Advective pore water input of nutrients to the Satilla River Estuary, Georgia, USA

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Abstract

In situ benthic flux measurements, pore water nutrient profiles, water column nutrient distributions, sediment grain size distributions and side-scan sonar observations suggest that advective transport of pore waters may be a major input pathway of nutrients into the Satilla River Estuary (coastal Georgia, USA). In situ benthic chamber incubations demonstrate the occurrence of highly variable, but occasionally very large sea floor fluxes of silicate, phosphate, and ammonium. Locally occurring benthic microbial mineralization of organic matter, as estimated by S35-sulphate reduction rate measurements, is insufficient to support these large fluxes. We hypothesize that the observed interlayering of permeable, sandy sediments with fine-grained, organic-rich sediments in the estuary provides conduits for advective transport of pore water constituents out of the sediments. Because permeable layers may extend significant distances beneath the salt marsh, the large fluxes observed may be supported by remineralization occurring over large areas adjacent to the estuary. Advective transport may be induced by pressure gradients generated by a variety of processes, including landward recharge by meteoric or rain waters if sand layers extend far enough into the maritime coastal lands. Alternatively, tidal variations across the salt marsh sediment surface may hydraulically pump water through the sediment system. Because these fluxes appear to be concentrated into small layers, this source may be a significant input of nutrients to the estuary even if permeable, sandy layers comprise a very small proportion of the seabed.

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1. Introduction

Processes controlling nutrient levels and distributions in estuarine environments must be understood to assess the impacts of human activities and global climate change on the chemical cycling and ecology of these coastal ecosystems. Such an assessment is a fundamental first step to the development of effective management strategies.

The coupling between benthic and pelagic systems is a critical component of the nutrient cycling within estuarine systems (Nixon, 1981). In general, this coupling is thought to consist of (1) water column biological production and subsequent deposition of organic materials at the seabed, (2) remineralization of the organic materials with the release of inorganic nutrients to the pore waters and (3) transport of dissolved nutrients back into the overlying water column. Molecular diffusion processes augmented locally by irrigation by benthic organisms are generally thought to be the primary pore water solute transport mechanisms. Because the nutrients fluxing back into the water column are derived primarily from fresh particulate materials deposited at the sediment surface, this flux is closely coupled to remineralization occurring in the upper few tens of centimetres of the sediment column. Thus, there is a close, local coupling between deposition, remineralization and benthic flux.

There is growing evidence for advective input of nutrients to coastal systems from primary ground water efflux or sea water that is advectively circulated over large
distances through the bottom sediments (Moore, 1996; Simmons, 1992). Primary ground water inputs may be tied to shallow aquifers that may be recharged through atmospheric and surface water inputs. Such systems may transport materials over great distances (hundreds of kilometres) and provide a direct conduit between coastal systems and terrestrial sources (Gallagher, Dietrich, Reay, Hayes, & Simmons, 1996; McClelland & Valiela, 1998). Advective pore water transport may also be provided by sea water that is recirculated through salt marsh and other permeable coastal sediments. These recirculation processes may transport fluids over centimetre distances that occur between sea floor sand ripples to kilometre distances given the right conditions and sedimentological formations (Jahnke, Nelson, Marinelli, & Eckman, 2000; Simmons, 1992; Whiting & Childers, 1989). Such transports increase the role of sea floor processes in controlling nutrient levels in estuarine and coastal water columns. Regardless of the characteristics and controlling processes, advective transport may significantly enhance the magnitude of solute and particulate exchange between the seabed and overlying water column. Mixing of salt and freshwater within the sediment matrix may also enhance diagenetic reactions (Moore, 1999). Because reaction products can be transported large distances, advective pore water inputs are relatively decoupled from the local deposition and remineralization processes and can channel the products of diagenetic reactions through relatively narrow pathways.

Here we present in situ benthic flux chamber, pore water, sediment incubation and side-scan sonar results from the Satilla River Estuary. These results indicate the occurrence of advective pore water nutrient inputs to the estuary that may be an important component contributing to nutrient levels within the estuarine waters. Because these inputs are controlled by advective transport, they may be influenced by processes and human activities occurring at significant distances from the estuary itself.

