Tracing Groundwater Flow on a Barrier Island in the North-east Gulf of Mexico

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Groundwater flow on St. George Island, a barrier island in the north-east Gulf of Mexico, was monitored downfield from wastewater systems using artificial tracer techniques. Sulphur hexafluoride and fluorescein dye were used to determine groundwater flow velocity, hydraulic conductivity, and dispersivity at selected sites on the island. Monthly hydraulic head measurements illustrate the aquifer’s dependence on recharge associated with rainfall. However, during periods of little rain, tidal stage also influences the direction and magnitude of groundwater flow within approximately 30 m of the waters edge. Estimated hydraulic conductivities ranged from 3 to 180 m day$^{-1}$, with an overall estimate of 36 m day$^{-1}$.

Groundwater tracers showed very little dilution and calculated longitudinal dispersivities were approximately 0.1 to 0.5 m, which is in the same range as previous studies of sandy aquifers.

The total groundwater flux into adjacent Apalachicola Bay was also evaluated using two independent techniques. Darcy’s law was applied using an estimated cross sectional area and the experimental horizontal transport rates to estimate the volumetric flow. In addition, a simple water balance calculation was used, which accounted for all known sources and sinks of water to the aquifer. The two independent approaches agreed very well, with an estimated groundwater flux from the surficial aquifer to Apalachicola Bay between $1 - 9 \times 10^6$ m$^3$ year$^{-1}$.

Keywords: groundwater; barrier island; coastal hydrology; estuary; coastal aquifer; Gulf of Mexico

Introduction

Submarine groundwater discharge (SGD) has been documented as being a significant source of nutrients to coastal waters (Valiela et al., 1978; Valiela & Teal, 1979; Capone & Bautista, 1985; Lapointe & O’Connell, 1989; Capone & Slater, 1990; Valiela et al., 1990; Corbett et al., 1999). The importance of groundwater discharge to the coastal environment is dependent on several variables, including the amount and type of nutrient enrichment in the groundwater, water column circulation, tidal flushing, porosity and permeability of the underlying strata, hydraulic head, and thus the corresponding groundwater flow. In areas with highly permeable soils, i.e. shallow coastal aquifers, groundwater may be a major pathway for the transport of anthropogenic contaminants, including nutrients, metals, and bacteria (Johannes, 1980; Weiskel & Howes, 1992). This is significant as coastal groundwaters contaminated with nutrients from onsite sewage treatment and disposal systems (OSTDS) have become more common and thus may promote eutrophication of nearshore ecosystems (Johannes, 1980; Nixon, 1986, 1992).

Apalachicola Bay is a highly productive bar-built estuary in the north-east Gulf of Mexico. St. George Island, like most barrier islands, forms the outer perimeter of the estuary and is critical to the bay’s productivity because its orientation determines the salinity distribution as well as other water quality features of the bay. The estuary is also an economic resource in north Florida, providing more than 90% of Florida’s oyster landings and the third highest catch of shrimp in the state (Livingston, 1984). Although the Apalachicola River is the major source of fresh water and nutrients to the estuary, increased population density on St. George Island has raised concerns of potential changes in local water quality and impacts on the oyster industry.

Barrier islands throughout the U.S. are experiencing similar increases in permanent and seasonal population. Many estuaries are sensitive to slight physical, chemical, and biological perturbations. Although barrier islands play a critical role in this balance, very little is known about the groundwater dynamics and potential impacts of contaminated groundwaters from these islands. A better understanding of groundwater flow within these islands can help establish responsible
comprehensive plans for local development and resource management. Conservative groundwater tracers were employed (fluorescein and sulphur hexafluoride) to evaluate the subsurface flow directions, velocity, and aquifer characteristics on St. George Island (Figure 1). Multi-level samplers (MLS) and 5-cm PVC monitoring wells were placed down-gradient from selected wastewater systems at three locations on the island. The wells were monitored over the course of the study for tracer concentrations along the flow path toward surface waters, as well as mapping of the piezometric surface at each location. In addition, the amount of groundwater entering the bay was estimated from two independent techniques, utilizing the tracer data and a simple water balance model calculation.

St. George Island

St. George Island is a microtidal barrier island in the panhandle of Florida. The island is approximately 48 km long and averages less than 0.5 km in width. The Dr. Julian G. Bruce State Park occupies the east end of the island. The climate in the region is mild, with a mean annual temperature of approximately 20 °C (Livingston, 1984). The mean annual rainfall over the area, recorded over the last 42 years by the NOAA weather station in Apalachicola, is approximately 140 cm. The rainfall recorded during the study period at Apalachicola was near the long-term average during each year. St. George Island receives less rain than that measured in Apalachicola. A new weather station located on the causeway leading to the island recorded an average of 63 cm year⁻¹ during this study. The tidal range in the bay and Gulf is typically less than 0.5 m.

The surficial aquifer on the island is composed of medium to fine sands overlying a silty clay barrier between 7.6 and 9.2 m below the surface that apparently forms a base to the aquifer (Livingston, 1984). Shallow auger holes (~3 m), dug during well construction, did not show significant lithologic variability at or between experimental sites on St. George Island. Water in the shallow fresh water lens is primarily derived from rainfall, and eventually discharges into Apalachicola Bay or the Gulf of Mexico. The clay layer may locally separate rain-derived freshwater from the surrounding salt water. St. George Island has a characteristically shallow water table (< ~0.5 m), which increases the probability of groundwater contamination and transport to surrounding marine waters.

