Ground Water Storage Effect on Streamflow for a Southeastern Coastal Plain Watershed

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Abstract

Ground water is known to substantially affect streamflow in regions where the two are interconnected. Likewise, streamflow can rapidly recharge surficial aquifers. Because of this interaction, accurate predictions of streamflow require accurate data on the relationships between streamflow and shallow aquifer conditions. Ground water levels in the surficial aquifer and streamflow were examined for a six-year period for two small watersheds in south-central Georgia within the Coastal Plain Physiographic region of the United States. Precipitation for the six-year period was below average. The shallow water table was elevated during average precipitation conditions, but receded as precipitation decreased. During the three-year period from 2000 through 2002, the riparian buffer was fully saturated approximately 40% of the time. Lower available aquifer storage was observed for late winter and early spring due to higher rainfall and lower evapotranspiration (ET). During recharge periods, which generally coincided with the period from December to March, the ground water gradient from the upland to the stream was approximately equal to the 3\% land surface slope. During other times of the year, the gradient fluctuated dramatically as a response to individual precipitation events, streamflow, and focused recharge. Based on ground water discharging to the stream for 40\% of the year, ground water discharge accounted for approximately 4\% of the annual precipitation for the period from 2000 through 2002. Aquifer saturation, the fraction of the available aquifer storage filled by ground water, was found to have a substantial effect upon streamflow. During periods when the aquifer was fully saturated, observed storm-generated streamflow peak rates were greater and of shorter duration. Comparison of similar magnitude events occurring in 1998 indicated that peak discharge was up to three times greater under saturated aquifer conditions than under nonsaturated conditions. Watersheds with substantial pond acreage appear to extend storm-flow over longer durations. These data indicate the impact of aquifer and pond storage on peak and total discharge in coastal plain watersheds.

Introduction

In cases where substantial interconnections between ground and surface water exist, ground water conditions are known to have a significant impact upon streamflow. This is particularly true in the Coastal Plain Physiographic region of the United States, where many shallow surficial aquifers supply base flow throughout much of the year to regional streams. Variably distributed runoff initiated by inconsistent saturation in the soil in areas surrounding the streams can dramatically impact streamflow in this region.

The Coastal Plain Physiographic region extends from Virginia to east Texas and comprises about 70\% of the southeastern United States land area. The surficial aquifers of the Southern Coastal Plain include sandy upland soils and alluvial deposits that have high infiltration rates and transmissivity. Surficial aquifers in south-central Georgia range in thickness from 10 m at the top of the landscape to 2 m at the stream. Materials with very low permeability underlie these aquifers, which generally prevent the downward movement of ground water (Asmussen 1971; Asmussen and Thomas 1974).

Data collected from 10 coastal plain watersheds in Georgia, ranging in size from 2.6 to 1494 km\textsuperscript{2}, indicated that base flow could make up 58\% to 82\% of total annual streamflow in this region (Shirmohammadi et al. 1984). Shirmohammadi et al. (1984) hypothesized that the proportion of streamflow made up by contributions from...
shallow aquifers would remain constant relative to the size of the drainage. Other estimates obtained from ground water data in the region indicate substantially lower ground water contributions for lower order watersheds (Bosch et al. 1996). These data were obtained downslope from a 1.8 ha upland field in the coastal plain. Estimates of ground water contributions to streamflow at this site were just 3% of the water budget for the area, or ~30% of the total streamflow. The implication is that lower order streams may contribute less base flow than do higher order streams.

A study that examined 13 years of rainfall and streamflow in relation to ground water elevation data found available alluvial storage varies with both time of year and antecedent rainfall conditions (Shirmohammadi et al. 1986). The study examined three subwatersheds of the Little River experimental watershed near Tifton, Georgia (Lowrance et al. 1988; Sheridan et al. 1995), ranging in size from 16.7 to 49.9 km² and focused upon the available storage within the surficial aquifer. Available storage is defined here as the empty pore space available to hold ground water between the aquiclude first encountered below the land surface and the land surface itself. Lower available storage was observed for late winter and early spring due to higher rainfall and lower ET (Shirmohammadi et al. 1986), which led to higher ground water conditions. Available storage increased in the summer and autumn due to higher ET and lower rainfall. Available storage was also found to substantially impact peak discharge rates. For comparable storms, instantaneous peak discharge rates were higher by an order of magnitude for wet conditions than for dry conditions. The study found that watersheds with wider channel cross sections have greater storage and respond slower to rainfall, yielding lower instantaneous peak discharges than watersheds with better defined channel systems. Their data indicated a minimum ratio of saturated alluvial depth to total alluvial depth of 0.35 to 0.60 was required for watersheds to generate significant streamflow for any rainfall event. They concluded that the antecedent condition of the alluvial zone was the primary factor controlling hydrologic response from coastal plain watersheds (Shirmohammadi et al. 1986).

