

The relative importance of diffuse groundwater discharge to the total chloride load is illustrated by the percent contribution of the three major sources—diffuse groundwater, springs, and tributaries—to each reach. Discharge to Reach 1 is dominated by Blue Spring, so chloride load from springs is correspondingly high (67.5%) to this reach. Diffuse groundwater contributes 20% of the chloride load to Reach 1. Diffuse groundwater discharge is the dominate source of chloride (75%) to Reach 2, where springs are lacking. This reach encompasses all the areas of the middle St. Johns River upstream of Sanford (U.S. 17), including lakes Monroe, Jesup, and Harney. Because chloride load to Reach 2 is dominated by diffuse groundwater, accurate estimates of chloride load from groundwater are most important there. The good match of observed and simulated chloride load to Reach 2 provides assurance that the estimated chloride load from diffuse groundwater within this reach is reasonable.

4.2.2 EXAMINING THE EFFECTS OF WATER WITHDRAWALS

Discharge of groundwater from a confined aquifer to surface water is proportional to the difference between the potentiometric surface (hydraulic head) of the aquifer and the surface water stage, if the underlying soils are permeable enough to allow vertical flow of groundwater. The potentiometric surface of the Upper Floridan aquifer is higher than the mean stage of both the middle St. Johns River and lower Ocklawaha River, so that groundwater flows into these rivers from the underlying aquifer. Surface water withdrawals would reduce river stage, and increase groundwater discharge and the associated chloride load to the rivers. In this section, the ECF and NCF groundwater flow models are used to quantify the increases in groundwater discharge and chloride load caused by water withdrawals. Modeling results show that groundwater discharge and chloride load does increase under a withdrawal scenario, but these increases are small compared to the volume and chloride budgets of the rivers.

Sensitivity of Groundwater Discharge to River Stage

The model sensitivity of groundwater discharge to river stage was first calculated by simple adjustment of river stage over a 1 ft (30 cm) range by 0.25 ft (7.5 cm) increments. As expected, the model response to changes in stage displayed nearly perfect linearity; the use of multiple increments of stage served as a check on the model response.

The increase in discharge per unit reduction in river stage, ΔQ/ΔH, is calculated separately for springs and diffuse groundwater within each river segment and ranges from 0.034 to 0.273 mgd cm^{-1} (Table 4-7). The percent change of discharge for a 1 cm decline in river stage is a convenient method of normalizing ΔQ/ΔH for comparison between segments. Segments with the smallest discharge (A and I) are most sensitive to changes in stage in a relative sense.

Table 4-7. Increase in discharge per unit reduction in river stage (mgd cm^{-1}) for springs and diffuse groundwater discharge by river segment.

Source	Increase in Discharge to River Segment*									Total By Source
	A	B	C	D	E	F	G	H	I	
Springs	—	0.034	0.084	0.006	0.001	—	—	—	—	0.126
Diffuse groundwater	0.047	0.239	0.112	0.036	0.033	0.096	0.113	0.061	0.036	0.773
Total by segment	0.047	0.273	0.196	0.042	0.034	0.096	0.113	0.061	0.036	0.898
% Change for a 1 cm change in river stage	1.74	0.49	0.15	0.15	0.12	0.31	0.30	0.46	1.07	0.27

* Segment A is in the lower Ocklawaha River; Segments B to I are in the St. Johns River.

Sensitivity of Groundwater Chloride Load to River Stage

The increase in chloride load per unit reduction in river stage is calculated from simulated chloride loads (see Table 4-4 and Table 4-5) and ΔQ/ΔH (Table 4-8). Chloride loads from springs are relatively insensitive to decreasing river stage. Chloride loads from diffuse groundwater increase 3 t d^{-1} for a 1 cm decline in stage across all segments.

Table 4-8. Increase in chloride load per unit decrease in river stage (t d^{-1} cm^{-1}) for springs and diffuse groundwater by river segment.

Source	River Segment*									Total by Source
	A	B	C	D	E	F	G	H	I	
Springs	—	0.015	0.119	0.015	0.001	—	—	—	—	0.15
Diffuse	0.005	0.646	0.309	0.116	0.086	0.447	1.170	0.231	0.010	3.02
Total by segment	0.005	0.661	0.428	0.131	0.087	0.447	1.170	0.231	0.010	3.17

* Segment A is in the lower Ocklawaha River; Segments B to I are in the St. Johns River.

The Effect of Water Withdrawals on Groundwater Discharge and Chloride Load

The preceding section provides the sensitivity of groundwater discharge and chloride load to unit reductions in river stage across all segments. In this section, we use estimates of stage reductions from a proposed water withdrawal scenario to quantify how groundwater discharge and chloride loads would respond. The withdrawal scenario includes a 107-mgd withdrawal from the lower Ocklawaha River, a 55-mgd withdrawal from the upper St. Johns River, and a 100-mgd withdrawal from the middle St. Johns River. Stage reductions were previously estimated for these scenarios from two surface water models: Hydraulic Simulation Program–Fortran (HSPF model) for the lower Ocklawaha River and upper St. Johns River (see Chapter 3), and EFDC model for the middle St. Johns River (see Chapter 5).

In the lower Ocklawaha River, mean stage decreases 4.3 cm due to the 107-mgd withdrawal. The combined 155-mgd withdrawal in the middle and upper St Johns River decreases mean stage from 1 to 5.5 cm, with greater reduction in upstream segments.

In response to these stage reductions, groundwater discharge increases 0.2 mgd in the lower Ocklawaha River, a 7.4% increase in diffuse groundwater discharge, but less than 0.03% increase of mean river discharge. In the St. Johns River, groundwater discharge increases 2.4 mgd, a 0.7% increase of modeled groundwater discharge and less than 0.1% increase of mean river discharge.

Chloride load to the lower Ocklawaha River increases 0.02 t d^{-1}, less than 0.2% of the mean chloride load. In the St. Johns River, chloride load increases 12.6 t d^{-1}. Although this chloride load might seem large in some contexts, it is less than 1% of the mean chloride load passing DeLand at SR 44. The increased groundwater discharge and chloride load resulting from water withdrawals is small compared to the overall discharge and chloride budget of these river segments.

In the middle St. Johns River, chloride load from groundwater is insensitive to changes in water level caused by a water withdrawal scenario even though diffuse groundwater discharge is the primary source of chloride load to the system. This result is explained by the relative difference between the hydraulic head difference driving the groundwater discharge and the expected change in water level due to a water withdrawal.

The hydraulic head difference driving groundwater discharge to the river is the difference between the hydraulic head of the Upper Floridan aquifer and the river stage. In the middle St. Johns River, the hydraulic head of the Upper Floridan aquifer is about 3 to 6 m (10 to 20 ft) NGVD29, while river stage is about 0.6 m (2 ft). The hydraulic head difference, then, is typically 2.5 to 5.5 m. Because groundwater discharge is a simple linear function of hydraulic head difference, a 5-cm decline in river stage increases discharge only 1% to 2%. This simple estimate for discharge increase is consistent with the 0.6 to 2.1% increase obtained by groundwater modeling for segments B through F (Table 4-7). The small ratio of stage change to hydraulic head difference (< 0.02) causes the insensitivity of groundwater discharge to water withdrawals. This ratio would need to be an order-of-magnitude larger for a 5-cm reduction in river stage to appreciably affect the chloride budget of the river.

4.3 Application to Surface Water Modeling

Simulated chloride load from diffuse groundwater discharge is required as a boundary condition for the EFDC model because of its importance to the total chloride budget of the middle St. Johns River, and because there are no direct observations of chloride load . Chloride load could have been estimated by difference in a total chloride budget, but this methodology is compromised by dependence on the accumulated error of all components in the budget. In addition, the spatial resolution of chloride load estimated from a chloride budget would be coarse because it is limited to locations along the river with coincident observations of discharge and chloride concentration.

To resolve spatially varying groundwater chloride load, the EFDC model is subdivided into 16 subregions. Steady-state groundwater discharge and chloride load, derived from the groundwater models, are assigned to each subregion (Figure 4-2). Although the EFDC model subregions do not perfectly align with the five middle St. Johns River groundwater segments, in general, Segment B comprises subregions 1 to 5, Segment C comprises subregions 6 to 7, Segment D comprises subregions 8 to 9, Segment E comprises subregions 10 to 13, and Segment F comprises subregions 14 to 16.

The total area of the EFDC model subregions (127.8 km^2) is smaller than the total area of the five middle St. Johns River groundwater segments (280 km^2) because the surface water and groundwater models differ in resolution and configuration. The EFDC model was configured to the geometry of the open waters of the middle St. Johns River and resolves the narrow channels between lakes. The groundwater models were configured to the geometry of the major features of the river, such as lakes Monroe, Harney, and Jesup, but the groundwater model grids are too coarse to resolve fine-scale open water areas. As a result, a groundwater model cell can include both open water and adjacent floodplain or uplands. The differences in model resolution require that output from the groundwater models be adjusted to properly represent diffuse groundwater flows through the EFDC model cells.

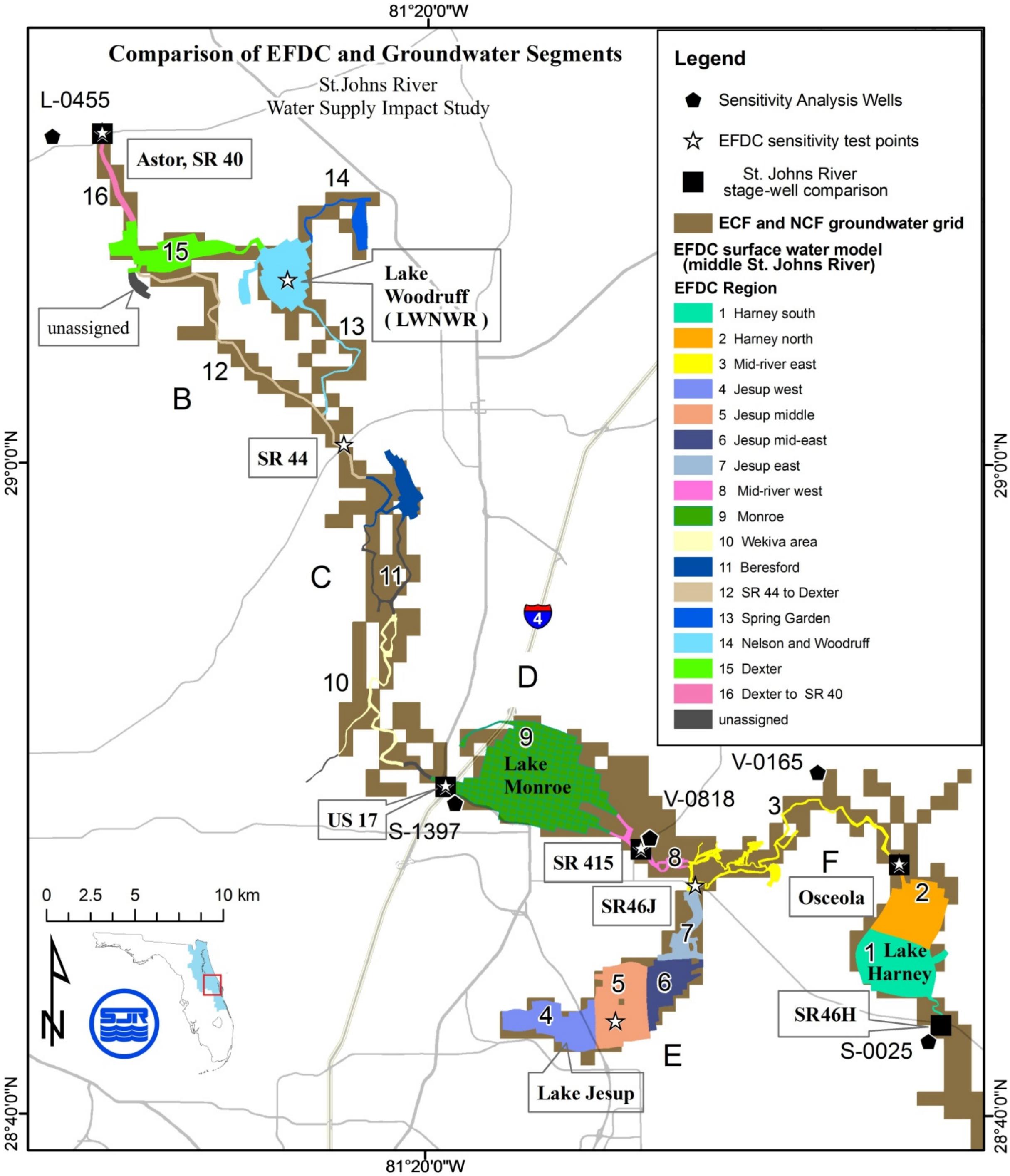


Figure 4-2. Sixteen subregions used to input groundwater discharge and chloride load to the EFDC model. Grid compared with five segments used by ECF and NCF groundwater models. The wells and surface water locations are referred to in Section 5.

Groundwater chloride concentrations are adjusted when transferring groundwater output to the EFDC model. Adjustment of chloride is necessary because the actual chloride concentration under the river channel is, in general, higher than the spatially averaged concentration supplied by the groundwater model. Chloride concentration is as important as chloride load for surface water modeling because river chloride levels (in the absence of significant evaporation) are limited by the level of the most concentrated source. Chloride concentrations are increased in certain EFDC model subregions, while chloride load is held constant by a compensating reduction of discharge.