Figure 1. St. George Island is a barrier island located in the north-eastern Gulf of Mexico. Samples were collected from three different experimental sites located on St. George Island. Rainfall data were collected from two weather stations indicated by open squares.
Study site

Experimental sites were selected according to location on the island, proximity to Apalachicola Bay, type of OSTDS, and the amount of time the residence were occupied. Care was given to locate sites adjacent to the bay with a similar beachfront, i.e. natural topographic gradients, no canals, etc. Three sites were chosen: a site on the far eastern end of the island located within the Dr Julian G. Bruce State Park (SP Site), a private residence near Bob Sikes Cut (BL Site) on the extreme western end of the island, and another private residence approximately 3 km west of the causeway (JA Site), between the other two locations. Each site had an average of three residents and was occupied all year round.

SP site

The experimental site within the state park was developed between August and October, 1997. Wells were installed primarily down-gradient from the septic system, and consist of twelve 5-cm monitoring wells and thirteen multi-level samplers (MLS) covering just under 8000 m² [Figure 2(a)]. The septic system is set back approximately 100 m from Apalachicola Bay.

BL site

The BL Site is also approximately 8000 m² with 60 m of bayfront access [Figure 2(b)]. The property is generally level and approximately 0 ·6 m above mean sea level. The septic system and drainfield is raised above ground level approximately 1 m to prevent inundation by storm waters during heavy rains. Wells were installed on either side of the drainfield and towards the bay waters during March, 1998. Monthly samples were collected from this site until March 1999, when the wells were removed due to construction on the property.

JA site

This site is located adjacent to Apalachicola Bay with approximately 45 m of bayfront access. The site is just under 6000 m² and has an aerobic wastewater system, installed in 1996. Due to the location of the house with respect to the OSTDS, wells were not positioned in a grid-type pattern as at the other sites. Wells were placed as close to the OSTDS and around the residence as possible. Fourteen wells including seven monitor wells and seven MLS were installed at the site during March 1998 [Figure 2(c)]. Samples were collected from this site until December, 1998, when the site had to be abandoned.

Methods

Well installation

Multi-level samplers and 5-cm PVC monitor wells were installed at all sites to an approximate depth of 3 m. Wells were installed using a hand auger with a 7·5 cm hollow barrel. To prevent the hole from back-filling during construction, a 10 cm PVC casing (outer-casing) was inserted into the hole and moved downward as the hole was dug deeper. Once the desired depth was reached (~3 m), the well was inserted into the hole, contained by the outer-casing. The outer-casing was then removed from the hole, allowing the aquifer materials to collapse around the sampler, thus isolating sampling points of the MLS at each level in the borehole. Initial studies on the island demonstrated that the in situ material was enough to isolate the well and it was not necessary to use borehole sealant above or between the sampling screens. Additional soil material, originally removed from the hole, was backfilled to complete the well as necessary. Wells were typically cut flush to the ground and covered with a removable 15 cm plastic cover. Broward Davis & Assoc., Inc surveyed the elevation of all wells relative to mean sea level.

Multi-level samplers.

Wells that use a slotted interval for sampling tend to provide integrated samples that are a mixture of different zones within the screened interval (Pickens et al., 1978). Nested wells or piezometers with short screens can be used to obtain samples from different depths, however this approach requires many boreholes and additional expense. The construction of a multi-level sampler allows for sampling of groundwater at closely spaced intervals in a vertical direction from a single borehole. The MLS device used in this study is a slight modification of wells used in previous groundwater studies (Pickens et al., 1978; Boggs et al., 1988; LeBlanc et al., 1991).

MLS devices were constructed using 1·9 cm OD PVC pipe and 0·6 cm OD polypropylene tubing. The PVC pipe acts as a housing for the smaller polypropylene sample tubing. A 3 m section of PVC pipe is first cut and one end (bottom) is capped. For this study, seven polypropylene tubes were attached to the outside of the PVC pipe by plastic cable ties or other fastening devices. The ends of the sampling tubes attached to the pipe were wrapped twice with 202 μm
Nyten mesh and spaced 30 cm apart. Enough tubing was left above the PVC pipe for easy access and sampling (~0.5 m). Upon installation of the well, the PVC pipe was filled with material removed from the borehole and then capped. Sample depths were identified at the top of each piece of tubing. Samples were collected directly from individual sample depths using a peristaltic pump.

**Monitor wells.** Wells constructed of 5-cm diameter PVC were installed at the experimental site for monitoring nutrient and tracer concentrations, as well as water table height. The wells were installed to approximately 3 m below ground-level. The last 0.4 m of the well consisted of slotted (254 µm) PVC screen. Water table height was monitored monthly using a Solinst model 101 water level meter. It was not necessary to adjust the head measurements to fresh water equivalent since water in the wells was consistently fresh throughout the study. Samples were collected using a peristaltic pump after purging three well volumes.

**Hydrologic tracers**

One of the most unequivocal ways to ascertain the rates and pathways through a hydrological system, hydraulic properties of an aquifer, and to link specific sites of contamination to discharge points is via artificial tracers. Ideally, tracers should have predictable properties, both intrinsically and in their interaction with the system into which they are introduced, i.e., tracers should be chemically stable, conservative, inexpensive, readily available, and easily detected. The large volume of most hydrological systems means that tracers need to be detectable at low concentrations and have low natural background concentrations.