With expanding global populations, water quantity and quality have become increasingly important. The total maximum daily load (TMDL) program initiated in the United States in 1972 relies heavily upon determining average, maximum, and minimum streamflow rates (Bosch 2003a, 2003b). Accurate assessment of streamflow rates requires an understanding of the interactions between shallow ground water and streamflow. Recently, negotiations between the states of Georgia, Alabama, and Florida have brought the importance of these interactions to the political front. These negotiations have clearly demonstrated that it is impossible to separate surface and subsurface waters when dealing with water rights. Better characterizations of the interaction between ground and surface water are required for accurate determination of streamflow rates and equitable distribution of water resources.

The objectives of this research were to (1) relate ground water conditions, water table elevation, gradient, and flow rates, to climatic patterns; (2) relate observed stream base flow to ground water gradients and ground water flow rates; and (3) relate streamflow response to ground water storage and gradient for a second order stream network in the coastal plain.

Methods

The period of record examined for this study was from 1997 through 2002. A site at the University of Georgia Coastal Plain Experiment Station Gibbs Farm near Tifton, Georgia, was selected for this research (Figure 1). The study site consists of two watersheds, 57 ha and 47 ha (Table 1). Soils in the watershed consist of loamy sands, with Tifton loamy sand (Plinthic Kandiudults; fine loamy, siliceous, thermic) being the dominant soil type (Calhoun 1983). The Tifton soil contains subsurface horizons with reduced infiltration rates that percolate water and initiate lateral flow during wet conditions (Hubbard 1983). The Tifton soil contains 7% to 14% plinthite from 0.8 to 1.4 m, evidence of a perched water table in this zone during wetter periods of the year (Daugherty and Arnold 1982). During the year, the shallow aquifer water table varies from 0 to 7 m below the ground surface, depending upon landscape position. The watershed contains dense riparian buffers in the floodplain. These buffers extend up to 50 m from the stream. The uplands consist of tilled fields and some forest (Figure 1). Rainfall was measured with a recording rain gauge located within the southern watershed.

The region is in the outcrop area of the Miocene series, the Hawthorn Formation (Asmussen 1971). This formation is the geologic parent material and is overlain by Quaternary sands. The Hawthorn Formation is believed to be continuous and serves as an aquiclude in the Tifton Upland (Stringfield 1966). The alluvium of present-day streams is

![Figure 1. Gibbs Farm north and south watersheds.](image-url)
incised into the Hawthorn Formation (Asmussen 1971). Only major rivers traversing the Tifton Upland, however, exhibit incised channels. Streams originating within the Tifton Upland flow over broad alluvial valleys covered by riparian vegetation, and incised channels are nonexistent. The geology and soils of the Tifton Upland are conducive to lateral subsurface flow of water within the vadose zone that eventually contributes to streamflow.

An initial set of monitoring wells (P1 through P5, F1 through F3, and W8901 through W8912) were installed in the upland and the riparian buffer surrounding the stream network in 1992 and 1993 (Figure 2). Transects extending across the south (W8913 through W8916) and north (W8917 through W8920) sections of the streams were established in January 2000 to better record the relationship between streamflow and the shallow water table gradients.

During installation, the depth to the confining layer was measured for the upland wells (P1 through P5, and F1 through F3), and the wells along the south (W8913 through W8916) and north (W8917 through W8920) cross sections. The depth to the confining layer varied from ~4.5 m from the land surface at the top of the landscape (P2 through P4) to ~2 m next to the stream. The slope of the confining layer was ~2.5%, slightly less than the 3% surface slope. In combination with the land surface elevation, the water table elevation and the depth to the confining layer were used to calculate percent saturation at each of the wells:

$$\% \text{ saturation} = 100 \left( \frac{W-D}{G-D} \right)$$

where $W$ is the water table elevation, $D$ is the confining layer elevation, and $G$ is the ground surface elevation. As porosity decreases in the aquifer, it requires less water to saturate the profile.