The constant, steady-state values for diffuse groundwater discharge, chloride load, and salinity assigned to each of the 16 EFDC model subregions are shown in Table 4-9. Salinity is the level matching the chloride load based on the chloride-salinity relationship of seawater: S = 1.80655 x Cl x 0.001, where S is salinity and Cl is chloride in mg L^{-1} (Stumm and Morgan 1981). The total chloride load of 460.0 t d^{-1} is slightly lower than the 487.7 t d^{-1} provided by the groundwater modeling. The difference is caused by adjustment of chloride concentration input to the EFDC model to better resolve the open water areas. Diffuse groundwater discharge was also reduced to 124.5 mgd in the EFDC model from 154.2 mgd in the groundwater models. Reduction of discharge allowed a closer approximation to observed chloride concentration entering the river channel while maintaining chloride load.

Table 4-9. Constant, steady-state inputs of diffuse groundwater discharge, chloride load, and associated salinity within 16 subregions of the EFDC model.

EFDC Model Subregion	Area (km^2)	Diffuse Groundwater Discharge (mgd)	Chloride Load ($t\ d^{-1}$)	Salinity
1	13.6	11.2	58.6	2.50
2	10.9	4.6	21.8	2.25
3	4.1	10.2	42.7	2.00
4	9.3	5.5	14.4	1.25
5	12.7	5.2	16.5	1.50
6	6.5	5.3	19.5	1.75
7	3.7	5.7	21.5	1.80
8	1.2	3.5	11.1	1.49
9	35.0	14.2	53.7	1.80
10	1.8	13.7	50.3	1.75
11	3.9	3.4	11.1	1.55
12	2.2	7.9	37.3	2.25
13	10.4	9.5	45.0	2.25
14	2.3	9.6	24.8	1.23
15	8.9	5.2	19.7	1.80
16	1.3	9.8	12.0	0.58
Totals	127.8	124.5	460.0	——

4.4 SUMMARY

Two steady-state groundwater flow models are used to estimate discharge and chloride load to the middle St. Johns River, upper St. Johns River, and lower Ocklawaha River. Results show that diffuse groundwater is the dominant source of chloride to the middle St. Johns River. Diffuse groundwater delivers 75% of the chloride load to the reach of the middle St. Johns River upstream of U.S. 17 including lakes Monroe, Jesup, and Harney.

The groundwater models are used to estimate the effects of water withdrawals on groundwater discharge and chloride load to the rivers. Model results show that the effects of water withdrawals are small compared to the overall discharge and chloride budget of these river segments. The increase of diffuse groundwater discharge caused by a 155-mgd water withdrawal is only 0.1% of the average river flow, and the increase in chloride load is less than 1% of the existing mean. The insensitivity of groundwater discharge to water withdrawals is explained by the overwhelming magnitude of the hydraulic head of the Upper Floridan aquifer (3 to 6 m) relative to the expected change in river stage (~5 cm).

The importance of diffuse groundwater discharge to chloride load highlights the importance of quantifying diffuse groundwater sources for surface water modeling. The simulated discharge and chloride loads are contributed to the EFDC model within 16 river subregions. A constant, steady-state value of diffuse groundwater discharge and constant salinity (derived from chloride) are assigned to each subregion.

The use of a constant salinity in the surface water model is justified by the results of Section 3, which show that chloride and salt composition are stable in time. The assumption of constant discharge to represent diffuse groundwater discharge for surface water modeling has not been examined and is the topic of Section 5.

5 Surface Water Modeling: Comparison of Steady-State and Transient Groundwater Discharge

5.1 Introduction

The steady-state approximation for groundwater discharge to the EFDC model requires testing because groundwater discharge does respond to seasonal and interannual variations of rainfall and pumping. Groundwater modeling (see Section 4.3) contributed steady-state, constant groundwater discharge to the EFDC model, but EFDC is a dynamic model that simulates hydrodynamic variables at hourly time scales. This section examines whether forcing the EFDC model-using steady-state groundwater discharge (groundwater discharge that remains constant over time) differs appreciably from using transient groundwater (groundwater discharge that varies over time.) The question of whether steady-state groundwater discharge is adequate for use as a boundary condition to the EFDC model was of primary concern to a National Research Council (NRC) review of the initial groundwater modeling work.

Comparing the effects of steady-state and transient groundwater discharge requires an estimate of the transient groundwater discharge. The model simulation period for the EFDC model spans the period 1996 to 2005, so time series of transient groundwater discharge are developed for that specific period. The estimates of transient groundwater discharge are made using observed time series of river stage and Upper Floridan aquifer hydraulic head.

The simulation period includes a range of river flow conditions. Mean annual discharge at DeLand for these individual 11 years ranged from a low of 639 mgd during the year 2000 to 3,038 mgd during the year 1995. The mean annual discharge for the period 1957 to 2009 was 1,938 mgd.

Two tests were used to compare the use of transient and steady-state groundwater discharge in the EFDC model. Test 1 was a straightforward comparison of EFDC model output generated using steady-state groundwater discharge, and then transient groundwater discharge. Test 2 is slightly more complex; it tested whether the choice of steady-state or transient groundwater alters the outcome of a water withdrawal scenario.

Both tests examine the primary output from the EFDC model. These are time series of water level, discharge, salinity, and water age at representative locations throughout the middle St. Johns River. Results show that EFDC model simulations made using steady-state groundwater discharge are nearly indistinguishable from simulations made using transient groundwater discharge.

5.2 METHODS

Test 1. Sensitivity of the EFDC Model to Transient Groundwater

Test 1 examines the sensitivity of the EFDC model to transient groundwater discharge compared with constant, steady-state groundwater discharge. Simulated output are generated at eight locations (EFDC model sensitivity test points, see Figure 4-2) using both steady-state and transient groundwater discharge. Simulated output included hourly time series of water level (cm), surface- and bottom-layer salinity, surface- and bottom-layer water age (d), and discharge (mgd). Each time series contained 96,360 hourly values covering the 11-year model simulation period of 1995 through 2005. Steady-state scenario refers to the EFDC model run using steady-state groundwater discharge, and transient scenario refers to the EFDC model run using transient groundwater discharge.

Sensitivity is assessed by comparing the time series simulated by the steady-state scenario with time series simulated by the transient scenario. Time series at the eight output locations are compared by the following methods:

- Mean and standard deviation of hourly differences
- Coefficient of correlation (r^2) between hourly differences
- Scatterplots of paired hourly values
- Median relative error (%) of hourly differences
- Histograms of hourly differences
- Percentiles of distributions of hourly differences

Test 2. Sensitivity of Water Withdrawal Response to Transient Groundwater

Test 2 examines how steady-state and transient groundwater affects the EFDC model's predicted response to a water withdrawal. Test 2 uses the same steady-state and transient groundwater scenarios used for Test 1 yet also includes a 155-mgd water withdrawal with each scenario. This produces a predicted response of the river to a water withdrawal using both steady-state and transient groundwater discharge.

The predicted response to a water withdrawal is calculated by first establishing a base run. The base run for steady-state groundwater discharge is the steady-state scenario from Test 1. Similarly, the base run for transient groundwater discharge is the transient scenario. Next, each base run is altered to simulate a 155-mgd water withdrawal. The predicted responses are the differences between time series produced by a base run and its matching withdrawal scenario. The final result is two sets of time series of predicted responses: one set is derived from steady-state groundwater discharge, and a second set is derived from transient groundwater discharge. These two sets of time series are compared using the following methods:

- Comparison of the average predicted response
- Histograms showing hourly differences of the predicted response
- Calculation of the 95th percentile of distributions for the differences of the predicted response.

Estimating Transient Groundwater Discharge

Transient groundwater discharge is estimated from time series of hydraulic head differences between the Upper Floridan aquifer and the river for each of the five groundwater model segments (B to F, see Figure 4-2). Observed Upper Floridan aquifer hydraulic head is obtained from five wells (S-0025, V-0165, V-0818, S-1397, and L-0455) and matched with river stage at five surface gauges (SR46H, Osceola, SR 415, U.S. 17, and SR 40; see Figure 3-1). Time series of groundwater discharge are then calculated for each of the five segments by using Equation 5.1. Next, the calculated discharge time series are assigned to each of the 16 EFDC model subregions. Finally, the calculated discharge time series are adjusted so that the mean of each discharge time series is equal to the original steady-state discharge value simulated by the groundwater models (see Table 4-9).

Equation 5-1, from the USGS MODFLOW package, is used to calculate groundwater discharge as a linear function of the hydraulic head difference between the Upper Floridan aquifer and the river (McDonald and Harbaugh 1988).

$$\boldsymbol{Q_i}(\mathbf{t}) = -\boldsymbol{C_i} \cdot \left(\boldsymbol{h_{river,i}}(\mathbf{t}) - \boldsymbol{h_{aquifer,i}}(\mathbf{t})\right) \qquad \textbf{(5.1)}$$

where

$Q_i(t)$	=	Flow to or from the groundwater system in well i at time t (positive Q is to the river)
$h_{river,\ i}(t)$	=	River stage near well i at time t
$h_{aquifer,\ i}(t)$	=	Upper Floridan aquifer hydraulic head in well i at time t
C_i	=	The conductance of the river–aquifer interconnection in the groundwater model cell containing well i

River stage is observed at daily intervals (daily averaged values), and Upper Floridan aquifer hydraulic head is observed at monthly intervals (instantaneous values). Upper Floridan aquifer hydraulic head is disaggregated from monthly to daily intervals by linear interpolation to pair Upper Floridan aquifer hydraulic head with river stage in Equation 5.1. Conductance (C_i) of the river–aquifer interface is obtained from the groundwater model cell containing a given well. Paired river stage and Upper Floridan aquifer hydraulic head is used to estimate transient groundwater discharge for the period of 1996 through 2005.

Observed Upper Floridan Aquifer Hydraulic Head and River Stage

Upper Floridan aquifer hydraulic head is generally higher than river stage (Figure 5-1 to Figure 5-5). Mean Upper Floridan aquifer hydraulic head ranges from 2.56 to 6.81 m, while mean river stage ranges from 0.36 to 1.09 m, NGVD29. Upper Floridan aquifer hydraulic head is distinctly lower in well S-0025 compared to the other four wells, likely due to the absence of a confining layer in this area (Phelps and Rohrer 1987).

The variability of Upper Floridan aquifer hydraulic head tends to be lower than for river stage, although it exhibits a distinct seasonality in response to rainfall. Peak Upper Floridan aquifer hydraulic head generally occurs in October through November following the wet season; and low Upper Floridan aquifer hydraulic head generally occurs in May through June following the dry season. The coefficient of variation (*CV*), defined as the ratio of the standard deviation to the mean, is a useful measure of variability (Moore and McCabe 1989). *CV* ranges from 9 to 19 % for Upper Floridan aquifer hydraulic head (with the highest variability for S-0025), and from 60 to 73% for river stage. Despite the greater variability of river stage compared with hydraulic head, the variability of the difference between the Upper Floridan aquifer hydraulic head and stage is generally dominated by hydraulic head, because hydraulic head is greater than river stage in absolute magnitude. The dominance of Upper Floridan aquifer hydraulic head on head difference is particularly noticeable for the three downstream wells, V-0818, S-1397, and L-0455.

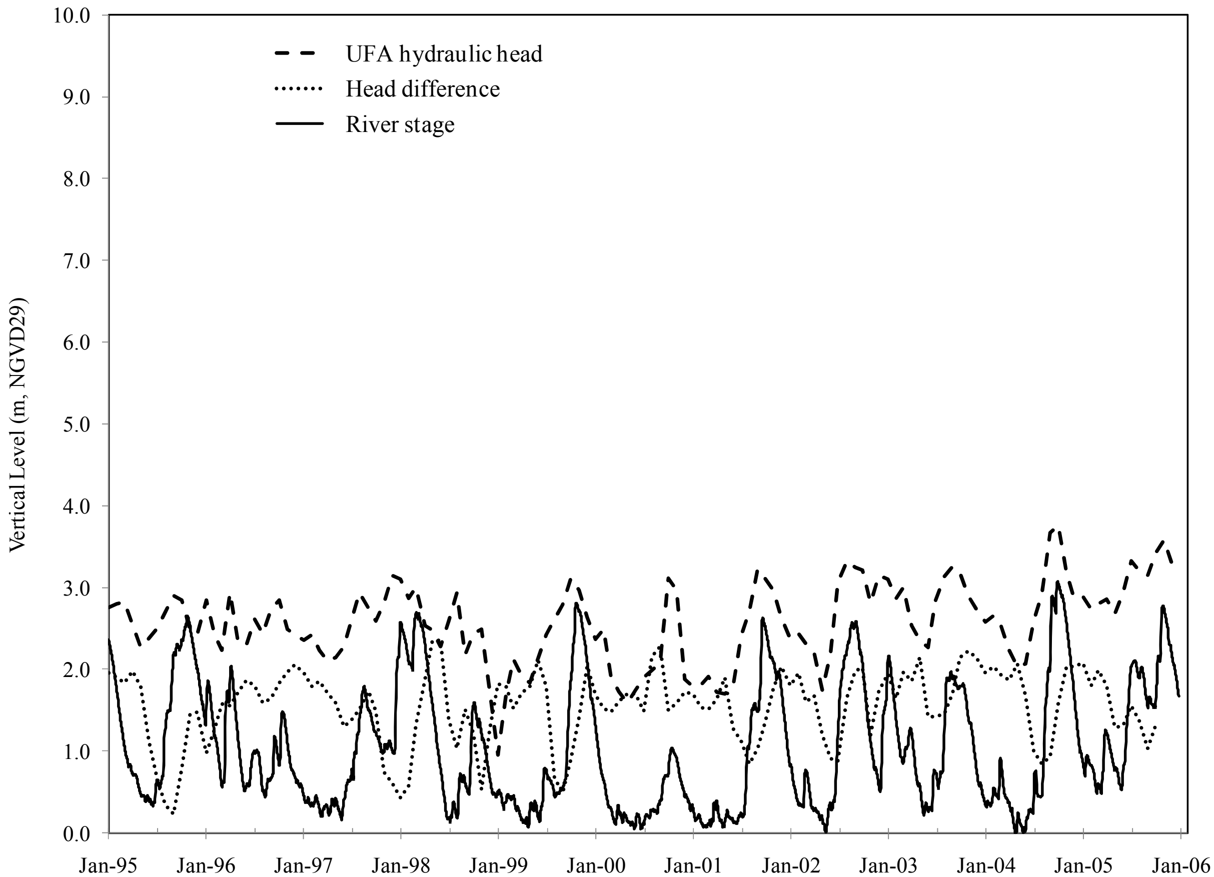


Figure 5-1. Upper Floridan aquifer hydraulic head in well S-0025 with adjacent river stage and head difference between the Upper Floridan aquifer and river.