**Fluorescein.** Sodium fluorescein (C20H10O4Na2), referred to as fluorescein dye or simply fluorescein, is an inexpensive highly water soluble fluorescent dye (Quinlan, 1989). Fluorescein is bright yellow-green to the eye and has a maximum excitation of 491 nm and maximum emission of 513 nm (Gaspar, 1987). Many groundwater tracing studies have employed this dye since it is inexpensive, easily detectable, non-toxic, and stable over time (Smart & Laidlaw, 1977; Smart, 1984; Duley, 1987; Quinlan, 1989; Reich, 1993).

However, the dye will break down if exposed to direct sunlight.

Samples were collected in 125-ml amber polycarbonate containers with a peristaltic pump. Samples were returned to the laboratory and analysed using a Turner Designs 10-AU-005 Fluorometer, which provides exact concentrations after calibration. The fluorometer was set up using the 10-045 daylight white lamp, 10-104R-C Combination Round Excitation Filter, 455-500 nm, 10-109R-C Combination Round Emission Filter, 510-700 nm, and the 10-063 Square Reference Filter as specified by manufacturer. The fluorometer was initially calibrated using fluorescein standards made in the laboratory.

**Sulphur hexafluoride.** SF6 was employed as a conservative tracer for subsequent experiments throughout the study period. Sulphur hexafluoride (SF6) is an inert water-soluble gas, is biologically and chemically inert, has a low background concentration (10^{-15} mol l^{-1}), and can be detected at extremely low concentrations (10^{-16} mol l^{-1}; Wanninkhof et al., 1985; Wanninkhof et al., 1991). SF6 has been used for gas exchange studies in rivers (Clark et al., 1994) and lakes (Wanninkhof et al., 1987; Upstill-Goddard et al., 1990; MacIntyre et al., 1995) as well as applications in atmospheric and oceanic sciences (Upstill-Goddard et al., 1991; Watson et al., 1991; Ledwell et al., 1993). The strong potential of SF6 as a geothermal and groundwater tracer has also been reported (Wilson & Mackay, 1993; Upstill-Goddard & Wilkins, 1995). In those studies, SF6 compared favourably with fluorescein dye applied in a 7.5 × 10^{-7} mass ratio of SF6 to sodium fluorescein.

Sulphur hexafluoride samples were collected in 30-ml serum vials with a peristaltic pump and extracted into a small headspace just before analyses. After purging the well, a sample was pumped into a serum vial and allowed to overflow for three bottle volumes. The vial was then sealed with a rubber septum and a crimp cap. To prevent loss of SF6 through the septa, the samples were stored on their sides until the samples could be extracted and analysed. Samples were extracted in the lab by adding a small headspace (typically 4 ml) of argon or ultra-high purity nitrogen to the sample. Simultaneously, a volume of water from the sample had to be removed.

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**Figure 2.** The SP study site (a) is located furthest east on the island and is within the Dr. Julian G. Bruce State Park. The SP site contains 12 5-cm monitoring wells (#1–12) and thirteen multi-level samples (#13–25). The BL study site (b) is located furthest west on the island and contains seven 5-cm monitoring wells (solid circle) and eight multi-level samples (x-circle). The JA site (c) contains seven 5-cm monitoring wells (solid circle) and seven multi-level samples (x-circle).
and discarded to allow for the headspace. The serum vials were slightly over-pressurized with 1 ml of nitrogen to allow several withdrawals for analysis (100 μl or less) by the gas chromatograph (GC).

SF6 samples were analysed with a Shimadzu model 8A gas chromatograph equipped with an electron capture detector. The gas chromatograph contained a stainless steel column (180 cm × 0.1 cm I.D.) packed with a molecular sieve 5A (80/100 mesh). Initially, a P5 mixture (95% argon, 5% methane) was used as a carrier gas with a flow rate of 25 ml min⁻¹. After having problems with carrier gas contamination, we switched to ultra-high purity nitrogen as a carrier at the same flow rate. Column and detector temperatures were set at 90 °C and 220 °C, respectively.

Headspace concentrations in ppmv (parts per million by volume, μl l⁻¹) of SF6 were determined by reference to a 1.04 ppm standard (Scott Specialty Gases). Headspace concentrations were converted to dissolved concentrations in μM using the ideal gas-equation. Replicates were collected for 10% of the samples. In addition, duplicate injections were run on the gas chromatograph every fifth injection. Precision between replicate samples and duplicate injections were usually better than 10%.

Results and discussion

Horizontal flow velocities

The water level relative to mean sea level (MSL) was monitored monthly throughout the study period in the 5-cm PVC monitoring wells at all the experimental sites. These measurements provide a snap shot of the piezometric surface and demonstrate how the hydraulic gradient changes over time. A representative hydraulic gradient for each site was calculated from water levels in the two inland-most wells to reduce the tidal influence at each site. During periods of little rainfall and continual evapotranspiration (ET), groundwater flow near the shoreline (within 15–20 m) is primarily influenced by tidal stage. In fact, groundwater flow can change directions during a 12-h period due to changing surface water levels (Figure 3). Piezometric surface maps from BL and JA sites during the early summer (June–July), 1998 suggested that the predominant groundwater flow direction near the wastewater systems was toward the interior of the island rather than the bay. It is evident then that the hydraulic gradient, and thus groundwater flow, is greatly dependent on the aquifer recharge associated with rainfall and ET. In fact, the similarities between rainfall and hydraulic gradient can easily be seen at all three sites (Figure 4). In addition to rainfall, seasonal changes in ET, higher during the summer months relative to winter, constrain the island’s hydrology and thus contaminant transport. During periods of low rainfall, contaminants introduced into the surficial aquifer would have additional time to be adsorbed, assimilated, or degraded, especially if the subsurface flow is toward the inner island rather than surface waters. Therefore, rainfall plays a very important environmental role on the island.