The wells were configured with pressure transducers and connected to data loggers for continuous monitoring of the water level. Water table elevations were adjusted back to a local benchmark (NGVD29). The system consisted of Druck pressure transducers (part no. PDCR 940 with a 0 to 5 psig range, Druck Inc., New Fairfield, Connecticut) monitored with Campbell CR10 recorders (Campbell Scientific Inc., Logan, Utah) on a half-hour basis. The measurement accuracy was checked twice per year to maintain data quality and corrected if necessary.

Stage recorders were installed at all surface runoff measurement sites in November 1997. At sites S21 and S23, circular culverts confined streamflow under road crossings (Figure 1). Culvert diameters were 0.66 m at S21 and 0.82 m at S23. Rating curves for the culverts were determined using Manning’s equation for open channel flow with circular cross sections with flow less than the culvert diameter and the Bernoulli equation for full flow (Brakensiek et al. 1979). Streamflow measurements were made and the formulas corrected for variations in roughness. Adjustments were made for flows that overtopped the road crossings using elevation data collected at the site. For these events, the road was assumed to act as a weir and the discharge added to the culvert flow (Kindsvater 1964). Wooden weirs were installed across the stream sections at S22 and S24. Site S22 contained a combination v-notch and rectangular weir with the low flows being confined to a 1.82 m wide v-notch with 1:10 side slopes. Flows with heads in excess of the 0.18 m deep notch

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**Table 1**

<table>
<thead>
<tr>
<th>Land Use</th>
<th>Gibbs South - S21</th>
<th>Gibbs North - S23</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>ha (%)</td>
<td>ha (%)</td>
</tr>
<tr>
<td>Ponds</td>
<td>3.2 (5.5%)</td>
<td>0.8 (1.7%)</td>
</tr>
<tr>
<td>Fields</td>
<td>28.9 (50.6%)</td>
<td>32.0 (68.4%)</td>
</tr>
<tr>
<td>Forest</td>
<td>24.2 (42.4%)</td>
<td>13.5 (28.9%)</td>
</tr>
<tr>
<td>Roads</td>
<td>0.8 (1.5%)</td>
<td>0.5 (1.0%)</td>
</tr>
<tr>
<td>Total</td>
<td>57.1 (100%)</td>
<td>46.8 (100%)</td>
</tr>
</tbody>
</table>

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Figure 2. Detailed illustration of well and streamflow instrumentation within the study watersheds in Figure 1 (elevations in M.S.L. NGVD 29).
at S22 were additionally confined within a 3.1 m rectangular weir. Site S24 was initially a combination v-notch and rectangular weir, with the low flows being confined to a 0.91 m wide v-notch with 1:5 side slopes. Flows with heads in excess of the 0.18 m deep notch at S24 were additionally confined in a 1.6 m rectangular weir. This weir washed out in September 1998 by runoff generated by Hurricane Earl. The weir was replaced in November 1998 with a combination v-notch and two rectangular weirs with the low flows being confined to a 2.62 m wide v-notch with 1:5 side slopes. Flows with heads in excess of the 0.26 m deep notch were additionally confined in a 2.62 m wide rectangular weir and flows with heads in excess of 2.67 m were routed through a second 2.62 m wide rectangular weir. Flow equations for the weirs were determined from Brakensiek et al. 1979. The stage at each site was measured with the same system used at the well sites. Initially, five-minute readings were recorded at the streams; however, this was changed in early 1999 to 10-minute readings.

Estimates of saturated hydraulic conductivity ($K_s$) of the aquifer materials were determined using procedures described in Bosch et al. 1996. Pump tests were conducted at 32 well sites, 18 in the upland and 14 in the riparian buffer. The pump tests yielded a single average $K_s$ for the saturated thickness. Texture characterizations of the vadose zone in the riparian buffer indicate an increase in clay content and a decrease in $K_s$ 1 to 2 m below the ground surface (Bosch et al. 1994).

Results

Precipitation

Precipitation totals for 1997 and 1998 were above or near average, but below average from 1999 to 2002 (Table 2). Monthly precipitation and cumulative departure from average precipitation were determined for the six-year period from 1997 through 2002 (Figure 4). The cumulative departure from average precipitation was calculated as the cumulative difference between the observed daily precipitation and the long-term daily average. Potential annual ET for the area is ~1354 mm (Hoogenboom 2003). Potential ET exceeds precipitation in all months except December through March.