1.1. Study location

The Satilla River Estuary is located on the Georgia coast in the southeastern US (Fig. 1). The estuary reaches through extensive salt marsh and maritime forest from the head of tide mark about 80 km upriver to St Andrews Sound, while the typical extent of salt water intrusion is just upstream from our deployment no. 10. Salinities across this region range from near zero to the west of station 10 to approximately 32 in St Andrews Sound. No significant freshwater input is provided by White Oak Creek or other dead end tidal creeks. Thus, tributaries within the Satilla Estuary are not significant point sources for nutrients to the main

Fig. 1. Study locations within the Satilla River Estuary.
channel. The surrounding areas are relatively pristine with no significant population centres or industrial complexes within the estuary. Like other coastal plain rivers, the Satilla has a relatively low flow rate, averaging 85 m$^3$/s$^{-1}$, which is lowest in February and March and maximum in October.

2. Methods

Expeditions were conducted in September 1998 and March 1999. River water samples were obtained primarily by a shipboard pumping system that also simultaneously measured temperature, conductivity, fluorescence and recorded time and location information. Once collected, individual samples were immediately filtered through 0.4 μm pore diameter polysulfone filters and stored in the dark at 4 °C. One transect from the freshwater endmember to the marine endmember was performed in September and two transects were performed in March. Transit and sampling required several hours for each transect, so a range of tidal stages were encountered during each sampling transect.

Sediment cores were recovered with two types of box corers, depending upon whether sandy or muddy sediments were encountered. At most locations, muddy sediments were recovered with a traditional, single-spade 20 x 30 cm box corer. Subcores of various diameters were removed from the box corer with intact overlying waters and processed as described later. At stations where the bottom sediments were composed primarily of sand, we used a smaller box corer fitted with a 7 cm-diameter core barrel that was equipped with a special ball-type check valve to minimize pore water washout. This corer was described in detail by Marinelli, Jahnke, Craven, Nelson, and Eckman (1998).

Pore waters were extracted from the muddy sites by centrifugation. The sediment cores were sectioned into 1 or 2 cm intervals in an N$_2$ atmosphere in a glove bag at in situ temperature and packed into 50 cm$^3$ centrifuge tubes. The tubes were centrifuged for approximately 5 min at 11,000 rpm in an Eppendorf® 5904 centrifuge. The sealed rotor was placed back in the glove bag. The centrifuge tubes were then opened, pore waters removed into plastic syringes, filtered through 0.4 μm pore diameter filters and stored in the dark at 4 °C until analysis.

At sandy sites, cores were recovered using the corer and methods described by Marinelli et al. (1998). The core barrels that were used on this corer have septae installed at centimetre intervals in the barrel walls. Once the sediment core was recovered, syringes with 18 gauge needles attached were used to penetrate the septae and remove the pore water samples. Sample withdrawal began at the shallowest depth and proceeded sequentially to the bottom of the core. Sample volume was typically \( \leq 3 \text{ ml cm}^{-1} \) depth. Downward percolation of pore waters to replace the volume withdrawn would displace pore water gradients by approximately 1.5 mm, a relatively insignificant distance considering the relatively coarse (1 cm) resolution of our pore water samples. Because samples were withdrawn sequentially from the top to the bottom of the core, this displacement due to percolation is not cumulative. Collected samples were filtered through 0.4 μm pore diameter filters and stored in the dark at 4 °C in polypropylene vials. While these methods work well for each type of sediment, they are exclusive of each other and sampling cores that contained both muds and sands in discrete layers posed a special problem. In these cases, only small amounts of pore waters were recovered and only selected analyses could be performed.