During October 1997, the 5-cm PVC monitoring wells at the SP site were monitored over a 24 h period to evaluate water level changes over a tidal cycle (Figures 3 and 5). The water level fluctuations in a well in response to changes in sea level may be used to determine certain aquifer characteristics (Carr & Van Der Kamp, 1969). Changes in the water level of a well in response to tidal loading occur as a result of mechanical loading of the aquifer at the oceanic extension, propagation and attenuation of the pressure wave created by the loading, and flow of groundwater from the aquifer to the borehole (Enright, 1990). For example, as sea-level increases, the aquifer bears a greater load creating a pressure gradient in the immediate vicinity of the loading. This pressure gradient (wave) is propagated inland and is attenuated since the matrix is bearing a portion of the load. Ferris (1963) showed the relationship between an aquifer’s transmissivity and the time delay that occurs between the tidal high or low and the maximum or minimum water level observed in an observation well at some distance from the tidal source and an aquifer’s transmissivity as

\[ T = \frac{x^2 S t_0}{4 \pi t^2}, \]

where \( T \) is the transmissivity (m² day⁻¹), \( x \) is the distance from the tidal source (m), \( S \) is the specific yield of the aquifer (dimensionless), \( t_0 \) is the period of the tidal signal (day), and \( t \) is the time delay to successive maxima or minima between the surface water body and the wells (day). The transmissivity \( T \) is the rate of groundwater flow through a vertical strip of aquifer one unit wide, extending the full saturated thickness of the aquifer, under a unit hydraulic head and is related to hydraulic conductivity by

\[ T = K b \]

where \( K \) is the hydraulic conductivity (m day⁻¹) and \( b \) is the saturated aquifer thickness (m). The specific yield is the ratio of the volume of water that drains by gravity to the total volume of aquifer.
Schwartz (1990) suggest that the specific yield for fine to coarse grain sands ranges from 0.23 to 0.28. We have assumed a conservative value of 0.20 for specific yield on all calculations. The thickness of the surficial aquifer used in these calculations is 8 m (average depth to clay horizon), based on data from boreholes (Livingston, 1984) and geophysical data (Ruppel, pers. comm.). Although this tidal approach is typically used in confined aquifers, the thickness of the surficial aquifer is more than 30 times the maximum change in tidal amplitude, allowing the assumption of confined conditions.

The lag time of the well response and the tidal attenuation increase with increasing distance from shore. Three wells (#1, #7, #9), which provided a transect from the shoreline to the residence [Figure 2(a)], were used to calculate a hydraulic conductivity based on the methodology described.
above. Results suggest that the hydraulic conductivity varies from 20–180 m day$^{-1}$ (Table 1). The largest (180 m day$^{-1}$) and the smallest (20 m day$^{-1}$) conductivities were observed in the wells furthest inland and shoreward, respectively. Schultz and Ruppel (1999) observed a similar trend on a barrier island off the coast of Georgia. Results from their tidal experiments and grain size analyses suggest that the hydraulic conductivity ranged from 0.07 m day$^{-1}$ nearest the surface water body to 12 m day$^{-1}$, approximately 35 m inland. Millham and Howes (1995) reported similar results from their study and earlier studies of coastal aquifers. They attributed the lower hydraulic conductivity along the shoreline to increased organic material in the sediment. The organic material is several times less dense than the inorganic matrix and occurs in smaller particle sizes, which could occupy pore spaces in the aquifer matrix. Unfortunately, sediments from St. George Island were not analysed for organic content, although we suspect that a similar trend in grain size does occur.

Tracer studies have also provided information about the hydrologic characteristics (dispersivity, hydraulic conductivity, etc.) of the experimental sites as well as providing groundwater flow paths and velocities. Two tracer experiments were conducted at the SP site, while only one experiment was conducted at each of the other two sites (Table 2). In each case, the tracer was dissolved in 190 l of tap water from the site. The tracer was then injected directly into a 5-cm PVC monitoring well (fluorescein) or directly into the drainfield (SF$_6$), bypassing the wastewater tank. The tracer was then followed by approximately 100 l of tap water, acting as a ‘chaser’. Wells were monitored at least monthly downfield from the injection sites over extended time periods (8–18 months).

The first experiment was conducted using fluorescein dye at the SP site. The dye was initially dissolved in approximately 190 l of tap water and injected on October 6, 1997 into SP-12, a 5-cm PVC monitoring well near the middle of the site. Wells were then monitored over an 18-month period. The tracer was only detected in two of the wells on the site, indicating very little vertical or horizontal dispersion of the plume. The tracer was first recorded in SP-17 after 30 days (Figure 6) at a depth of approximately 60 cm below mean sea level (MSL), about the same depth the tracer was injected into the aquifer. The tracer was also monitored at two additional depths from this well, approximately 5 cm and 150 cm below MSL. The tracer was never detected at either of these two depths at SP-17, nor at any other location or depth until much later. The tracer reached its peak concentration at the 60 cm depth (1660 µg l$^{-1}$) after 41 days (Table 3), diluted by 3 orders of magnitude. Based on these observations, the horizontal transport

![Figure 4](image_url)