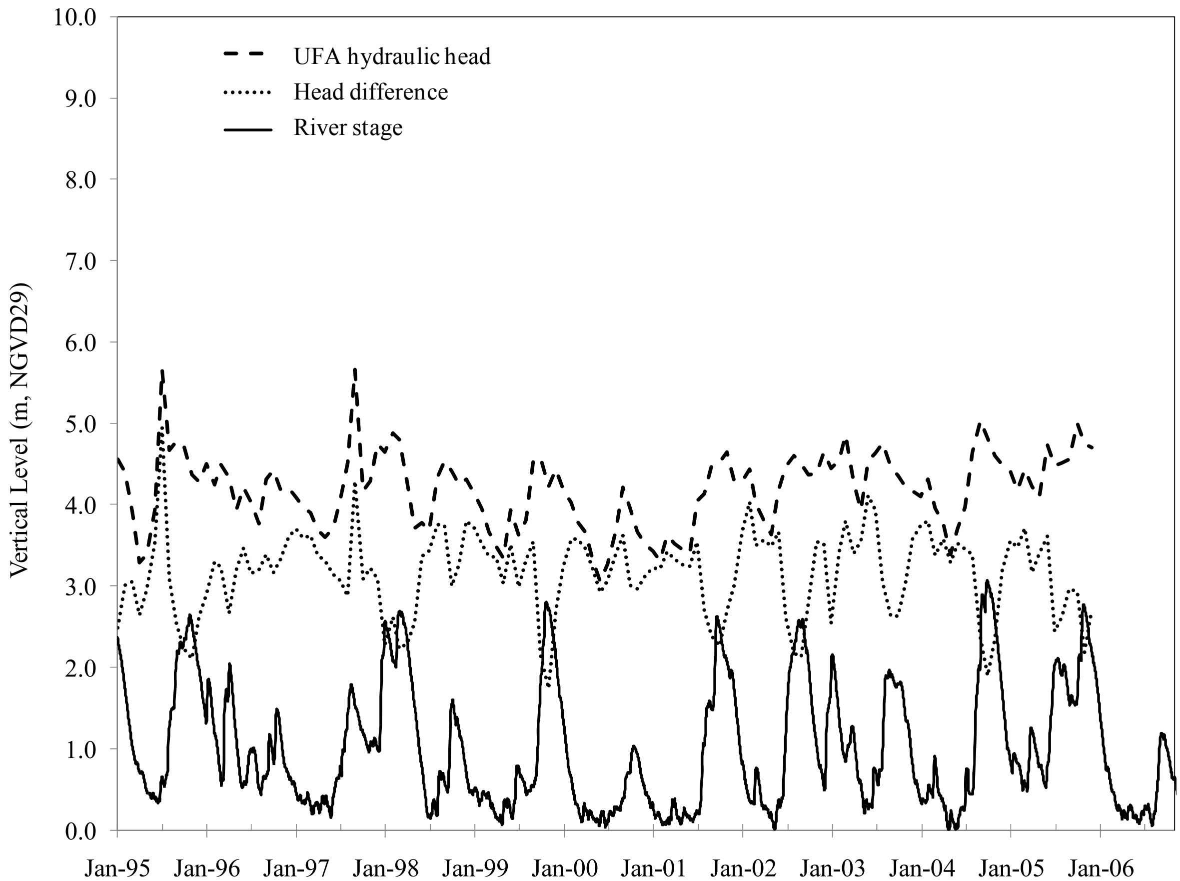
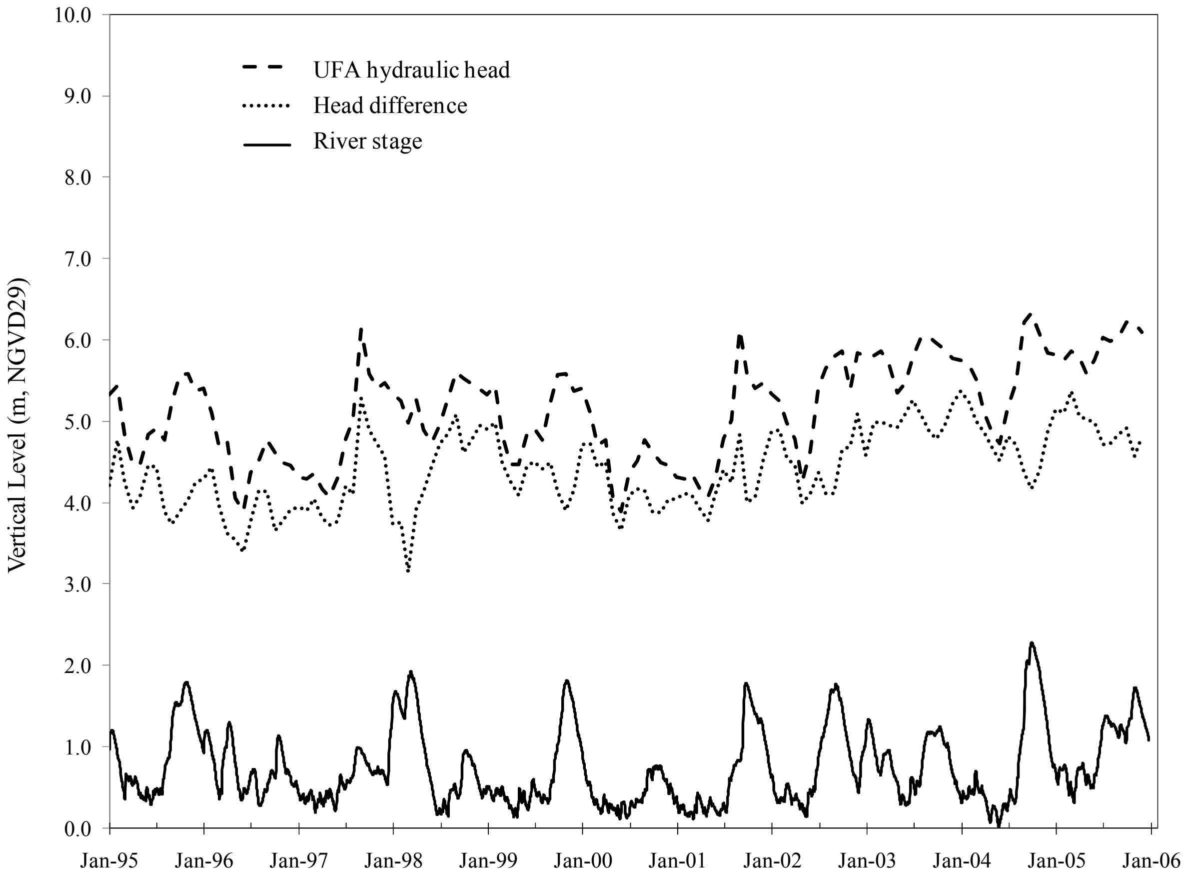


Figure 5-2. Upper Floridan aquifer hydraulic head in well V-0165 with adjacent river stage and head difference between the Upper Floridan aquifer and river.



10.0
9.0
8.0
7.0
6.0
5.0
4.0
3.0
2.0
1.0
0.0
Vertical Level (m, NGVD29)
UFA hydraulic head
Head difference
River stage
Jan-95
Jan-96
Jan-97
Jan-98
Jan-99
Jan-00
Jan-01
Jan-02
Jan-03
Jan-04
Jan-05
Jan-06

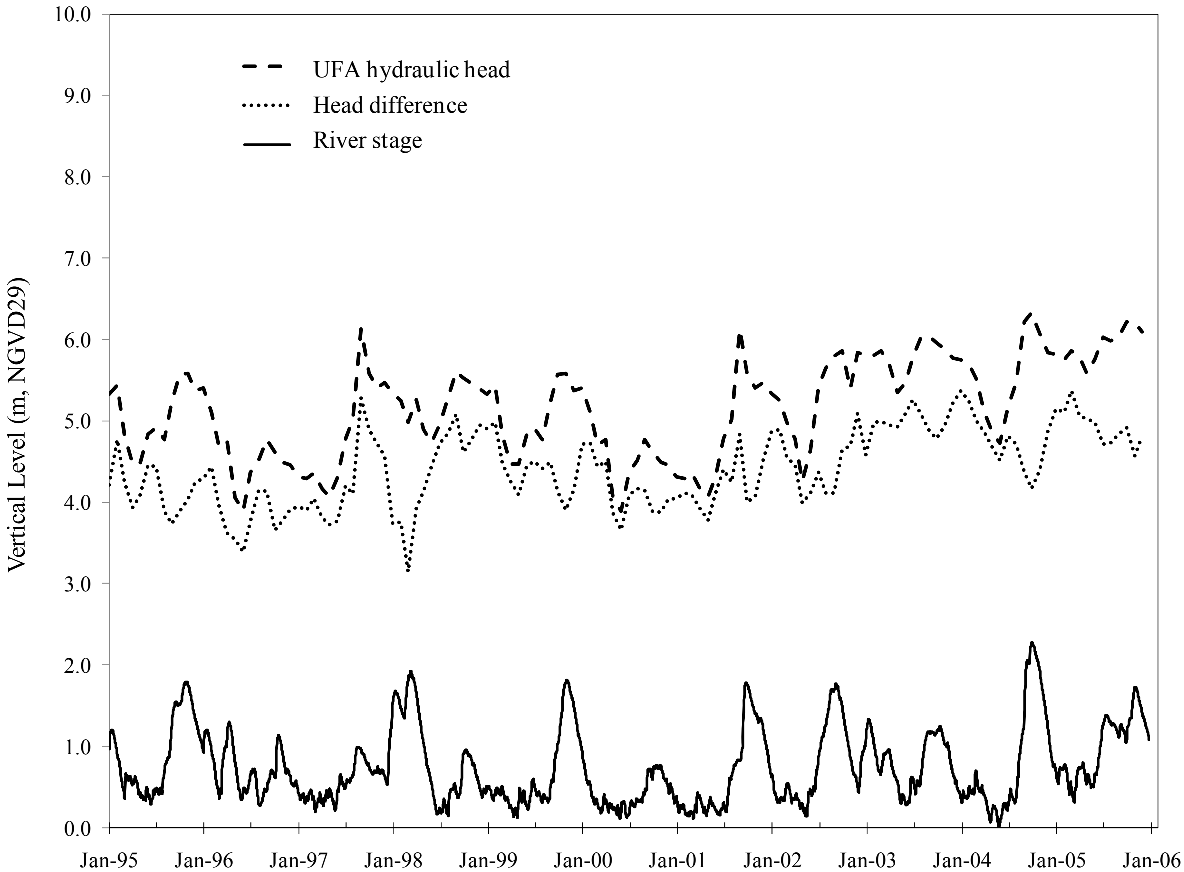


Figure 5-3. Upper Floridan aquifer hydraulic head in well V-0818 with adjacent river stage and head difference between the Upper Floridan aquifer and river.