Sulphate reduction rates were determined on duplicate intact sediment cores utilizing the $^{35}$S–SO$_4^{2-}$ incubation method (Jorgensen, 1978). Intact sediment cores were injected with the radiotracer at centimetre intervals and incubated at in situ temperature. The incubation was terminated by sectioning the cores and fixing the sediments with 20% zinc acetate. In the laboratory, the reduced $^{35}$S was recovered by distillation with boiling acidic Cr$_2$O$_7$ solution (Canfield, Raiswell, Westrich, Reaves, & Berner, 1986; Fossing & Jorgensen, 1989; Zhabina & Volkov, 1978). Sulphate concentrations were determined on pore waters by the method of Tabatabi (1974).

In situ benthic flux chamber incubations were performed with an automated system as described by Jahnke and Jahnke (2000) at the locations, times and water depths listed in Table 1. This system, originally designed for deployment by manned submersible, consists of a single, 30 cm diameter PVC chamber. Mounted on the instrument frame around the chamber are two pressure cases that house the controlling electronic circuit boards and battery, an impeller pump used to stir the chamber waters, a pressurized hydraulic reservoir used to close the chamber lid, a rack containing eight spring-loaded sampling syringes and one spring-loaded injection syringe for injecting an inert tracer at the beginning of the incubation. Details of the sampler design are provided by Jahnke and Christiansen (1989). The impeller pump was connected to two nozzles which gently directed the expelled waters across the centre of the chamber and counterclockwise within the chamber. This provided gentle mixing while minimizing pressure gradients that would enhance pore water exchange. This is discussed further in the interpretation of the Br$^-$ tracer results.

The instrument was slowly lowered to the river bottom from a small boat on a line that was secured to a surface float. After a preset time interval, the controlling electronics actuated the hydraulic system, closing the chamber lid and beginning the incubation. The first
operation was to inject the tracer (0.3 M NaBr) into the chamber waters so that the chamber volume could be calculated and the chamber lid seal could be verified. Eight samples were then sequentially withdrawn at a preset time interval. At the end of the incubation period, the instrument was recovered and the samples removed for processing.

After the chamber was back aboard, the individual sample syringes were removed and the contents were immediately filtered through a 0.4 µm pore diameter polysulfone filter and stored at 4°C in the dark in polypropylene vials. In the laboratory, the chamber samples were analyzed for pH (glass electrode), NH4+ (modified from Koroleff, 1976), NO3-/NO2- (using an Alpkem RFA 300 analyser and published methods), PO43-, Si(OH)4 (Strickland & Parsons, 1972), titration alkalinity (Gran titration), Cl- (AgNO3 colourmetric titration using a LabConco digital chloridometer), and Br- (Morris & Riley, 1966). Pore waters were analyzed for all constituents named above except NO3-/NO2- and Br-. Pore waters were acidified with HCl to a pH of approximately two prior to analysis for PO43-.

The lower 45 km of the river’s course were imaged using side-scan sonar to identify sedimentary environments based on acoustic character. Profiles were collected using a Klein system 2000 digital towfish and data were stored/processed using a Triton-Elics ISIS sonar processing unit. Side-scan data were typically collected using a 150-m range scale, yielding an effective swath width of 300 m, which was adequate to image the whole channel in one or two passes. Individual lines were mosaiced using DelphMap software to produce a composite image of the riverbed.

### Table 1

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a Accuracy of locational equipment employed was less for September 1998 than for March 1999.
b Bottom water salinity and exact water depth not determined for box core collections.

### 3. Results

Grain size analyses and side-scan sonar data show a variety of sediment types and sedimentary environments along the river course (Figs. 2 and 3). Bottom surveys of sediment grain size indicate that the river bottom is very heterogeneous, with surface sediments ranging from fine-grained muds to medium sands. In the upper reaches of the river (Crow Harbor Reach and upstream), channels exhibit sandy bedforms migrating over muddy sediments or muddy sediments exposed at the riverbed (Fig. 3). Below Crow Harbor Reach, channel sediments become more sand-dominated until sand dominates in St Andrews Sound. All along the river’s course, its banks (i.e. the salt marsh/channel interface)
are characterized by broad, gently dipping ramps or by steep erosional scarp faces. In both cases, numerous subaqueous, bank-parallel lineations are commonly observed. A comprehensive description of the Satilla River sedimentary environment is outside the scope of this paper; complete details will appear in a later publication (Alexander, in preparation).