**Figure 4.** Hydraulic gradient measured at the three experimental sites and rainfall show a similar trend. The Northwest Florida Water Management District (NWFWMD) measures rainfall on the causeway leading to the island. Hydraulic gradients were measured between pairs of wells furthest from surface waters to avoid tidal influence. The difference in rainfall between September 1997 and 1998 is primarily due to a more active hurricane season in 1998. Open bars: SP site; grey bars: BL site; black bars: JA site; circles: rainfall.
rate calculated for the tracer to this well is 0.11 m day\(^{-1}\). However, when this tracer was initially injected into the subsurface there was very little hydraulic head (gradient <0.001) driving the water (Figure 4). Soon after the experiment commenced, the island received enough rain to increase the hydraulic gradient by more than five times (0.0052) the initial gradient. A rate of 0.40 m day\(^{-1}\) results if the horizontal transport rate is calculated from the time of the rain event (2 November 1997). It is unlikely that the tracer had moved a significant distance from the point of injection before that rain event. The effective flow rate could have been still higher as there is no way of knowing the exact time following the rain event that the plume began moving.

The tracer was not detected in any wells until it reached SP-9, even though there are other monitor wells in-between the injection point and this well. This is probably due to very little vertical migration. Monitoring well SP-9 has a screened interval around 10–50 cm below MSL, approximately the same as the injection point. The two wells sampled between SP-9 and SP-17 have a deeper screened interval (>50 cm

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**Figure 5.** The hydraulic head of three monitor wells (open circles) relative to tidal changes (closed square) shows the attenuation of the tidal signal as it propagates inland. Well SP-9 (a) is located 9 m and well SP-7 (b) is located 39 m from the shoreline. Well SP-1 (c) is furthest inland at 97 m from the shore. Note the difference in scales of the well head as the tidal signal propagates inland. Closed squares: bay; open circles: well.

**Figure 6.** Results of tracer experiment in which fluorescein was injected into SP-12 [site map, Figure 2(a)] and wells downgradient were sampled over approximately 18 months. The tracer only appeared above detection limits in two wells, SP-17 (squares) and SP-9 (circles).

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**Table 1.** Results from the tidal experiment at the SP site conducted in October, 1997

<table>
<thead>
<tr>
<th>Well #</th>
<th>Distance from tidal signal (m)</th>
<th>Lag time (day)</th>
<th>Hydraulic conductivity (m day(^{-1}))</th>
</tr>
</thead>
<tbody>
<tr>
<td>Well #1</td>
<td>97</td>
<td>0.20 ± 0.06</td>
<td>180 ± 100</td>
</tr>
<tr>
<td>Well #7</td>
<td>39</td>
<td>0.13 ± 0.04</td>
<td>70 ± 40</td>
</tr>
<tr>
<td>Well #9</td>
<td>9</td>
<td>0.06 ± 0.02</td>
<td>20 ± 10</td>
</tr>
</tbody>
</table>

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**Table 2.** Summary of tracer experiments conducted on St. George Island

<table>
<thead>
<tr>
<th>Experimental site</th>
<th>Tracer employed(^{a})</th>
<th>Date injected</th>
<th>Injection concentration(^{b})</th>
</tr>
</thead>
<tbody>
<tr>
<td>SP site</td>
<td>Fluorescein</td>
<td>6 October 97</td>
<td>1.6 \times 10^6 \mu g l(^{-1})</td>
</tr>
<tr>
<td></td>
<td>SF(_{6})</td>
<td>23 June 98</td>
<td>1250 ± 270 nM</td>
</tr>
<tr>
<td>JA site</td>
<td>SF(_{6})</td>
<td>23 June 98</td>
<td>1360 ± 400 nM</td>
</tr>
<tr>
<td>BL site</td>
<td>SF(_{6})</td>
<td>23 June 98</td>
<td>1560 ± 190 nM</td>
</tr>
</tbody>
</table>

\(^{a}\)Tracers were injected directly into the wastewater drainfield, bypassing the wastewater tank, except for the fluorescein experiment. Fluorescein was injected into well SP-12.

\(^{b}\)The volume of injection was 190 l during each experiment. Concentrations were measured, not calculated.
Table 3. Times of peak tracer concentrations and horizontal transport rates (HTR) for each sampling location at the SP site during both experiments

<table>
<thead>
<tr>
<th>Tracer experiment</th>
<th>Sample location</th>
<th>Distance from injection (m)</th>
<th>Time (days)</th>
<th>Maximum tracer concentration*</th>
<th>HTR* (m day⁻¹)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Fluorescein</td>
<td>SP-17a</td>
<td>4.5</td>
<td>41</td>
<td>1660</td>
<td>0.11</td>
</tr>
<tr>
<td>6 October 97</td>
<td>SP-17b</td>
<td>4.5</td>
<td>12</td>
<td>1660</td>
<td>0.40</td>
</tr>
<tr>
<td></td>
<td>SP-9</td>
<td>55</td>
<td>410–450</td>
<td>210</td>
<td>0.12–0.13</td>
</tr>
<tr>
<td>SF6</td>
<td>SP-12</td>
<td>31</td>
<td>73</td>
<td>640</td>
<td>0.42</td>
</tr>
<tr>
<td>23 June 98</td>
<td>SP-21a</td>
<td>54</td>
<td>210</td>
<td>830</td>
<td>0.26</td>
</tr>
<tr>
<td></td>
<td>SP-21b</td>
<td>23</td>
<td>137</td>
<td>830</td>
<td>0.17</td>
</tr>
<tr>
<td></td>
<td>SP-23a</td>
<td>67</td>
<td>290</td>
<td>210</td>
<td>0.23</td>
</tr>
<tr>
<td></td>
<td>SP-23b</td>
<td>13</td>
<td>80</td>
<td>210</td>
<td>0.16</td>
</tr>
<tr>
<td>SF6</td>
<td>JA-2</td>
<td>3</td>
<td>175</td>
<td>240</td>
<td>0.02</td>
</tr>
<tr>
<td>23 June 98</td>
<td>JA-4a</td>
<td>7</td>
<td>230</td>
<td>453</td>
<td>0.03</td>
</tr>
<tr>
<td></td>
<td>JA-4b</td>
<td>4</td>
<td>55</td>
<td>453</td>
<td>0.07</td>
</tr>
</tbody>
</table>