Annual precipitation deficits were substantial from 1999 through 2002 (Table 2). The greatest deficits occurred from April to August. Regional potential ET demands are high (Table 2) and most precipitation is expected to be taken up by ET during this period. Exceptions are observed during large thunderstorm events that may generate surface runoff. Large deviations from average precipitation were observed from October 1999 to June 2000 and again from October
2001 to June 2002. The month of July is typically a month when greater precipitation is observed (Table 2); however, July totals were also low in 1997, 1999, 2000, and 2001. The deviation from the long-term average precipitation for the entire six-year period was 892 mm. From January 1997 to September 1998, cumulative rainfall was 410 mm above average, while from October 1998 until September 2002, the cumulative rainfall was 1302 mm below average.

### Table 2
Monthly Precipitation Totals at the Site

<table>
<thead>
<tr>
<th></th>
<th>1997 (mm)</th>
<th>1998 (mm)</th>
<th>1999 (mm)</th>
<th>2000 (mm)</th>
<th>2001 (mm)</th>
<th>2002 (mm)</th>
<th>Long Term Average Precipitation (mm)</th>
<th>Long Term Average Potential ET (mm)</th>
</tr>
</thead>
<tbody>
<tr>
<td>January</td>
<td>133</td>
<td>117</td>
<td>112</td>
<td>92</td>
<td>36</td>
<td>79</td>
<td>115</td>
<td>56</td>
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<tr>
<td>February</td>
<td>204</td>
<td>118</td>
<td>50</td>
<td>56</td>
<td>8</td>
<td>41</td>
<td>105</td>
<td>72</td>
</tr>
<tr>
<td>March</td>
<td>51</td>
<td>228</td>
<td>40</td>
<td>131</td>
<td>249</td>
<td>97</td>
<td>127</td>
<td>94</td>
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<tr>
<td>April</td>
<td>70</td>
<td>92</td>
<td>28</td>
<td>34</td>
<td>28</td>
<td>69</td>
<td>97</td>
<td>135</td>
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<tr>
<td>May</td>
<td>80</td>
<td>91</td>
<td>65</td>
<td>6</td>
<td>33</td>
<td>13</td>
<td>88</td>
<td>160</td>
</tr>
<tr>
<td>June</td>
<td>81</td>
<td>26</td>
<td>222</td>
<td>50</td>
<td>142</td>
<td>58</td>
<td>114</td>
<td>171</td>
</tr>
<tr>
<td>July</td>
<td>89</td>
<td>151</td>
<td>89</td>
<td>110</td>
<td>79</td>
<td>183</td>
<td>141</td>
<td>167</td>
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<tr>
<td>August</td>
<td>39</td>
<td>92</td>
<td>62</td>
<td>67</td>
<td>53</td>
<td>64</td>
<td>121</td>
<td>149</td>
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<tr>
<td>September</td>
<td>163</td>
<td>267</td>
<td>108</td>
<td>281</td>
<td>86</td>
<td>84</td>
<td>87</td>
<td>119</td>
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<tr>
<td>October</td>
<td>153</td>
<td>2</td>
<td>46</td>
<td>21</td>
<td>33</td>
<td>119</td>
<td>57</td>
<td>102</td>
</tr>
<tr>
<td>November</td>
<td>185</td>
<td>33</td>
<td>29</td>
<td>103</td>
<td>15</td>
<td>132</td>
<td>59</td>
<td>73</td>
</tr>
<tr>
<td>December</td>
<td>205</td>
<td>45</td>
<td>65</td>
<td>90</td>
<td>18</td>
<td>89</td>
<td>115</td>
<td>57</td>
</tr>
<tr>
<td>Annual total</td>
<td>1451</td>
<td>1261</td>
<td>916</td>
<td>1042</td>
<td>780</td>
<td>1026</td>
<td>1228</td>
<td>1354</td>
</tr>
</tbody>
</table>

Deviation from term average | 223 | 33 | -312 | -186 | -448 | -202 | — | —
Water Table Response

Observed water table elevations were compared to precipitation patterns. An example of this comparison is illustrated with the data from well W8902 (Figure 5) which is located just inside the wood-line (Figure 2). During the wetter years, the water table elevation in the riparian buffer increased from November to April and receded from May to October (Figure 5). The water table fluctuated from high conditions during the winter months to low conditions during the summer months, responding to ET demands. At W8902, a 1.5 m difference in water table elevation was observed between high conditions and low conditions (Figure 5). While the water table elevation still increased during the winter during years where precipitation was low, the decrease in the subsequent spring was more evident (1999–2000). In addition, ground water recharge was delayed from November to February or March.