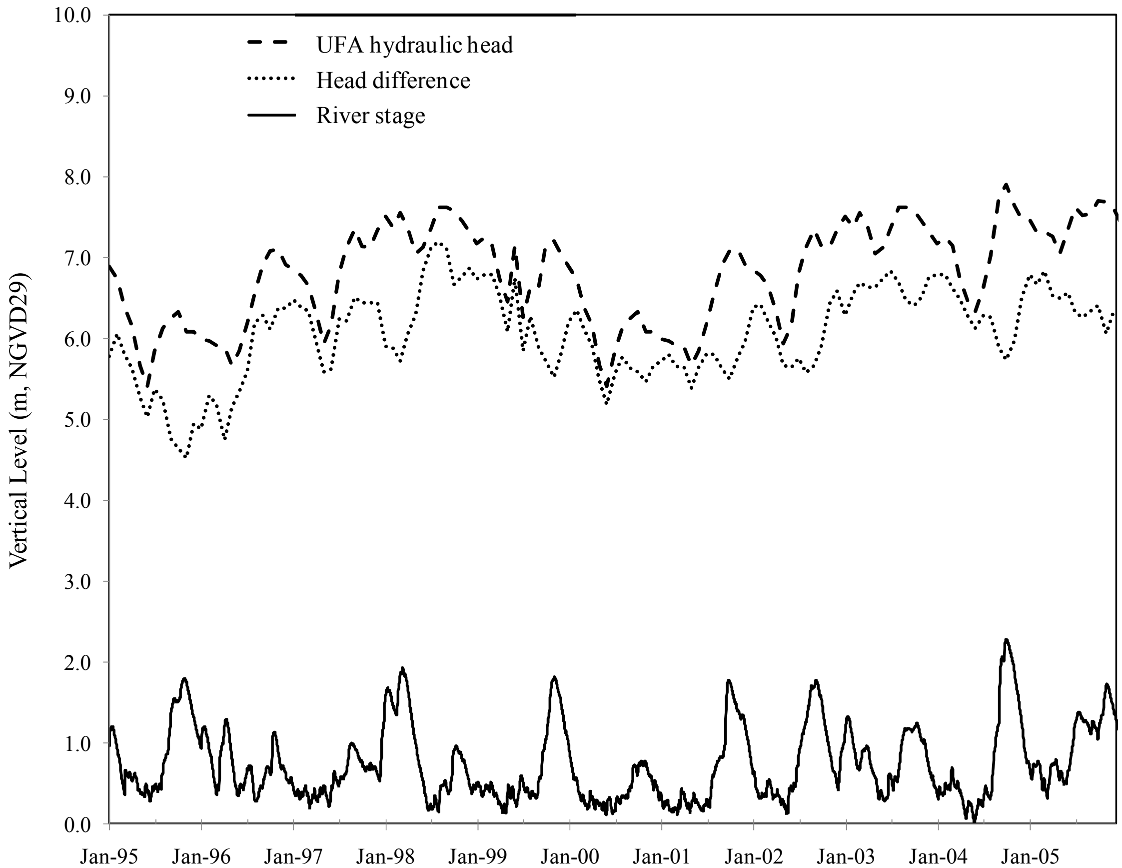


Figure 5-4. Upper Floridan aquifer hydraulic head in well S-1397 with adjacent river stage and head difference between the Upper Floridan aquifer and river.

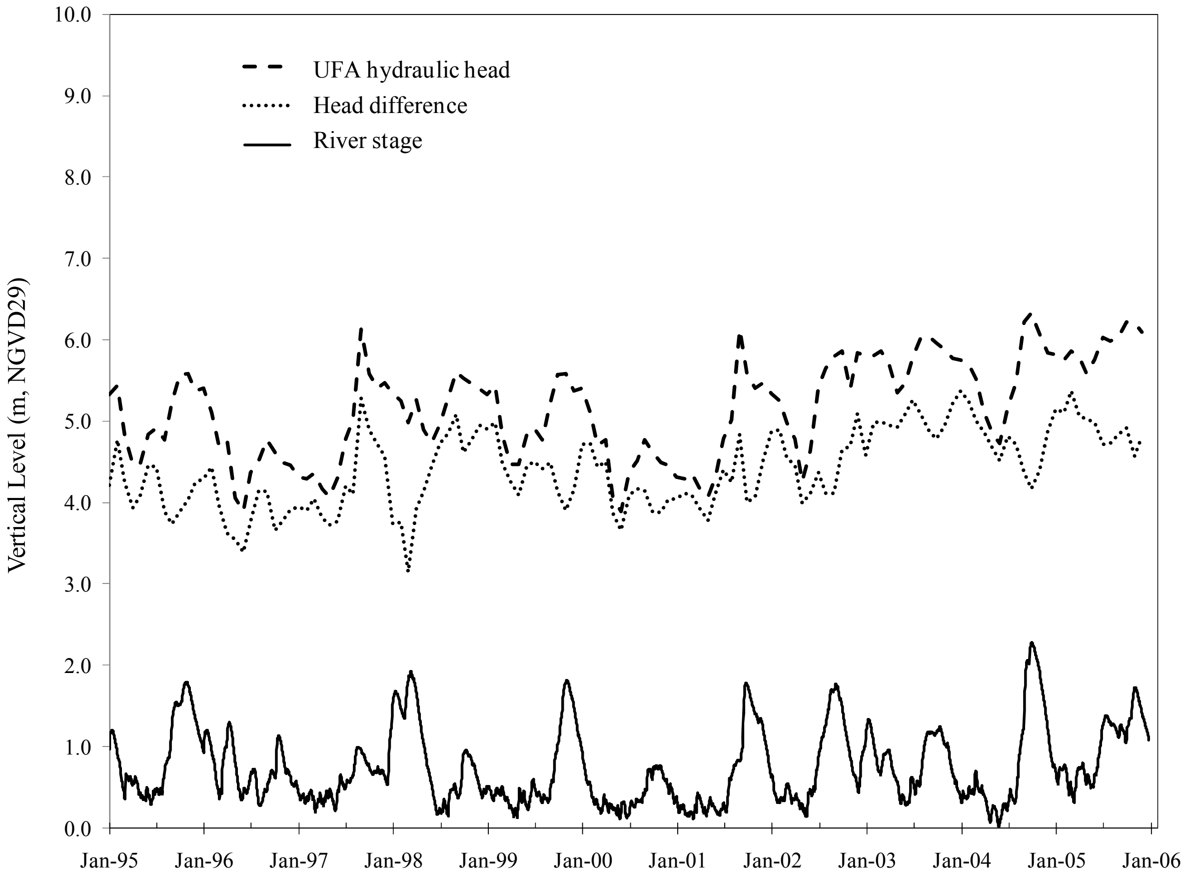


Figure 5-5. Upper Floridan aquifer hydraulic head in well L-0455 with adjacent river stage and head difference between the Upper Floridan aquifer and river.

Calculated Transient Groundwater Discharge

Time series of groundwater discharge is calculated, using Equation 5.1, from the time series of head differences shown in the above plots (Figure 5-6). Well S-0025 exhibits the greatest variability of discharge ($CV = 32\%$) and well S-1397 the least ($CV = 9.5\%$).

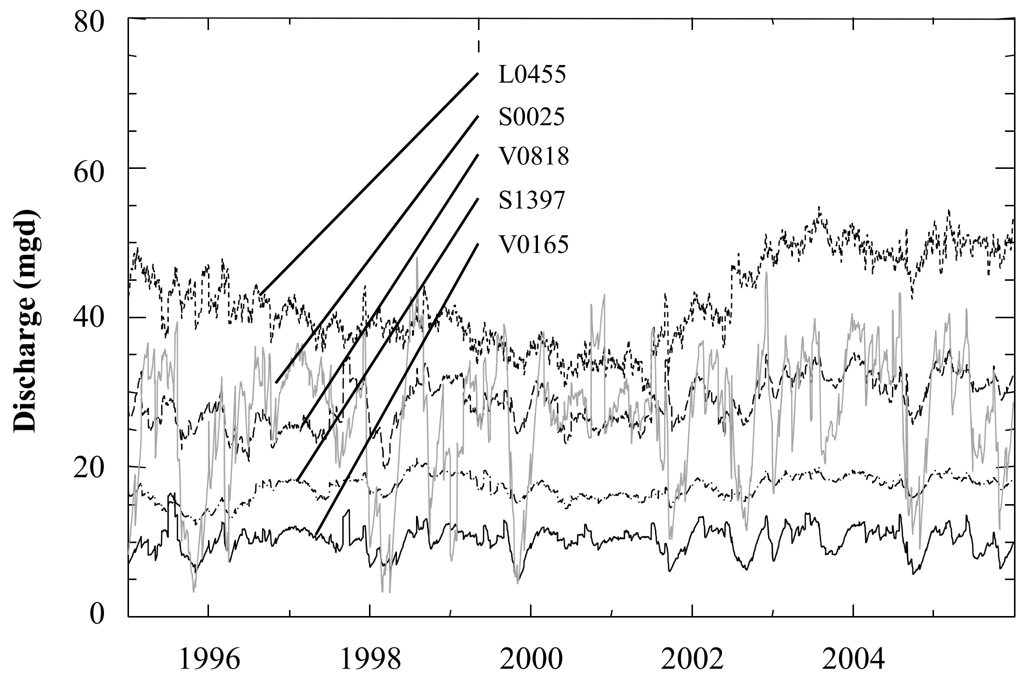


Figure 5-6. Calculated transient groundwater discharge at five well locations near the middle St. Johns River.

Converting Transient Groundwater Discharge to EFDC Model Boundary Conditions

The discharge time series assigned to each EFDC model subregion are adjusted so that their mean value equals the constant, steady-state groundwater discharge assigned to each subregion (see Table 4-9). This adjustment is made as follows:

For any given EFDC model subregion, let Q_{TR} equal the mean of the transient groundwater discharge time series assigned to the subregion (Table 5-1) and Q_{SS} equal the simulated steady-state groundwater discharge (see Table 4-9). Then the transient groundwater discharge is adjusted by multiplication of all values by the ratio Q_{TR}:Q_{SS}. The adjustment ensures that the long-term mean discharge and chloride load assigned to each EFDC model subregion are the same for both steady-state and transient groundwater discharge. The adjusted time series have the same coefficient of variance as the original time series. Differences between model runs comparing steady-state and transient groundwater boundary conditions are then due solely to groundwater transience.

Table 5-1. Assignment of time series of transient groundwater discharge associated with each of five wells to EFDC model subregions.

Well	L-0455	S-1397	V-0818	V-0165	S-0025
EFDC model subregion	1, 2, 3, 4, 5	6, 7	8, 9, 12, 13	14	10, 11, 15, 16

Transient discharge calculated at well S-0025 was assigned to four EFDC model subregions. The discharge variability at S-0025 is likely much greater than for other areas of the middle St. Johns River. Discharge at S-0025 has the greatest variability (gray line, see Figure 5-6) among wells because of its location near a zone of low Upper Floridan aquifer pressure head and within a river reach with large stage variations. Upper Floridan aquifer pressure head throughout the middle St. Johns River is higher than at S-0025 (Kinnaman 2005). Stage variations are lower downstream of Lake Harney. Lake Harney experiences the greatest fluctuation of stage during high flow events among all middle St. Johns River segments due to a constriction downstream of the lake's outlet. The constriction is formed by the river's shift to a geologically older valley as it flows around a block fault and passes between the DeLand Ridge to the north and Geneva Hill to the south (Pirkle 1971). Using S-0025 to represent transient groundwater discharge, then, overestimates discharge variability in areas downstream of Lake Harney. This overestimation of discharge variability is conservative for comparing steady-state and transient groundwater boundary conditions.

Final Transient Groundwater Boundary Conditions

The sum of all transient groundwater discharges assigned to the 16 EFDC model subregions is plotted in Figure 5-7 along with the associated transient chloride load. Transient chloride loads are calculated as the product of transient discharge and constant chloride concentration (equivalent to salinity) (see Table 4-9). The total transient groundwater discharge has a *CV* of 12.5%, a mean of 124.5 mgd, and a standard deviation of 15.6 mgd. The total transient chloride load has a *CV* of 13.5%, a mean of 460 t d^{-1} and a standard deviation of 62.5 t d^{-1}.

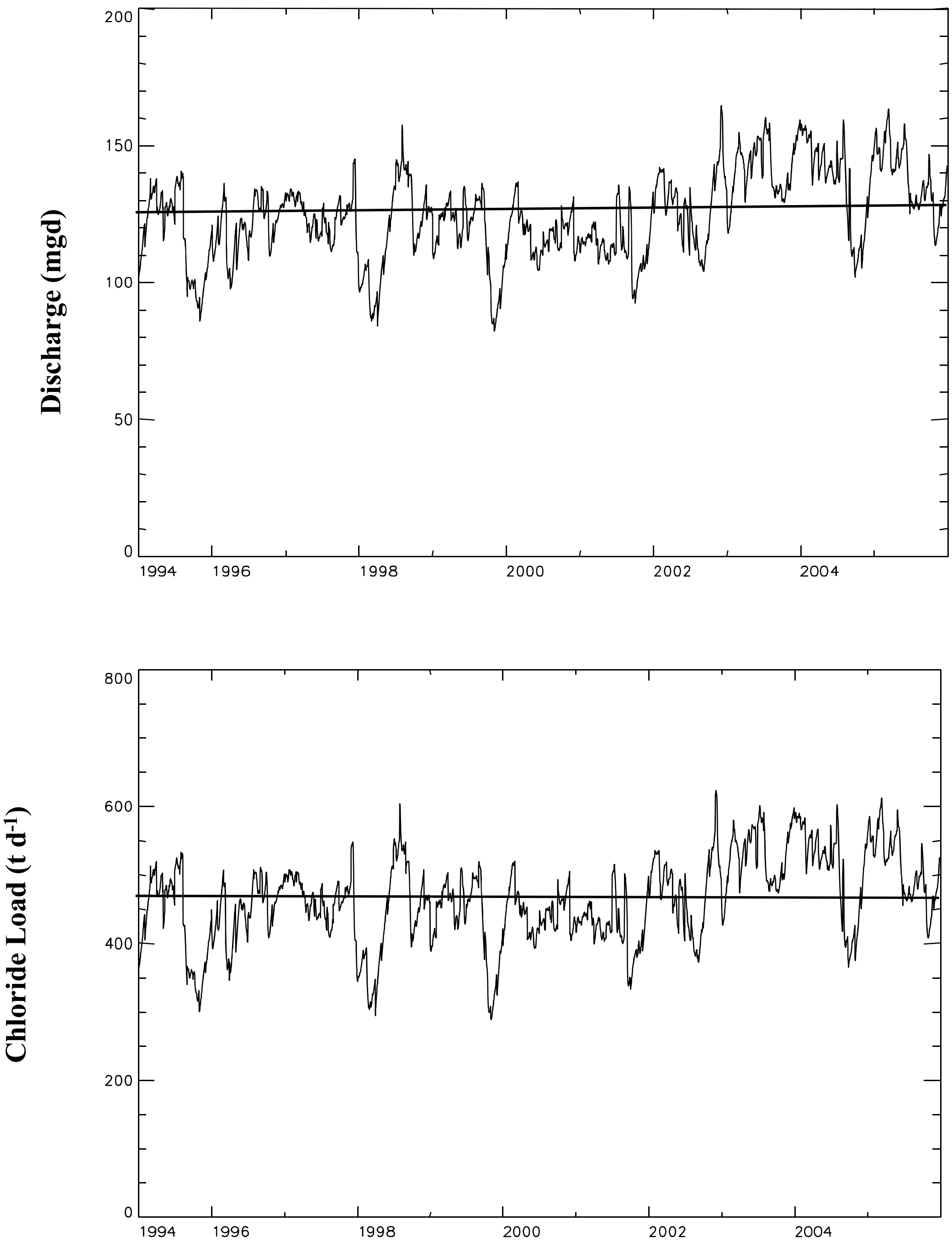


Figure 5-7. Total transient groundwater discharge (top) and chloride load (bottom) input to the EFDC model. Steady-state discharge (124.5 mgd) and load (460 t d^{-1}) are represented as time series by the horizontal lines. The steady-state values equal the mean of the transient time series.