*aConcentrations of fluorescein are in μg l⁻¹ and SF₆ are in nM.
*bHorizontal transport rates were calculated for each sampling location, except SP-21b, SP-23b, and JA-4b, by dividing the distance from the site of injection by the time of the peak tracer concentration at that sampling location. SP-21b, SP-23b, and JA-4b were calculated using the distance between the upfield well and the difference in time of maximum concentration between the upfield well.

The higher value corresponds to the average over the study period. Horizontal transport rates were calculated for wells SP-21 and SP-23. The time until the maximum concentration is presented as a range, since the actual peak was not measured.

below MSL), so the tracer could pass above the sampling interval without detection. Wells SP-7 and SP-8 (Figure 2) have a 50 cm screened interval starting approximately 150 cm below MSL. In addition, well SP-13 was not sampled during the fluorescein experiment. By the time the fluorescein dye reached detectable concentrations in SP-9, samples were only being collected monthly. The exact time of peak concentration is not known, since there was no detectable amount of fluorescein in the well one month (October 1998) followed by an easily detectable amount (210 μg l⁻¹) in the subsequent month (November 1998). Therefore, the horizontal transport rate is reported as a range (0.12–0.13 m day⁻¹), accounting for the possibility that the concentration may have peaked any time between the two sampling periods (Table 3). The dye was never detected in any of the MLS wells (SP-21, 22, 23, or 24) on either side of the centre transect (SP-12 to SP-9).

Sulphur hexafluoride was injected into the respective drainfields at all three experimental sites in June 1998. The tracer was not detected in any wells at the BL site throughout the study period. Although it is possible that the tracer could have passed above or below the monitoring wells surrounding the drainfield, we think this is highly unlikely as the closest wells to the drainfield were all MLS with several sampling intervals. The tracer could not have passed above the well, since the shallowest screened interval was just below the water table. In addition, it seems unlikely that the tracer would move to a depth of 3 m below the ground surface before moving horizontally, thus missing the well. Therefore, the loss of tracer is believed to have occurred due to the design of the drainfield. The elevated drainfield increases the residence time of the wastewater, and thus the trace gas, in the unsaturated zone. The SF₆ tracer may have degassed before ever reaching the water table, decreasing the concentration below the detection limit. The drainfields at the other two sites are not raised, reducing the amount of time the gas tracer may emanate from solution.

At the SP site, the tracer was found to be above the detection limit in several wells, with the highest concentrations measured in SP-12, SP-21, and SP-23 (Figure 7). The peak concentrations measured in wells SP-12 (640 nM) and SP-21 (830 nM) are diluted by less than a factor of two, equivalent to about the same volume as the chaser injected into the drainfield immediately following the tracer. The tracer concentration was only 210 nM at SP-23, which could be associated with dilution of the tracer or may indicate that the centre of mass of the plume did not move through the area of the well. In either case, there was not a significant amount of dilution of the tracer over a long period of time (>290 days). Horizontal groundwater velocities calculated for each well range from 0.16 to 0.42 m day⁻¹ (Table 3). Two estimates of velocity were calculated for wells SP-21 and SP-23. The higher value corresponds to the average over the
entire experiment, including periods of increased flow, i.e. heavy rainfall. The slower rate is indicative of the groundwater movement between two different wells. For instance, an average velocity for SP-21 was calculated to be 0.26 m day$^{-1}$ over the course of the experiment. However, between the period of time that the maximum concentration was measured at SP-12 and then reached SP-21, the groundwater velocity was only 0.17 m day$^{-1}$. Estimating the velocity between wells provides a more realistic flow field and may provide a more accurate estimate of aquifer characteristics, since the hydraulic gradient may change due to rain events in the time the tracer travels between wells. The hydraulic gradient ranged from 0.0005 to 0.0045 throughout this tracer study, with the lowest gradients measured during June and July 1998.

Results obtained from the JA site are similar to those observed during the first experiment at the SP site. When the tracer was injected in June 1998, there was a very small hydraulic gradient (<0.0008). In addition, the hydraulic gradient measured in early July 1998 was negative, moving water toward the centre of the island rather than toward the bay and the monitoring wells. During this experiment, the tracer was only detected in two wells (JA-2 and JA-4), and it is not certain if the tracer’s centre of mass moved through the area of the wells (Figure 8). If a horizontal transport rate is calculated from the time of injection to the time when the maximum concentration in the monitoring wells appeared, an average velocity of 0.02 m day$^{-1}$ and 0.03 m day$^{-1}$ is obtained for wells JA-2 and JA-4, respectively (Table 3). However, this includes the periods of very little transport (low gradient) and the time the groundwater was moving in the opposite direction and should, therefore, be considered a minimum for this period. Calculating the time between corresponding tracer maximum and distance between the two wells provides an estimated groundwater velocity of 0.07 m day$^{-1}$ between these wells. Again, there was very little dilution of the injected tracer, on the order of 3–5 times less than the injected value.