Large rainfall events were observed in September 1998, June 1999, and September 2000. During these periods, the water table rose dramatically; however, it was followed by an equally dramatic drop in the water table when this recharge was observed during the summer months. This appears to be an indication of rapid recharge followed by an equally rapid period of ground water discharge and/or plant uptake. Conversely, if higher rainfall was received during the late fall and winter months, the water table remained high (December 1997 and March 1998). This period of recovery coincides with the period during which the potential ET demands are low (Table 2). Recharge is expected to occur in December through March when precipitation exceeds potential ET for the area.

Figure 5. Well W8902 water table elevation and cumulative departure from the average precipitation from 1997 through 2002.

Figure 6. Water table elevations at wells P2 (ground surface at 31.1 m) and W8913 (ground surface at 27.4 m) from 2000 through 2002.
Ground Water Gradients

Gradients from the upland wells into the riparian buffer were examined for the period of record. Wells W8913 through W8916 related to flows recorded at sites S21 and S22, while wells W8917 through W8920 related to flows recorded at sites S23 and S24. As indicated by water tables at wells P2 and W8913, a gradient from the upland to the riparian buffer existed throughout most of the three-year period from 2000 through 2002 (Figure 6); however, periods exist where the gradient is reversed. During these periods, ground water flows from the vicinity of the stream into the riparian buffer and upland.

On July 22, 2000, 58 mm of rain was recorded at the site. Following this rainfall, a 0.75 m water table increase was observed at well W8913, while a 0.4 m increase was observed at well P2. The increase at W8913 was observed within one day following the precipitation, while the increase at P2 was over a three-week period. During a four-day period from September 3 through September 6, 2000, 143 mm of rain fell at the site. A 1.3 m increase was subsequently observed in the water table at W8913 and a 3.3 m increase was observed at P2. Over a one-month period from December 10, 2001, to January 12, 2002, 70 mm of rain fell at the site. A 1 m increase was subsequently observed in the water table at W8913 and a 2 m increase was observed at P2. Sustained long-term precipitation appears to lead to substantial upland recharge, whereas the riparian buffer appears to respond more rapidly to single, large-volume precipitation events. The data indicate the buffer region responds to infiltration of surface runoff from the upland area and bank overflow from the stream. This is also a function of the degree of saturation within the buffer. Under saturated conditions, no recharge can occur, leading to larger streamflow events.

Surface depressions within the riparian buffer capture and hold surface runoff during large precipitation events, leading to focused recharge at these depressions. During periods of greater precipitation, as was observed in July 2000 and January 2002, the water table within the riparian buffer appeared to be responding to such recharge rather than ground water flow from the upland. Because a large volume of surface runoff can be generated from the upland, this can lead to periods of reversed gradient. When the water table in the riparian buffer rises rapidly in response to focused recharge and streamflow, the gradient may reverse, leading to flow from the near-stream area into the upslope portions of the buffer.

During periods when the water table was low and significant precipitation occurred, the water table within the riparian buffer surrounding the stream reacted rapidly to focused recharge and streamflow. In the case of the December 2001 to January 2002 water table increase, the water table responded to smaller, but sustained, precipitation events. In some cases, as observed in September 2000 and early 2002, the water table within the buffer remained high. This appears to have been a function of the initial recharge provided by the stream and ground water flow from the upland. Because the ground surface surrounding the riparian wells is low, flooding in the stream leads to focused recharge at the well sites as seen for January and February 2002 (Figure 6). In these cases, the water table elevation within the buffer rose more rapidly than that in the upland.

The water table at the wells along the north watershed stream behaved in a similar fashion to those along the south watershed stream. Ground water gradients in this watershed were not, however, as large as those observed in the south watershed (Figures 6 and 7). The water table elevation at wells P4 and W8920 responded to the same events as did the south corridor wells (Figure 7). Recharge occurred from July to September 2000, December 2001 to January 2002, and again at the end of 2002. Focused recharge and/or streamflow again appear to lead to ground water recharge along the stream.