Estuarine water nutrient concentrations from both the September and March expeditions are plotted against salinity (Fig. 4). The results display the complex biogeochemical cycling of nutrients that occurs within this active estuary. Note that sampling and transit for the 20+ stations in each transect required several hours and therefore covered a range of tidal stages. Since the water locations of a particular salinity vary with the tidal cycle, nutrients are plotted against salinity, not location. In September, (Fig. 4 upper panels) NH$_4^+$ displays a dramatic removal zone in the salinity range of 5–15 with a pronounced production zone at the higher salinities, roughly 20–30. Although not as dramatic, the same features are observed in the PO$_4^{3-}$/C0$_3$ results with evidence for removal at low salinities and release at higher salinities.

A very different pattern is observed for Si(OH)$_4$ and NO$_3^-$/NO$_2^-$. Silicate displays a consistent decrease from the high concentration river endmember to low concentration sea water endmember. There is also evidence of NO$_3^-$ and NO$_2^-$ release throughout the mixing zone. Nitrate exhibits a large maximum of over 20 µM centred at a salinity of approximately 10. Nitrite exhibits a broad maximum with a highest value of approximately 5 µM at a salinity of nearly 25. This very large and broad maximum in NO$_3^-$ suggests that there is a strong source of NO$_3^-$ throughout the estuary. The oxidation of NH$_4^+$ that is already in the water column cannot be the sole source of this NO$_3^-$ because NO$_3^-$ concentrations greatly exceed the NH$_4^+$ concentrations. More likely, the NO$_3^-$ production is supported by the oxidation of NH$_4^+$ that is ultimately derived from the sediment pore waters.

March estuarine transects (Fig. 4 lower panels) display similar patterns but for some nutrients, the absolute variations are less, most likely due to reduced metabolic rates because of lower water temperatures (September: 27–29 °C, March: 15 °C). During March, PO$_4^{3-}$ and NH$_4^+$ distributions tend to be devoid of prominent features and concentrations are generally low throughout the estuary. Silicate concentrations again show a dramatic decrease from river water values of near 150 µM to less than 50 µM at a salinity of 25. Nitrate again displays non-conservative behaviour with a marked production region at salinities less than 10. However, overall NO$_3^-$ concentrations in March are significantly reduced from the warmer, September values. Note that NO$_2^-$ was not individually measured on the March samples and values represent the sum of NO$_3^-$ plus NO$_2^-$. 

In addition to the concentration trends observed within the estuarine mixing zone, it is important to note the absolute values of the concentrations as well. Ammonium and PO$_4^{3-}$ concentrations are always less than 4 and 1.5 µM, respectively. Silicate concentrations
range from approximately 150 \( \mu M \) at the river endmember to 10–20 \( \mu M \) in St Andrews Sound. Nitrate displays the greatest seasonal variation—ranging from approximately 22 to 0 \( \mu M \) in September and only 4–0 \( \mu M \) in March. It has been previously noted that PO\(_4^{3-}\) is a particularly sensitive indicator of enhanced exchange between pore waters and oxygenated overlying waters. This is because the injection of oxygenated overlying waters not only dilutes the pore water PO\(_4^{3-}\), but also tends to stimulate the production of iron oxyhydroxides which have a strong affinity for PO\(_4^{3-}\) and reduce pore water concentrations further by adsorptive removal. Phosphate for CHR2 pore waters is complex. The core was comprised of discrete sand and mud layers, and was collected as for muddy sediments and sampled by sectioning. While the layers that were dominated by sands could not be sampled, the intermixed sandy/muddy layers immediately adjacent to the sandy layers display very low PO\(_4^{3-}\) concentrations (Fig. 5a).

Ammonium and Si(OH)\(_4\) pore water concentrations for CHR are also lower in the upper few centimetres, consistent with enhanced pore water exchange. The effect is not as dramatic here as for PO\(_4^{3-}\) because the effect is primarily dilution, with little if any loss to the solid phase. In the case of Si(OH)\(_4\), the low tide, low salinity water column concentration approaches the surface pore water concentration, minimizing solute exchange.