### Aquifer characteristics

Many aquifer characteristics may be obtained from tracer experiments like those performed on St. George Island. Estimates of such characteristics are needed before further predictive tools may be utilized. For instance, an estimate of the aquifer’s hydraulic conductivity may be calculated using the relationship

$$K = \frac{nv}{i}$$

where $v$ is the horizontal transport rate (m day$^{-1}$), $K$ is the hydraulic conductivity (m day$^{-1}$), $i$ is the hydraulic gradient (dimensionless), and $n$ is the sediment porosity (dimensionless). The porosity of St. George Island sediment is approximately 0.30–0.35, which was measured gravimetrically from samples collected during installation of the
monitoring wells. This agrees with values reported by Domenico and Schwartz (1990) for fine-medium grain sands. The average groundwater velocity is calculated using the tracer results. Unfortunately, the hydraulic gradients measured at the St. George Island sites have been shown to change dramatically throughout the course of the tracer experiments, increasing the error associated with this calculation. To reduce these errors, the hydraulic conductivity can also be calculated using the estimated groundwater velocity obtained between two monitor wells. In this case, the hydraulic gradient is constrained to the time period between the arrival of the maximum tracer concentration in the two wells, greatly reducing the associated error. Estimates of hydraulic conductivity using Equation 3 range from 4–5 to 75 m day\(^{-1}\) and 2–7 to 7·1 m day\(^{-1}\) for the SP site and JA site, respectively (Table 4). These values compare well with other studies of coastal sandy aquifers (Garabedian et al., 1991; Robertson et al., 1991; Mas-Pla et al., 1992; Millham & Howes, 1995) and with the estimates obtained from the tidal method at the SP site (Table 1). Although the tidal estimates of hydraulic conductivity appear slightly higher than the tracer results, we constrained the value to within an order of magnitude. It should be noted that the hydraulic conductivity in sands varies over 4–5 orders of magnitude, so we have been able to significantly reduce the uncertainty at these locations.

Hydrodynamic dispersion, another important aquifer characteristic, refers to the mechanical and diffusive mixing of solutes within the porous media. Diffusive dispersion is a consequence of a concentration gradient, while mechanical dispersion is mixing that occurs due to local variations in velocity around some mean velocity of flow (Domenico & Schwartz, 1990). Hydrodynamic dispersion may act to dilute the solutes, spreading them both vertically and horizontally, as they travel downgradient. Many studies have shown that the values of dispersivity are scale dependent, increasing with the scale of measurement (Theis, 1963; Fried, 1975; Lee et al., 1980; Pickens & Grisak, 1981; Gelhar et al., 1985). Zou and Parr (1994) described a mathematical approach to estimate longitudinal (same direction of flow) and transverse (perpendicular horizontally and vertically to flow) dispersivity using single well tracer breakthrough curves, like those shown in Figures 6–8, assuming there is prior knowledge of the flow direction and magnitude. Longitudinal (\(a_x\); m) and transverse (\(a_y\); m) dispersivity may be calculated using

\[
\begin{align*}
a_x &= \frac{v\Delta t^2}{4(\ln(R) - \Delta t)} \\
a_y &= \frac{y^2v\Delta t^2}{4B - A}
\end{align*}
\]

where

\[
\begin{align*}
A &= 4t(x^2 - v^2t_{max}^2)\ln R \\
B &= x^2\Delta t - v^2t_{max}^2\Delta t \\
R &= \frac{C_{max}t_{max}}{Ct}
\end{align*}
\]
and \( v \) is the linear velocity of the flow field (m day\(^{-1}\)), \( t_{\text{max}} \) is the time of maximum concentration (\( C_{\text{max}} \)) observed (day), \( t \) is the time of some observed concentration (\( C \)) at the same well as \( C_{\text{max}} \) (day), \( \Delta x \) is the difference between \( t_{\text{max}} \) and \( t \), \( x \) is the longitudinal distance from the injection point to the well, and \( y \) is the distance of the well off the centre transect. Using this approach, the longitudinal dispersivity at the SP site ranged from 0·1 to 0·5 m throughout both experiments (Table 4). The largest value obtained (0·5 m) was calculated for the well furthest from the injection site, perhaps as a result of the scale dependence of this parameter as described above. Longitudinal dispersivities calculated for JA site were an order of magnitude lower than those observed at the SP site (Table 4). However, the scale of observations at the JA site was much smaller (7 m) compared to that of the SP site (67 m), again showing a scale dependence of this parameter. These values of longitudinal dispersivity are comparable to those reported in several studies of sandy aquifers under a variety of field-scales (Figure 9; Gelhar \textit{et al}., 1985; Knopman & Voss, 1991; Garabedian \textit{et al}., 1991; Mas-Pla \textit{et al}., 1992) and suggest a relatively homogeneous aquifer.