5.3 RESULTS

5.3.1 TEST 1. SENSITIVITY OF THE EFDC MODEL TO TRANSIENT GROUNDWATER

Mean and Standard Deviation

Computed mean and standard deviation of the hourly differences between the steady-state and transient scenarios are shown in Table 5-2 for water level, discharge, salinity, and water age at each output location. The transient scenario was subtracted from the steady-state scenario, so a positive difference means the steady-state scenario is greater than the transient scenario. Discharge is not calculated for Lakes Jesup and Woodruff because there is no defined downstream direction of flow at these open-lake locations. Only bottom salinity and water age are shown because surface values are essentially identical. Mean differences are extremely small for all variables and stations.

Table 5-2. Mean and standard deviation of hourly differences between the steady-state and transient scenarios.

Location	Water Level (cm)		Discharge (mgd)		Bottom Salinity (PSS78)*		Bottom Water Age (d)	
	μ	σ	μ	σ	μ	σ	μ	σ
Osceola	0.034	0.309	-0.0012	5.026	-0.007	0.013	0.14	0.30
SR46J	0.044	0.251	0.0007	4.476	-0.005	0.025	0.18	1.83
Lake Jesup	0.043	0.253	n/a	n/a	-0.005	0.016	0.45	1.23
SR 415	0.042	0.223	-0.2225	10.626	-0.007	0.015	0.30	0.82
U.S. 17	0.043	0.220	0.0005	11.826	-0.005	0.013	0.12	0.81
SR 44	0.016	0.079	0.0019	12.871	-0.002	0.009	0.06	0.55
Lake. Woodruff	0.003	0.017	n/a	n/a	0.005	0.023	-0.25	0.93
SR 40	0.001	0.001	0.0084	15.556	-0.001	0.009	-0.01	0.49
Total	0.028	0.203	-0.0621	9.452	-0.003	0.017	0.13	1.00

* The Practical Salinity Scale 1978 (PSS78) (Lewis and Perkin 1978); (see Chapter 5)

Coefficient of Correlation

The coefficient of correlation (r^2) shows that the steady-state scenario explains more than 98% of the variability of the transient scenario for all output variables (Table 5-3).

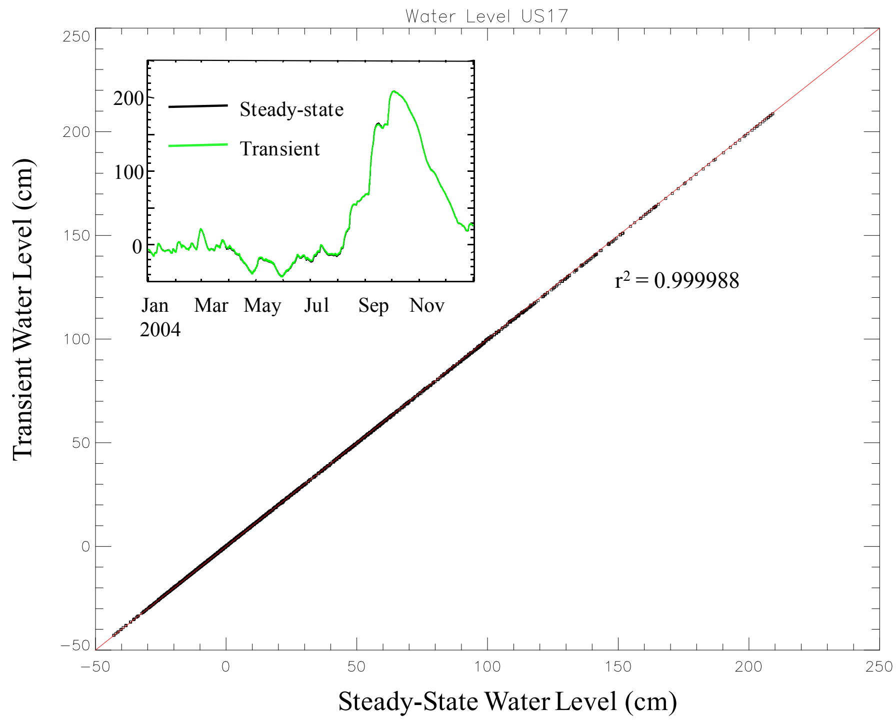
Table 5-3. Coefficient of correlation (r^2) between steady-state and transient scenarios.

Variable	Osceola	SR46J	Lake Jesup	SR 415	U.S. 17	SR 44	Lake Woodruff	SR 40
Water level	1.0000	1.0000	1.0000	1.0000	1.0000	1.0000	1.0000	1.0000
Discharge	1.0000	0.9997	n/a	1.0000	1.0000	1.0000	n/a	0.9999
Salinity	0.9988	0.9929	0.9953	0.9983	0.9979	0.9980	0.9884	0.9973
Water age	0.9999	0.9974	0.9991	0.9994	0.9993	0.9990	0.9905	0.9991

Scatterplots of Paired Values

Scatterplots of paired steady-state and transient values illustrate the high degree of correlation between the steady-state and transient scenarios. Scatterplots at all locations show a nearly straight-line relationship, so only one location, U.S. 17, is plotted here. This location was selected because it is the downstream boundary of the groundwater segments most affected by diffuse groundwater discharge. Figure 5-8 shows water level (top) and discharge (bottom), and Figure 5-9 shows salinity (top) and water age (bottom). (The paired data were resampled to include only every tenth point for practicality in creating the plot.) The red line is the 1:1 line of perfect fit. The inset plot is a sample time series for the year 2004. This year is selected to illustrate temporal variability because it has a particularly wide range of discharge between a dry spring and wet fall.

Simulated water level and discharge for the steady-state scenario are essentially identical to the transient scenario (Figure 5-8). Salinity and water-age pairs differ only slightly from the line of perfect fit (Figure 5-9). Salinity differences of about 0.05 occur when ambient salinity is above 0.8.



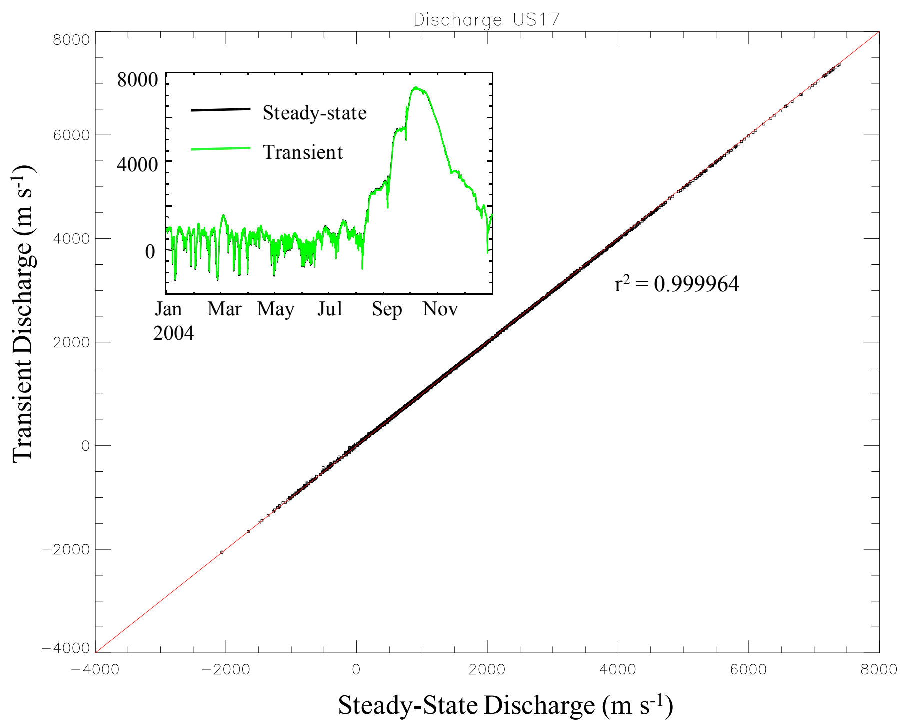
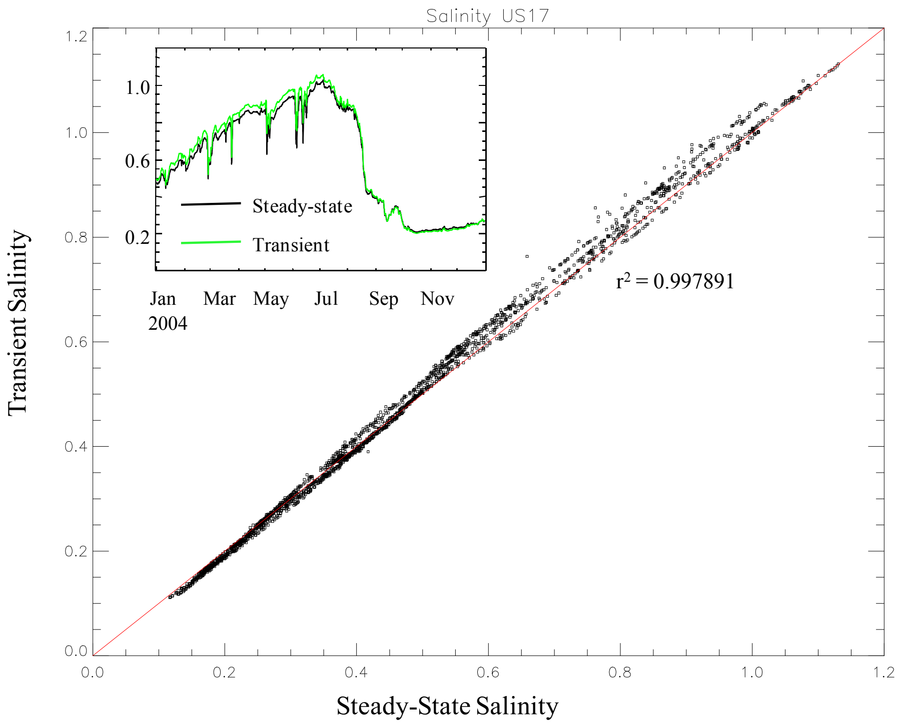


Figure 5-8. Scatterplot of paired hourly water level (top) and discharge (bottom) (1996 through 2005) at U.S. 17 for the steady-state and transient scenarios. The inset plot shows a sample time series for 2004.



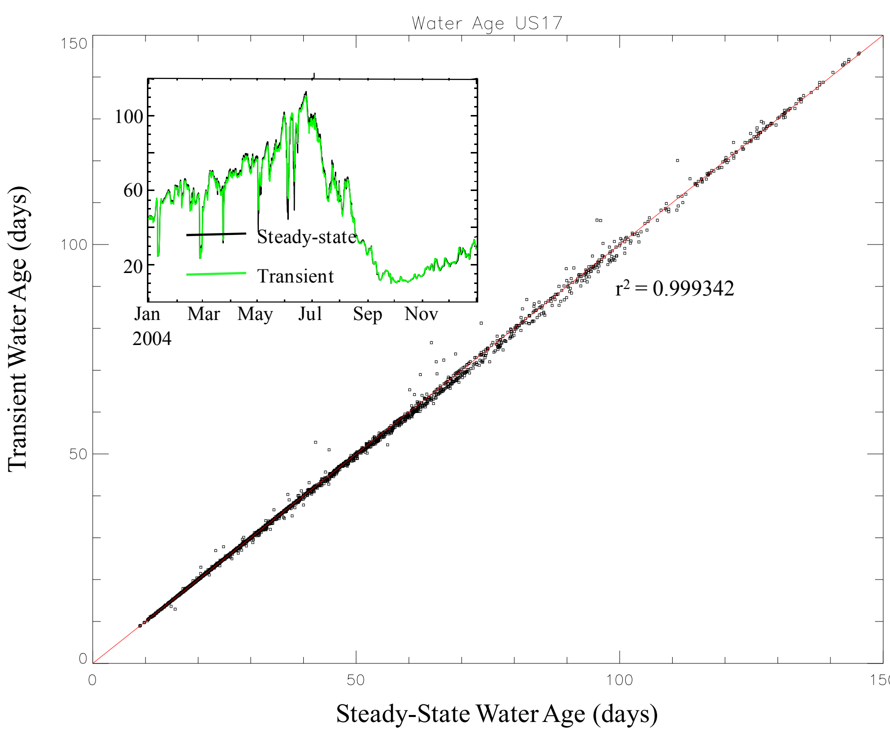


Figure 5-9. Scatterplot of paired hourly salinity (top) and water age (bottom) (1996 through 2005) at U.S. 17 for the steady-state and transient scenarios. The inset plot shows a sample time series for 2004.