During recharging periods, when the ground water in the upland and the riparian buffer is high, the water table is within 1 to 2 m of the land surface in the upland landscape position and at, or near, the land surface in the riparian buffer. The ground water gradient during these conditions is ~3%. This is approximately equal to the land surface slope. During dry conditions, the gradient ranges from 1% to 0%, and occasionally becomes negative (Figures 6 and 7). Because greater increases are observed near the stream, the gradient from these locations can be higher, directing flow toward both the stream and upslope toward the field edge.

Aquifer Saturation

The confining layer in the upland was ~5 m below the land surface, while it was ~2 m below the land surface in the riparian buffer. The available aquifer thickness at well P2 was 4.6 m, while it was 1.7 m at well W8913. Throughout the observation period, the aquifer saturation varied from 75% to 0% at well P2. During the same period, the aquifer saturation at well W8913 varied from 100% to 0%. During the three-year period from 2000 through 2002, 100% saturation, with the water table at the land surface, was observed at W8913 for approximately 40% of the time, or 146 days per year. Slightly lower aquifer saturations were observed along the northern transect. In the riparian buffer, the aquifer was close to 100% saturation from August 2000 to May 2001. During this period, very little storage remained for ground water flow from the upland or for infiltration from precipitation events.

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Ground Water Discharge

During periods of recharge, when the ground water is consistently flowing toward the stream, the height of the water table at the wells nearest the stream remains at, or near, the land surface. The flow depth in the stream remained at, or near, the channel depth of 0.5 m. During high water table conditions, the ground water gradient over the 90 m from well P2 to the stream on the south transect ranged from 3% to 4%, or approximately equal to the surface slope. Along the north transect, the ground water gradient over the 100 m from well P4 to the stream ranged from 1% to 2%.

Pump tests yielded $K_\text{s}$ values for the upland wells from 1.98 to 0.03 m/day, while values in the riparian buffer varied from 1.85 to 0.03 m/day (Bosch et al. 1996). The average for 32 measurements was 0.66 m/day and the standard deviation was 0.62 m/day. These values agree with other measurements collected in similar alluvial material (Shirmohammadi et al. 1986; Stephens et al. 1968). For the surface material from 0 to 0.23 m in the riparian buffer, $K_\text{s}$ averages 1.44 m/day (Bosch et al. 1994); however, considerable variability occurs with depth.

Ground water flow to the stream can be calculated based on the Dupuit assumptions (Todd 1980). We assumed the ground water discharged freely into the stream from both stream-banks and the gradient and streamflow depth remained constant during recharge. If a gradient of 3% and an average $K_\text{s}$ of 0.7 m/day is used for the recharge periods, flow rates during these periods would be 0.16 m$^3$/day/unit width of stream. The stream length in the south watershed is ~900 m, while it is 800 m in the north watershed. Assuming this discharge exists along the entire stream section, the average daily discharge during periods of flow from the ground water to the stream would be 142 $m^3$ for the south watershed and 126 $m^3$ for the north watershed.

Saturated conditions surrounding the stream lead to seepage faces near the stream ~40% of the time, or 146 days/year. This saturation generates seepage faces near the stream. Shirmohammadi et al. (1986) found similar results. They further found this saturated period most frequently occurred from January to May. Based upon the saturation data, a period of 146 days of seepage was used, yielding estimates of base flow on an area basis of ~40 mm per year from each watershed. This is ~4% of the 2848 mm of precipitation received from 2000 through 2002, the period over which data were collected. This saturation generates seepage faces near the stream ~40% of the time, or 146 days/year.

As expected, sustained periods of surface water runoff occurred primarily during the period from January to March, when ET is low and ground water is at, or near, the land surface in the riparian buffer. Sustained flow was also observed throughout the remainder of the year during periods of greater precipitation. These data agree well with other observations collected in the region. Longer periods of streamflow occurred during high water table conditions when the aquifer in the riparian buffer was close to 100% saturated. During these periods, base flow was providing sustained streamflow.

The peak flow observed at S21 in the southern watershed occurred on March 8, 1998, when the flow stage reached 1 m above the culvert inlet. This flow resulted from 32 mm of precipitation on March 7 and 140 mm on March 8. The precipitation was fairly constant over a 16-hour period from 0700 to 2300 on March 8, with the peak rate (6.6 mm over a five-minute period) observed at 0940. The peak flow observed for the event was 7.6 m$^3$/sec, observed at 1330 on March 8. The event generated approximately three days of storm-flow. At S23 in the northern watershed, the peak flow observed for this event was 8.5 m$^3$/sec at 1325 on March 8. The storm-flow duration at S23 was two days. At the time of the March 8 event, the water table was high and the aquifer fully saturated. There was no available storage within the aquifer. Runoff was directed to the stream quickly and flowed off the watershed in a fairly short period.