Unfortunately, calculation of the transverse dispersivity had to be estimated since the tracer never consistently appeared in any well except for those along the centre transect. Assuming well SP-20 (5 m off the centre line) had a similar curve to SP-23, a maximum transverse dispersivity of 0·04 m was estimated. In any event, it can be inferred that the transverse dispersivity is relatively small, since the tracer only appeared in one well along each ‘picket fence’ (wells were spaced <5 m apart), including the fence 67 m from the injection site. The small dispersivities estimated for St. George Island indicate that very little dilution of unreactive contaminants occurs in the subsurface before being released into surface waters. This finding may be significant since regulators frequently rely on dilution for attenuation of wastewater contamination in groundwater (Robertson \textit{et al}., 1991).
Groundwater flux into Apalachicola Bay

Estimating the groundwater flux from St. George Island into Apalachicola Bay will allow estimates of nutrient fluxes from wastewater systems, assuming one can determine concentrations at the point of discharge. This is important, as it will offer more information for better management decisions in the future. We have used two independent techniques to evaluate the volume of groundwater exiting into the bay: (1) discharge estimates via tracer studies; and (2) a water balance approach. Tracer experiments have provided estimates of groundwater velocity and hydraulic conductivity at two different locations along the bay side of the island. Since this water must eventually discharge into Apalachicola Bay, an estimate of the volumetric flow rate \( Q \) (m³ year⁻¹) may be calculated, if the cross sectional area of the aquifer is known \( A \) (m²), by application of Darcy’s law

\[
Q = vnA
\]  

where \( v \) is the groundwater velocity (m year⁻¹) estimated from the tracer experiments; and \( n \) is the effective porosity. The cross sectional area of the aquifer was calculated by assuming the length of the island (48 000 m) and the approximate depth to the clay layer near mean high tide (8 m). The estimate of the volumetric rate is presented as a range assuming the island (48 000 m²) and the approximate depth to well installation and calculated dispersivities suggest a yearly basis. Using this assumption the groundwater balance and assuming the aquifer is at steady state on the island. We acknowledge that there are significant assumptions, a range in the groundwater movement was monitored on different locations along the island. The groundwater movement was monitored on different points on the island. We acknowledge that there are significant uncertainties associated with this assumption and have presented the estimate as a range in order to be more conservative. Using Equation 9 and the stated assumptions, a range in the groundwater flux from the surficial aquifer into the bay was estimated to be between \( 3-9 \times 10^6 \) m³ year⁻¹. Although this discharge is very small in comparison to the Apalachicola River, the largest river in Florida, which discharges an average \( 2.2 \times 10^8 \) m³ year⁻¹ (Fu & Winchester, 1994), it could provide a significant amount of nutrients and bacteria to local bodies of water, especially semi-enclosed embayments around the island.

Flux was also estimated using a simple water balance and assuming the aquifer is at steady state on a yearly basis. Using this assumption the groundwater flux can be calculated by totalling the sources to the surficial aquifer (rainfall and potable water from the mainland) and subtracting the sinks (evapotranspiration), accounting for the area of the island. The average rainfall measured on the causeway leading to the island during 1997 and 1998 was approximately 63 cm year⁻¹. Multiplying by the area of the island provides a flux of \( 15 \times 10^8 \) m³ year⁻¹. The total amount of water pumped over to the island in 1998 was \( 0.5 \times 10^6 \) m³ year⁻¹, less than 4% of the annual rainfall. Although some of the water pumped to the island is lost in transit, the uncertainty is minimal since the volume is so much less than the other parameters. Evapotranspiration is one of the most difficult parameters to obtain for a water balance. The Northwest Florida Water Management District has estimated the evapotranspiration for the island to be approximately 80% using a water-use demand model (AFSIRS Ver. 5.5, developed by A. G. Smajstrla, University of Florida). Again, we have presented our estimates as a range to try and include the uncertainties of each parameter. Assuming evapotranspiration is between 50–90% of the total rainfall (a value of \~50% has been applied in similar settings, Bokuniewicz & Pavlik, 1990), the total groundwater that must exit the aquifer to continue steady state conditions is between 2–8 \times 10^6 \) m³ year⁻¹. However, some of this groundwater moves toward and ultimately discharges into the Gulf of Mexico rather than Apalachicola Bay. Conservatively, it can be assumed that half of the groundwater moves in each direction, although the topography would suggest that more than 50% of the groundwater moves toward the bay. This balance results in a calculated groundwater flux into the bay of around \( 1-4 \times 10^6 \) m³ year⁻¹, nearly the same range as the \( 3-9 \times 10^6 \) m³ year⁻¹ estimate based on the tracer measurements.

Summary

The transport and ultimate discharge of groundwater into coastal zones may be a significant process in the geochemical and nutrient budgets of many nearshore marine waters, especially in areas served by onsite sewage treatment and disposal systems. In order to better understand the potential impacts of wastewater systems on surface water environments a better understanding of the groundwater flow and general hydrology of offshore barrier islands is needed. Groundwater movement was monitored on St. George Island, a barrier island in the Northeastern Gulf of Mexico, downgradient from three different onsite sewage treatment and disposal systems. It is hoped that results from this study may help in the development of better management plans to protect this and other delicate ecosystems.

Tracer experiments and results from a tidal experiment suggest that the hydraulic conductivity ranges
from about 3 to 180 m day$^{-1}$, with the higher values estimated for wells located further inland. Calculated longitudinal dispersivities suggest a relatively homogeneous surficial aquifer and that there is very little dilution of the OSTDS tracer plume as it moves downgradient. This is also evident by the small change in tracer concentration. Horzontal transport rates ranged from 0.01–0.42 m day$^{-1}$ and 0.02–0.07 m day$^{-1}$ at the SP and JA field sites, respectively. Differences in flow velocity are associated with large deviations of hydraulic head in response to tides and aquifer recharge associated with rain events. The flow velocities, as well as a simple water balance, were used to estimate the total flux of 1–9 × 10$^6$ m$^3$ year$^{-1}$ groundwater entering Apalachicola Bay.

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References