Median Relative Error

Median relative error (*MRE*), the median value of the time series of relative errors, provides a good estimate of error between the steady-state and transient scenarios (Table 5-4). *MRE*, expressed as a percentage, is calculated as *MRE* = 100 x |ss-tr|/ss, where ss and tr are time series from the steady-state and transient scenarios, respectively. Although the reach of river between U.S. 17 and SR 40 had the largest absolute difference of discharge (Table 5-6), it had the smallest *MRE*, less than 0.5%. The large absolute errors are the result of small relative differences to large absolute values. The largest *MREs* (1.13% to 3.48%) occur for salinity, the variable with the smallest absolute differences. Even if the transient scenario was regarded as exact, an *MRE* of a few percent indicates that the steady-state scenario is a good representation of river salinity.

Table 5-4. Median relative error (MRE; %) for differences between the steady-state and transient scenarios.

Output Location	Water Level *MRE* (%)	Discharge *MRE* (%)	Salinity *MRE* (%)	Water Age *MRE* (%)
Osceola	0.41	0.33	1.78	0.45
SR46J	0.46	1.35	3.12	1.15
Lake Jesup	0.47	n/a	2.16	0.69
SR 415	0.44	0.52	2.23	0.58
U.S. 17	0.43	0.48	1.89	0.32
SR 44	0.22	0.41	1.28	0.33
Lake Woodruff	0.06	n/a	3.48	2.17
SR 40	<0.001	0.46	1.13	0.40

Histograms

The variability in the scatterplots shown previously can be visualized using histograms of paired differences. Histograms at U.S. 17 are plotted for water level (Figure 5-10), discharge (Figure 5-11), salinity (Figure 5-12), and water age (Figure 5-13). The histogram of water-level differences show a slight left-hand skew, meaning water level for the transient scenario is slightly greater than for the steady-state scenario. All histograms show small differences–98% of water-level differences are within ±0.75 cm, 94% of discharge differences are within ±20 mgd, 97.5% of salinity differences are within ±0.05, and 96.3% of water-age differences are within ±2.5 d.

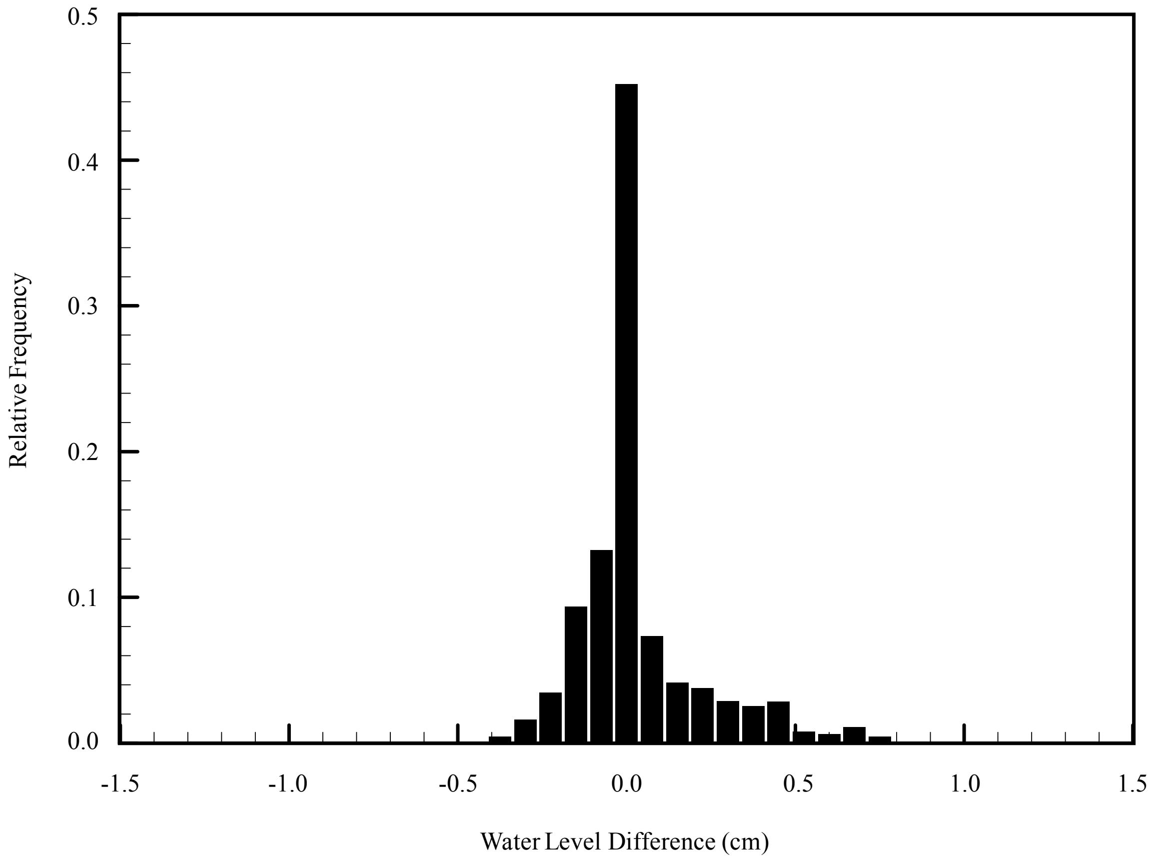


Figure 5-10. Histogram of hourly water level differences (cm) between steady-state and transient scenarios at U.S. 17.

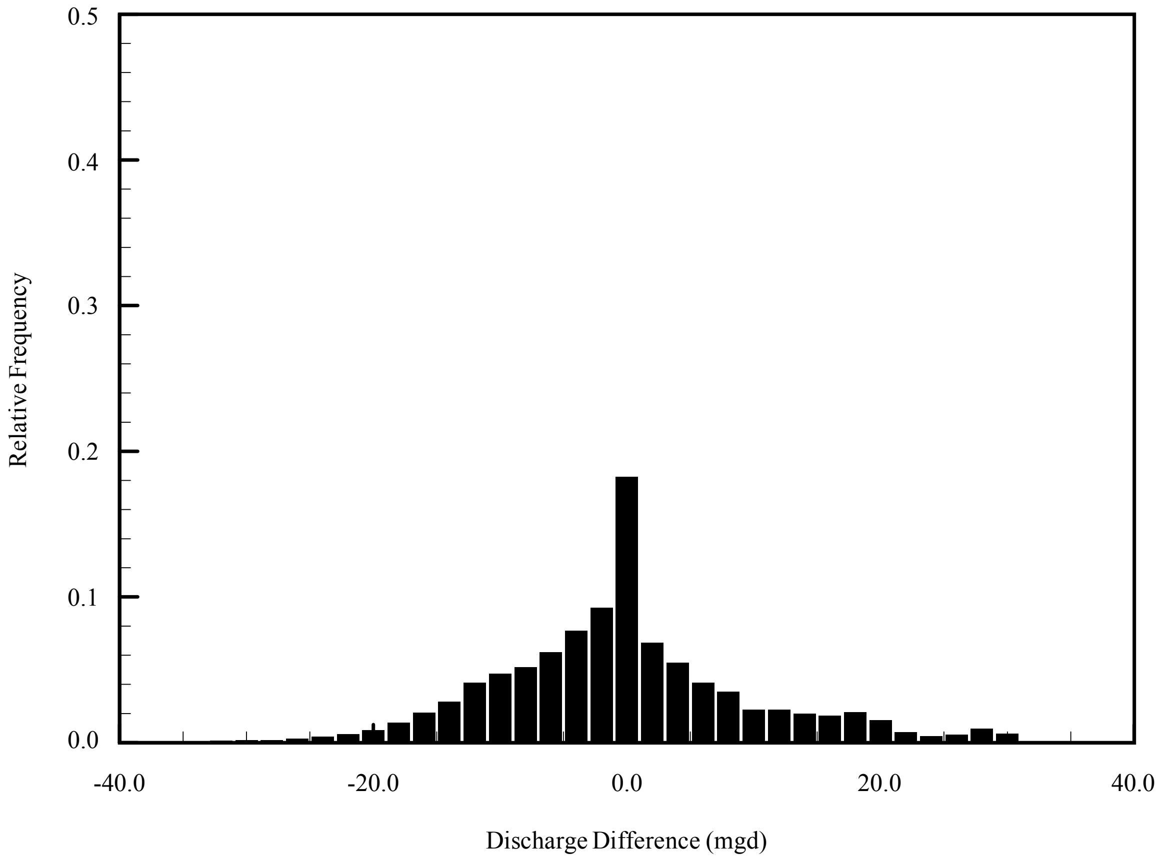


Figure 5-11. Histogram of hourly discharge differences (mgd) between steady-state and transient scenarios at U.S. 17.

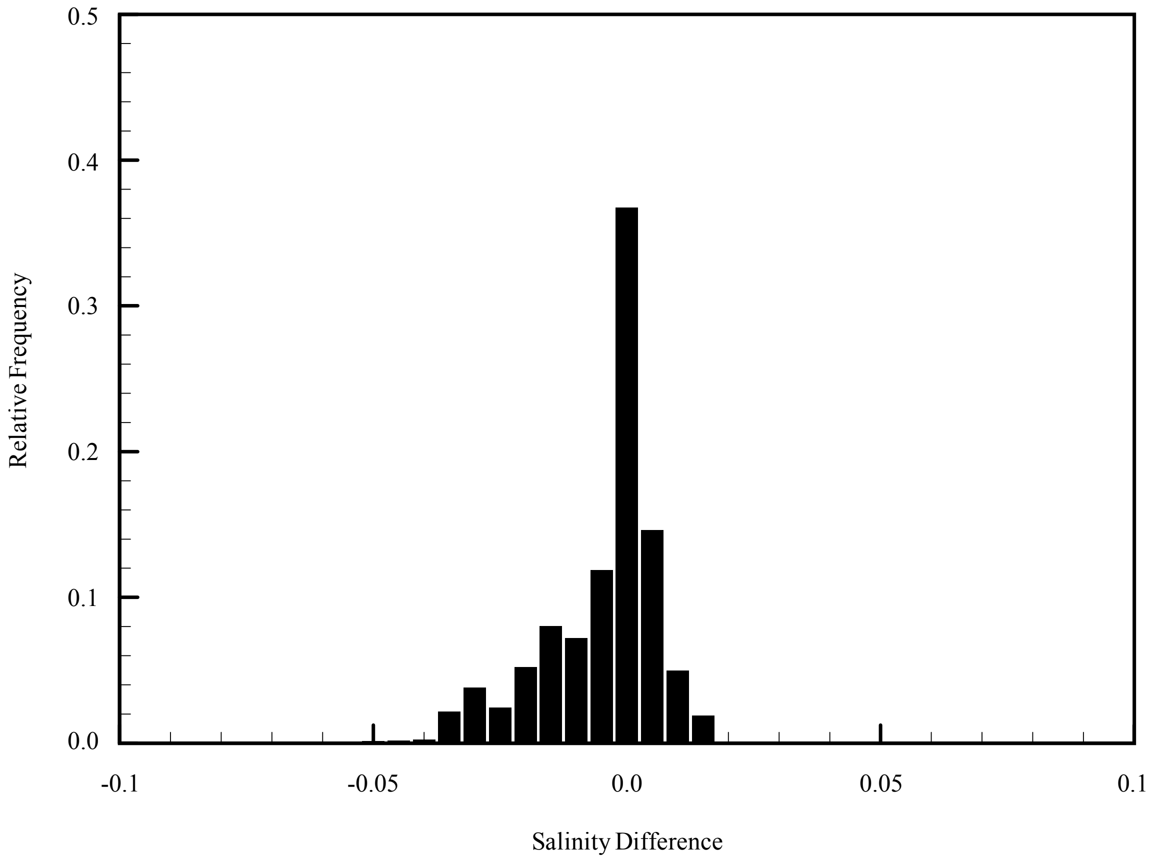


Figure 5-12. Histogram of hourly salinity differences between steady-state and transient scenarios at U.S. 17.