A similar volume of precipitation fell on September 3, 1998. The September 3 event was the largest precipitation event over the period of record, with 19 mm on September 2 and 148 mm on September 3. Precipitation began at 1745 on September 2, peaked (6.6 mm over a five-minute period) at 0245 on September 3, and ended at 1240 on September 3. The greatest volume of the precipitation fell from 0100 to 1100 on September 3. The peak flow observed at S21 from this event was 2.6 m$^3$/sec, observed at 0450 on September 3. This event generated approximately eight days of storm-flow at S21. The peak flow observed for this event at S23 was 9.2 m$^3$/sec at 0340 on September 3. The observed storm-flow duration at S23 was approximately eight days. At the beginning of this event, the water table at W8902 was 1.2 m below the land surface; however, the water table rose to the land surface on the day of the event and receded back to 0.8 m below the land surface by September 13.

The peak flow observed at S21 from the March 8 event was approximately three times as large as that observed for the September 3 event. In addition, the streamflow generated by this event occurred over a much shorter duration, the result of little aquifer storage. Aquifer storage spreads out the discharge over a longer duration by storing water in the ground before it discharges into the stream. While both of these events peaked in a fairly short time, the peak was much larger for the event which occurred under saturated conditions. In contrast, the peak flows observed at the northern watershed for these two events were similar in magnitude. The lower acreage in ponds (Table 1) on this watershed would lead to more rapid and less sustained discharge from this watershed. This is consistent with what was observed during other rainfall events on these watersheds. The large and relatively equal peak flows observed between the March 8 and September 3 events in the northern watershed indicate little available storage during either event.
Discussion

Estimates of ground water discharge from the upland to the stream during the study period varied from 4% to 18% of the precipitation observed over the study area. These estimates agree closely with previous observations from the area (Bosch et al. 1996), but are considerably lower than base flow estimates obtained from hydrograph separation techniques applied to watershed flow observations from the Little River Watershed (Shirmohammadi et al. 1984), where they found that base flow generated from ground water contributed, on average, 70% of total annual streamflow. Long-term hydrologic records from the Little River Watershed indicate that annual streamflow in the region accounts for ~30% of the precipitation (Sheridan 1997). Using these data, base flow estimates for the watershed account for 21% of the precipitation. The unusually dry conditions during our study may have led to ground water yields below average. Wetter conditions would have led to a longer period of saturation in the riparian buffer, which would lead to greater ground water contributions to streamflow. In addition, the relatively small upland area in this watershed may also lead to less ground water flow into the stream. The relatively short stream length led to low estimates of ground water recharge. Longer stream lengths would yield estimates more in line with watershed scale observations.

Significant contributions to base flow are also made by focused recharge from surface runoff generated in the upland. Data collected near the study site indicate surface runoff may be as high as 24% of precipitation from conventionally tilled upland fields (Bosch 2003, unpublished data). This is considerably greater than the 5% to 13% observed at the watershed scale (Shirmohammadi et al. 1984). The difference between these two appears to be surface runoff which infiltrates within the riparian buffer, later translating into short duration ground water flow.

Watershed observations indicate 5% to 13% of streamflow is derived from direct surface runoff (Shirmohammadi et al. 1984). Runoff data collected at this site indicate as much as 44% of the upland surface runoff infiltrates once it reaches the edge of the riparian buffer (Sheridan et al. 1999). Under unsaturated conditions within the vadose zone, this runoff is captured in depressional areas within the buffer, leading to focused recharge. As demonstrated by the ground water data (Figures 5, 6, and 7), this focused recharge leads to short duration, high water table conditions throughout the riparian buffer. While some of this water redistributes throughout the unsaturated zone, a portion also discharges into the stream. Our data indicate this focused recharge may make up a large component of the base flow. The controlling factor determining whether the focused recharge would generate streamflow would be the saturation in the alluvial material surrounding the stream. Under saturated conditions the water table would rise to levels capable of generating streamflow, while under dry conditions it would not.