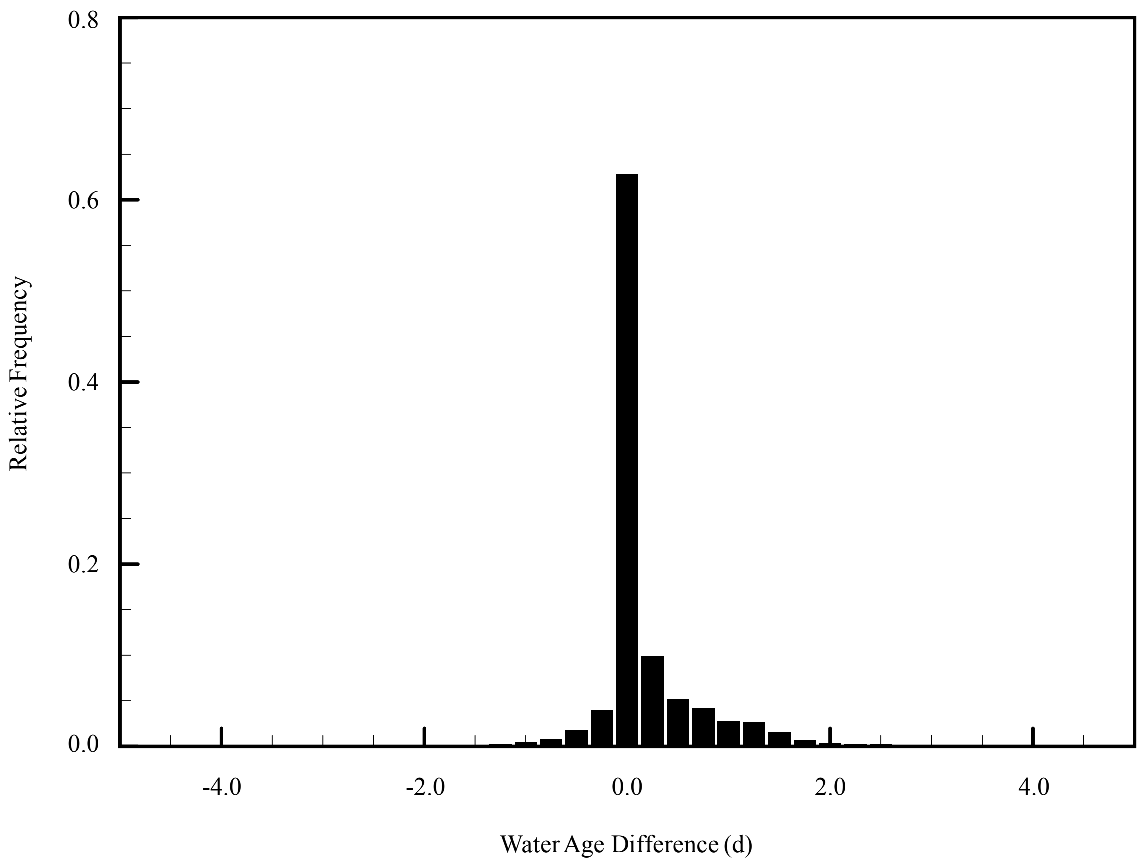


Figure 5-13. Histogram of hourly water-age differences (d) between steady-state and transient scenarios at U.S. 17.

Percentiles of Distributions

Extreme values for differences were examined to show the full numeric range of the calculated differences. Although histograms are useful for displaying the distribution of differences for the bulk of the paired data, they do not discriminate extreme values well. Extreme values are shown by identifying the maximum difference and the 50th, 75th, 95th, and 99th percentiles. The N-th percentile is the value that exceeds N% of all values.

Water level upstream of U.S. 17 is more sensitive to transient groundwater than at downstream locations (Table 5-5). The downstream differences are artificially small because of proximity to the specified water-level boundary condition at SR 40. The absolute water-level differences, even above U.S.17, are small; 95% of all differences are less than 7 mm, and the absolute maximum water-level difference does not exceed 1 cm.

Table 5-5. Percentile and maximum values for absolute differences of hourly water level (cm) between the steady-state and transient scenarios.

Output Location	Percentile				
	50%	75%	95%	99%	Maximum
Osceola	0.170	0.347	0.654	0.867	0.910
SR46J	0.113	0.251	0.560	0.798	0.837
Lake Jesup	0.114	0.254	0.563	0.802	0.838
SR 415	0.096	0.217	0.504	0.734	0.774
U.S. 17	0.093	0.211	0.499	0.730	0.770
SR 44	0.031	0.074	0.182	0.278	0.314
Lake Woodruff	0.010	0.018	0.036	0.052	0.076
SR 40	0.001	0.001	0.003	0.004	0.006
All Records	0.049	0.168	0.489	0.754	0.910

Maximum hourly discharge differences are at times greater than 100-mgd downstream of U.S. 17 (Table 5-6). These differences appear substantial relative to the magnitude of the proposed withdrawal of 155 mgd. Even at the 95th percentile, discharge differences were 9.1 to 31.3 mgd. These differences are explained by small phase changes of a much larger fluctuating ambient discharge. The maximum discharge difference of 113 mgd, for example, occurred at SR 40 on 17 March 2005. This difference occurred during a five-day period, beginning on 12 March, when discharge dropped from 2,500 to -2,000 mgd, a net change of 4,500 mgd. The discharge difference between the steady-state and transient scenarios is only 2.5% of the total change in ambient discharge at this time.

Table 5-6. Percentile and maximum values for absolute differences of hourly discharge (mgd) between the steady-state and transient scenarios.

Output Location	Percentile				
	50%	75%	95%	99%	Maximum
Osceola	3.3	5.7	10.1	12.5	29.4
SR46J	2.5	4.8	9.1	11.9	43.9
SR 415	6.8	12.2	21.0	26.9	68.0
U.S. 17	6.8	13.1	23.9	31.1	101.7
SR 44	7.3	14.3	26.3	33.0	100.2
SR 40	9.9	17.3	31.3	39.2	113.4
All Records	3.2	8.5	21.2	31.5	113.4

Maximum hourly salinity differences are greatest at the outlets of Lake Jesup (SR46J; SR46J is a location where State Road 46 crosses the St. Johns River near Lake Jesup), Lake Monroe (U.S. 17), and Lake Woodruff (Table 5-7). For all locations, 95% of salinity differences were less than 0.05 and 99% were less than 0.07.

Table 5-7. Percentile and maximum values for absolute differences of hourly salinity between the steady-state and transient scenarios.

Output Location	Percentile				
	50%	75%	95%	99%	Maximum
Osceola	0.007	0.014	0.033	0.047	0.058
SR46J	0.017	0.030	0.049	0.060	0.143
Lake Jesup	0.009	0.017	0.037	0.047	0.055
SR 415	0.009	0.017	0.035	0.045	0.083
U.S. 17	0.007	0.014	0.032	0.039	0.112
SR 44	0.005	0.010	0.022	0.028	0.044
Lake Woodruff	0.013	0.025	0.049	0.070	0.105
SR 40	0.005	0.010	0.022	0.027	0.037
All Records	0.008	0.017	0.037	0.052	0.143

Maximum differences for water age of 8 to 30 days occur between the mouth of Lake Jesup and SR 40 (Table 5-8). These extreme values are large but infrequent. 95% of the water-age differences are less than 4 days, and 99% are less than 6 days. The maximum difference (30.1 d) at SR46J occurred on 25 August 1999 during a 12-hour period of rapid flushing when water ages throughout the system were rapidly declining. By the end of the day, water-age differences returned to near zero.

Table 5-8. Percentile and maximum values for absolute differences of hourly water age (days) between the steady-state and transient scenarios.

Output Location	Percentile				
	50%	75%	95%	99%	Maximum
Osceola	0.060	0.230	0.730	1.280	4.840
SR46J	0.800	1.640	3.700	6.200	30.100
Lake Jesup	0.690	1.200	3.100	4.400	6.300
SR 415	0.130	0.510	1.900	3.480	15.740
U.S. 17	0.120	0.400	1.440	2.980	16.170
SR 44	0.090	0.310	1.100	2.340	8.290
Lake Woodruff	0.192	0.499	1.870	4.050	14.710
SR 40	0.100	0.310	1.071	1.850	12.380
All records	0.190	0.630	2.080	4.100	30.100

5.3.2 TEST 2. PREDICTED RESPONSE TO WATER WITHDRAWAL

This section tests differences between the predicted effects of a 155-mgd water withdrawal made using steady-state and transient groundwater discharge. Results from section 5.3.1 showed that the steady-state and transient scenarios are very similar. It is expected that both these models will predict similar responses to a water withdrawal. Therefore, the predicted response of a water withdrawal will not depend on the use of steady-state versus transient groundwater discharge.

Comparison of the Average Predicted Response

The predicted effects of a water withdrawal using steady-state groundwater discharge are calculated by using the steady-state scenario as a base condition. This scenario is modified to simulate the removal of 155 mgd of water from three locations. The difference between time series produced by the base condition and withdrawal scenario is the predicted response.

A predicted response is calculated separately here using both steady-state and transient groundwater discharge. The predicted response to a water withdrawal at U.S. 17 is shown in Table 5-9. The mean and standard deviations shown in the table are calculated from hourly differences over a 10-year period. The average drop in water level is 2.3 cm at this location, and this predicted change is identical using either steady-state or transient groundwater. The average change in salinity and discharge are also essentially identical for steady-state and transient groundwater. (The average drop in discharge of 102.9 mgd is less than 155 mgd, because one of the three withdrawal locations is downstream of U.S. 17.)

Only water age has a different average-predicted response to a water withdrawal when using steady-state versus transient groundwater discharge. The average increase of water age is incrementally larger. The difference of the predicted response for mean water age is 0.402 - 0.369 = 0.033 d. A detailed examination, using histograms, of the difference of the predicted response for all variables follows below.

Table 5-9. Mean (μ) and standard deviation (σ) of predicted response to a 155-mgd water withdrawal at U.S. 17 using both steady-state and transient groundwater discharge.

Predicted Variable	Steady-State Groundwater		Transient Groundwater	
	μ	σ	μ	σ
Water level (cm)	-2.3	1.7	-2.3	1.7
Discharge (mgd)	-102.9	67.5	-102.9	67.4
Salinity	-0.012	0.037	-0.012	0.037
Water age (d)	0.369	7.106	0.402	7.003

Histograms of Hourly Difference of the Predicted Response

The hourly difference of predicted response to a water withdrawal are shown as histogram plots for water level (Figure 5-14), discharge (Figure 5-15), salinity (Figure 5-16), and water age (Figure 5-17) at U.S. 17. The difference of predicted response are about an order of magnitude smaller than the differences for simulated variables shown for Test 1 (Section 5.3.1). For example, the hourly difference for simulated salinity (see Figure 5-12) is ±0.05, while the difference for the predicted response of salinity to a water withdrawal is ±0.005. Transient groundwater discharge has less effect on the predicted response to a water withdrawal than on the simulation of a given hydrodynamic variable.

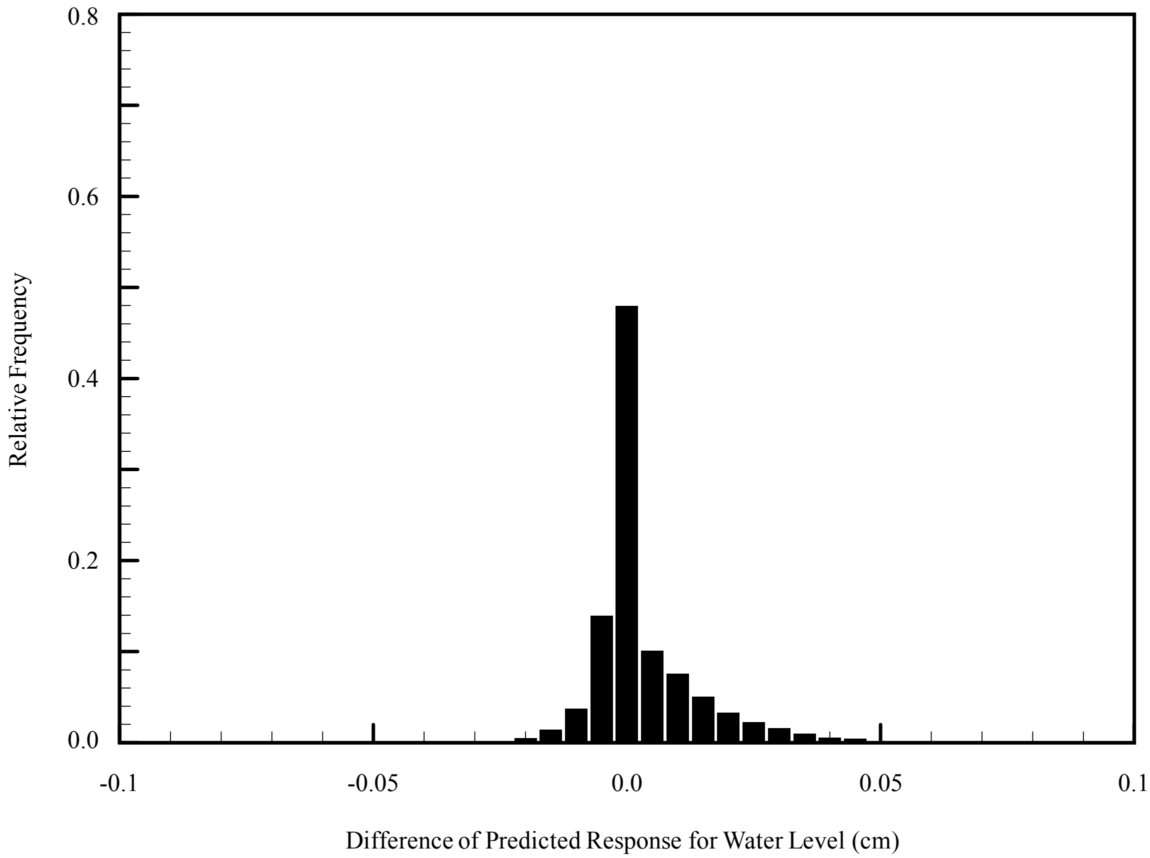


Figure 5-14. Histogram of difference of predicted response in water level at U.S. 17 using steady-state and transient groundwater discharge.

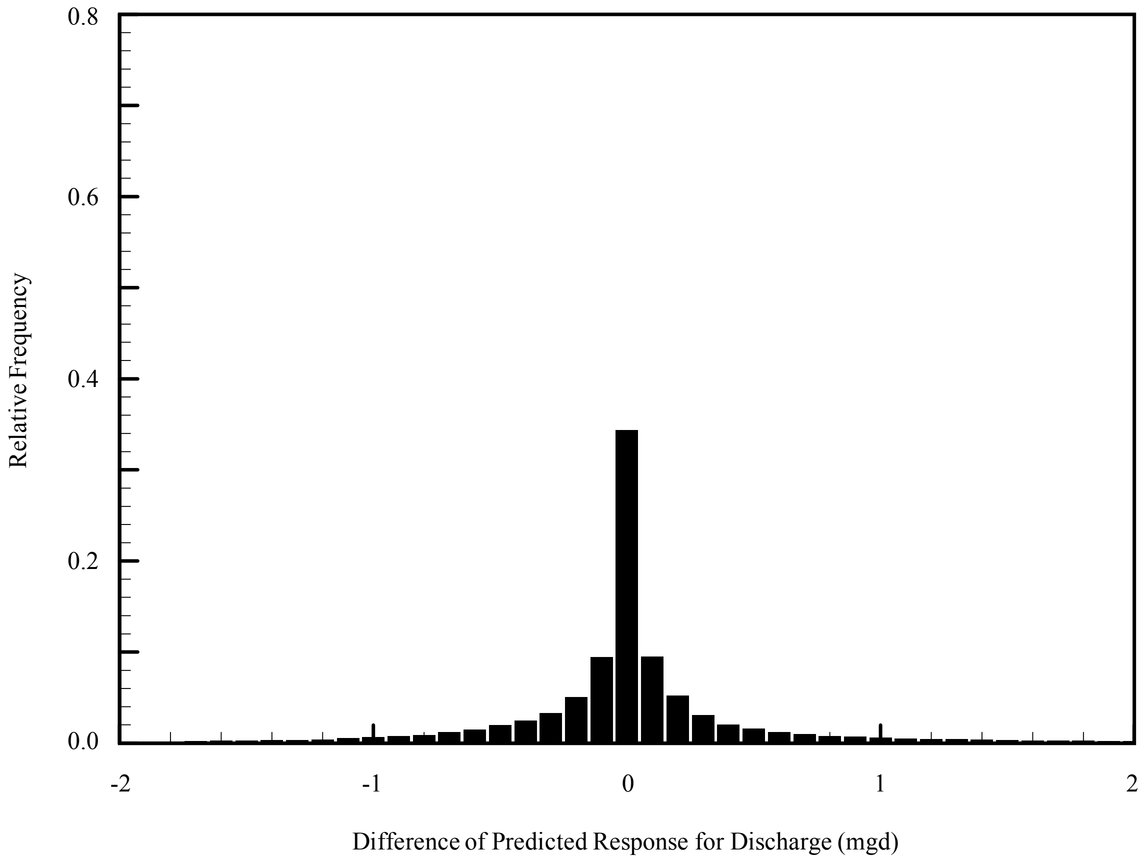


Figure 5-15. Histogram of difference of predicted response in discharge at U.S. 17 using steady-state and transient groundwater discharge.

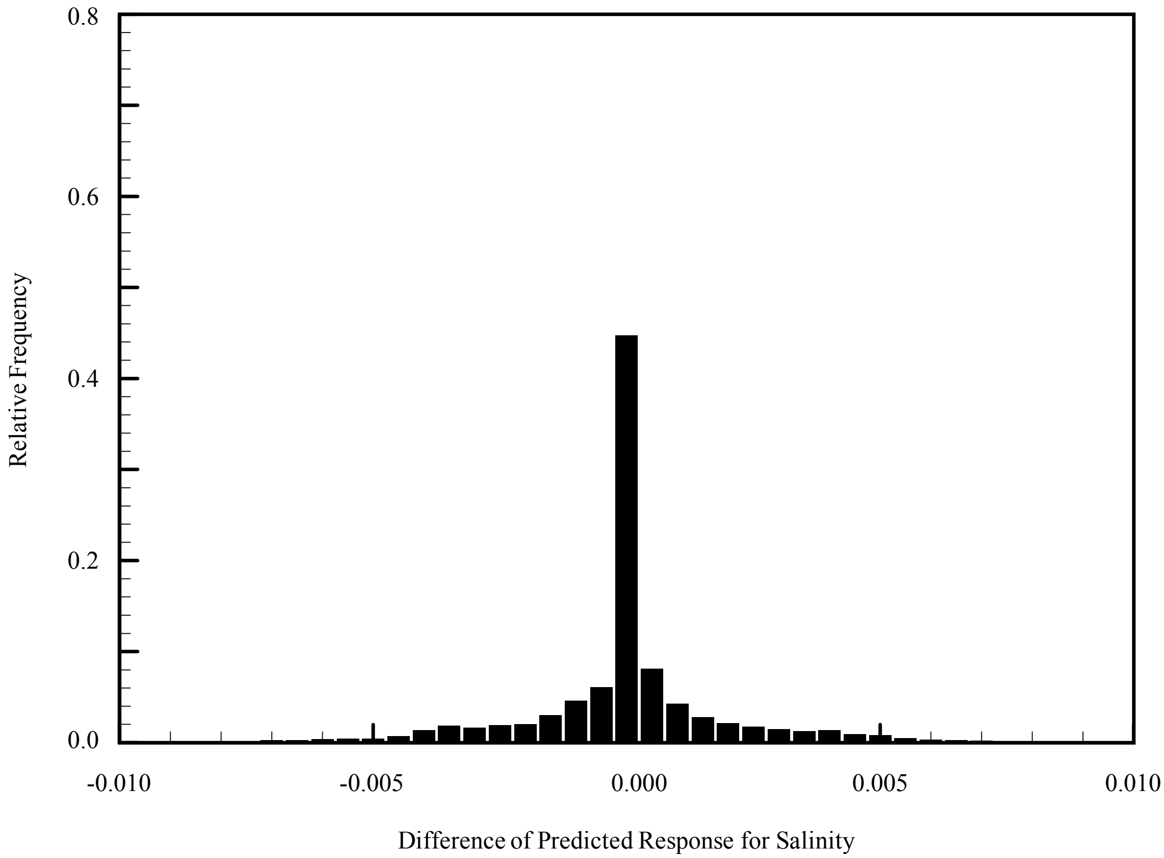


Figure 5-16. Histogram of difference of predicted response in salinity at U.S. 17 using steady-state and transient groundwater discharge.

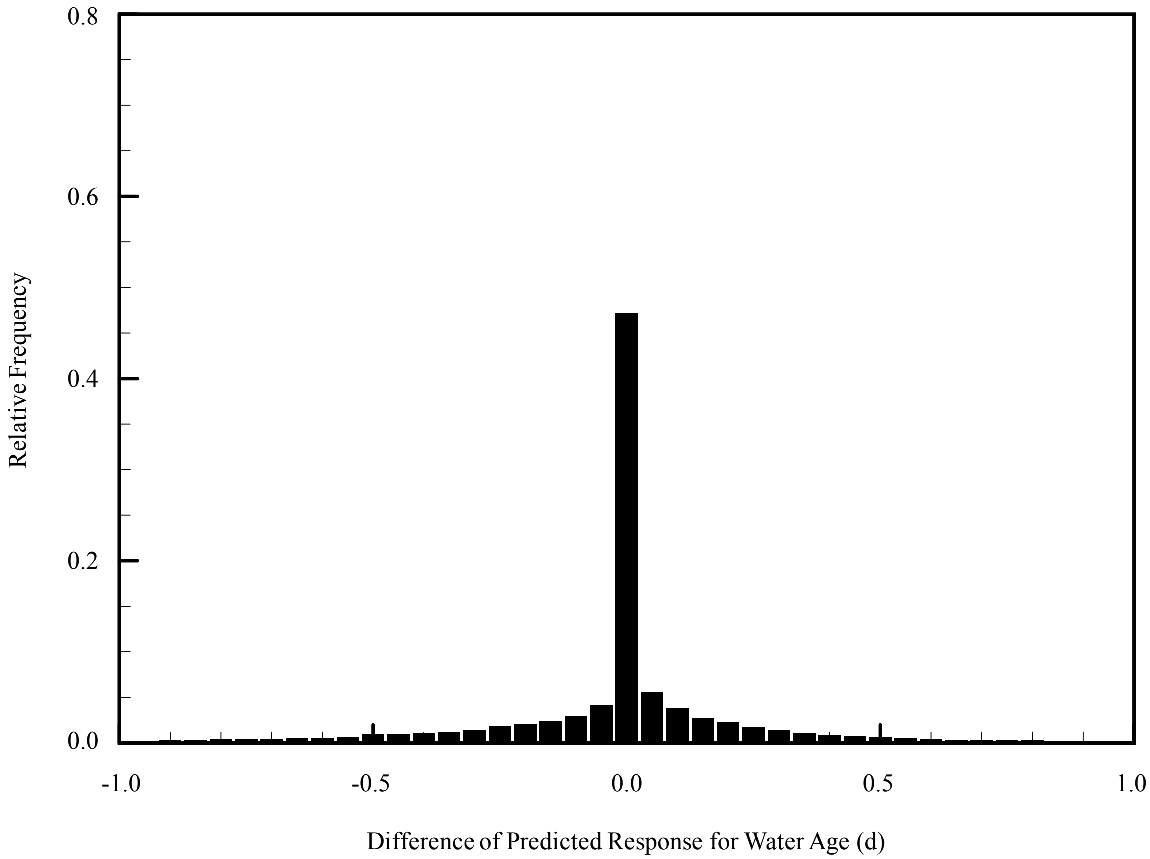


Figure 5-17. Histogram of difference of predicted response in water age at U.S. 17 using steady-state and transient groundwater discharge.

Ninety-Fifth Percentile for Difference of the Predicted Response

The above histograms do not clearly show extreme values. Extreme values for difference of predicted response to a water withdrawal are shown in Table 5-10 for the 95th percentile. Even for extreme values, transient groundwater discharge has a negligible effect on predicted response. For water level, 95% of the difference of predicted response differ by less than 0.3 mm. Extreme differences for discharge are less than 3 mgd, for salinity are less than 0.01, and for water age are less than 1.5 days.

Table 5-10. Extreme high (95th percentile) for absolute value of difference of predicted response to a water withdrawal using steady-state and transient groundwater.

Variable	Osceola	SR46J	Lake Jesup	SR 415	U.S. 17	SR 44	Lake Woodruff	SR 40
Water level (cm)	0.033	0.031	0.030	0.027	0.027	0.010	0.003	0.001
Discharge (mgd)	0.58	1.12	n/a	1.44	2.34	2.22	n/a	2.74
Salinity (PSS78)*	0.003	0.005	0.001	0.005	0.006	0.008	0.003	0.005
Water age (days)	0.164	1.619	0.199	1.190	1.199	1.469	0.171	0.954

* The Practical Salinity Scale 1978 (PSS78) (Lewis and Perkin 1978); (see Chapter 5).

5.4 SUMMARY

The use of steady-state groundwater discharge as a boundary condition to the EFDC model is well justified based on comparison of the EFDC model results generated using both steady-state and transient groundwater. Similarly, the predicted response of hydrodynamic variables to a 155-mgd water withdrawal is unaffected by the choice of either steady-state or transient groundwater discharge.

6 CONCLUSIONS

Groundwater is an important contribution to the base flow of the middle St. Johns River. Groundwater enters as diffuse upward flow through the river and from numerous springs. The influence of groundwater is greatest in the middle St. Johns River and lower portions of the upper St. Johns River because the underlying Hawthorn Formation is thin or absent in these area. In the lower St. Johns River and upper portions of the upper St. Johns River, the Hawthorn Formation comprises a thick confining layer that inhibits upward flow of groundwater to the river. Diffuse groundwater flow is the primary source of chloride load to the middle St. Johns River.

Steady-state groundwater flow models are adequate for estimating groundwater discharge for use as boundary conditions to a surface water model. A dynamic surface water model of the middle St. Johns River was insensitive to replacement of steady-state groundwater discharge with transient values. Groundwater flows, and particularly their associated chloride loads, are important for surface water modeling, but for this system, temporally constant discharge and chloride values are adequate for model boundary conditions.

The assumption of constant density (and lack of constituent transport) is reasonable for groundwater flow modeling in the WSIS study area. Chloride-derived density differences do not affect the model's flow calculations significantly. Simulated vertical discharge would likely increase less than 3% if density differences were considered. Given that simulated groundwater flows are simulated correctly, a constituent transport model is not needed to estimate chloride loads. Chloride loads can be simply estimated as the product of spatially varying concentration and discharge outside of the groundwater model because chloride concentrations and salt composition do not vary in time.

Diffuse groundwater is the largest source of chloride to the middle and upper St. Johns River, providing approximately 937 t d^{-1}. In the middle St. Johns River upstream of U.S.17, diffuse groundwater supplies 75% of the chloride load to the river. Model estimates of chloride load are within 12% of observed loads for a large portion of the middle St. Johns River.

The existing groundwater flow models provide adequate boundary conditions for groundwater discharge and salinity to a surface water model of the middle St. Johns River. The assumptions of steady-state and constant-density used by the groundwater models are reasonable for this purpose. The addition of these boundary conditions allows the surface water model to simulate salinity and water age, in addition to river stage and discharge, to meet the larger objectives of the WSIS.

Finally, the groundwater flow models show that neither groundwater discharge nor chloride load is altered appreciably by lowering of river stage due to a proposed water withdrawal. This result is found because the hydraulic head in the aquifer that drives the groundwater discharge is much larger (3 to 6 m above the river surface) than the maximum expected reduction in river stage due to a withdrawal (5 cm).

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