Geologic data collected on the Little River Watershed indicate that deposited sands (Asmussen 1971) may dominate the east side of many of the coastal plain streams. While these sands may not make up a large fraction of the area of the watersheds, they may provide the conduits necessary for considerable ground water flow to the stream. The hydraulic conductivities of these sands are expected to be one or two orders of magnitude higher than the hydraulic conductivities measured in this study. Given the same gradients as those observed, these areas of higher conductivity would be expected to increase the ground water flow to area streams considerably, and could feasibly yield base flow values as large as those observed in previous watersheds studies.

Variability of $K_s$ with depth would also significantly influence the ground water yield. Previous studies have found that $K_s$ varies from 0.7 m/day in the saturated zone to 1.4 m/day in the near surface soil (Bosch et al. 1994). As the aquifer nears the soil surface, the ground water would flow through these zones of higher conductivity, increasing the volume of water flowing to the stream. While the saturated thickness for these layers would be small (0.25 to 1 m), they could still make substantial contributions to base flow during fully saturated conditions.

The water table in the riparian buffer was relatively unchanged by the March 8, 1998, event, with the water table near the land surface throughout the period, but rose dramatically during the September 3, 1998, event. These two events illustrate the impact of the aquifer saturation. Because the water table was fully saturated for the March event, the surface runoff event occurred over a fairly short time frame. Data collected from larger watersheds indicate that the majority of the streamflow (73%) occurs in the first four months of the year (Sheridan 1997) during this period of greater aquifer saturation. For the September event, the aquifer saturation was close to 50% and available storage was high. Within the southern watershed, the peak flow for the March event was approximately three times greater than that observed for the September event. Similar relationships between available aquifer storage and peak surface discharge have been found for the region. Shirmohammadi et al. (1986) found that for comparable storms, peak discharge rates were higher by an order of magnitude for fully saturated conditions than for dry conditions.

Our data also indicate that ponds within the watershed may dramatically impact peak and total discharge. Peak flows within the southern watershed, where more pond acreage exists, were substantially reduced during the drier summer months by the storage provided in the aquifers and the ponds. Peak flows within the northern watershed, where less pond acreage exists, did not appear to be reduced as substantially during these drier periods. Examination of the other larger precipitation events indicates similar ground water-streamflow relationships; however, during periods of very low saturation (0% to 25%), more of the surface runoff went toward filling the ground water without continued generation of streamflow. The discharge for these events were short duration (one to two days).

Conclusions

The University of Georgia Coastal Experiment Station Gibbs Farm study site received 110% of the average annual precipitation during the two-year period from 1997 through 1998 and 76% of the average annual precipitation during the four-year period from 1999 through 2002. The water table at the site responded accordingly, with higher water table conditions from 1997 to 1998 and reduced water table...
conditions from 1999 to 2002. While periods of higher water table conditions were observed during the years with lower precipitation, the water table dropped rapidly as ET increased. Available storage in the surficial aquifer varied from 100% to 25% in the upland and from 100% to 0% in the riparian buffer. During the three-year period from 2000 through 2002, the riparian buffer was fully saturated ~40% of the time; however, these observations were made during drought conditions. Greater saturation would be expected during average precipitation patterns. Lower available storage was observed for late winter and early spring due to higher rainfall and lower ET. Available storage increased in the summer and autumn due to higher ET and lower rainfall. As storage in the surficial aquifer decreased, the likelihood of direct surface runoff to the stream increased.

When the aquifer is not fully saturated, the water table in the riparian buffer rises dramatically in response to greater rainfall; however, an equally dramatic drop in the water table follows this rise if this recharge occurs during the summer months. This appears to be an indication of rapid recharge followed by an equally rapid period of ground water discharge and ET. During these periods, surface runoff in the upland leads to focused recharge within the riparian buffer. As the localized mound in the ground water dissipates into the surrounding area, the water table elevation recedes.

Discharge in the small watersheds was found to be impacted by aquifer saturation, precipitation volume and rate, and pond storage. Comparison of similar magnitude events occurring in 1998 indicated that peak discharge was up to three times larger under saturated aquifer conditions than under nonsaturated conditions. The watershed without the available pond storage did not exhibit these differences, indicating a pond effect as well. Available saturation was also found to affect flow duration. If more aquifer storage was available, the duration of the discharge event was increased. Greater aquifer saturation led to more intense, shorter duration runoff events.

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References