Pore water nutrient concentrations from the muddy sediments from St Andrews Sound (SAS 1 and 3, Fig. 5a and b) also exhibit dramatic increases with sediment depth. In contrast, significantly lower pore water nutrient concentrations are measured at the sandy SAS sites (SAS 2 and 4). In addition to overall lower concentrations, the profiles from the sands are characterized by an upward curvature. Such curvature may be the result of uptake within the upper 10 cm or, more likely, enhanced transport and exchange with bottom waters. The latter could result from irrigation activities by macrobenthic organisms but because these are sandy, permeable sediments, it is more likely that bottom current-induced advective exchange is responsible (Hütte & Gust, 1992; Marinelli et al., 1998). Note that SAS4 is interlayered. It was collected as a sand core in a barrel with septae, which does not allow for sampling from muddy sediment interlayers. Below 15 cm, the core was entirely mud.

Sulphate reduction rate profiles for St Andrews Sound and Crow Harbor Reach locations are displayed in Fig. 6. In St Andrews Sound, values increase with sediment
depth from approximately 200 nmol cm$^{-3}$ d$^{-1}$ to a distinct subsurface maximum of approximately 600 nmol cm$^{-3}$ d$^{-1}$ at 7 cm. Below this depth, values decrease rapidly to very low rates at 20 cm. Pore water SO$_4^{2-}$ concentrations range from approximately 24 to 18 mM indicating that SO$_4^{2-}$ availability never limits SO$_4^{2-}$ reduction rates at this site. The lower values observed near the surface may be the result of significant oxic remineralization processes in the near surface sediments.

In general, SO$_4^{2-}$ reduction rates are lower at the Crow Harbor Reach site. Measured rates in the upper 5 cm range from approximately 50 to 150 nmol cm$^{-3}$ d$^{-1}$. Variations among the individual points are sufficient that identification of a subsurface maximum is equivocal. Below 7 cm, rates again decrease to very low levels at 20 cm. Measured pore water SO$_4^{2-}$ concentrations range from 8 to 11 mM. This range is consistent with the salinities displayed in Fig. 5.

Examples of in situ benthic flux chamber results are provided in Figs. 7 and 8. In general, the results can be divided into two groups; group 1 deployments are those in which concentrations increased dramatically, consistent with enhanced rates of solute exchange (Fig. 7). In the group 2 deployments, concentrations changed...
only a little, if at all, which is consistent with molecular diffusive solute exchange rates (Fig. 8) for the duration of these deployments.

The results in Fig. 7 display the dramatic changes in chamber water concentrations that occurred for the group 1 deployments during the relatively short 1.7 h incubation time. Ammonium concentrations increase from near zero to over 500 μM; Si(OH)₄ concentration changes from water column values of approximately 100 to over 500 μM and PO₄³⁻/Cₐ increases from near zero to between 50 and 150 μM. Coinciding with these dramatic changes were more subtle changes in NO₃⁻/Cₐ, salinity and Br⁻ tracer concentration. Nitrate is observed to decrease from approximately 12 to 4 μM, salinity increases from approximately three to eight while the excess concentration (concentration above that expected for sea water at the measured salinity) of the Br⁻ tracer decreases from around 3.5 mM to less than one. All of these results are from the Crow Harbor Reach region.

In contrast, the group 2 deployments display little or no significant concentration change in the chamber waters during the incubation (Fig. 8). In St Andrews Sound (BECI 3), consistent increases in PO₄³⁻, NH₄⁺ and Si(OH)₄ are observed during the incubation. However, instead of increasing by hundreds of micromoles, observed increases are limited to a few micromoles for PO₄³⁻ and 20–25 μmol for NH₄⁺ and Si(OH)₄. For BECI 6, which was deployed at the entrance of the main estuarine channel and BECI 8, which was deployed in Crow Harbor Reach, no significant concentration change is observed for any constituent over the 2 h incubation period.

Fig. 5. (continued)
It is important to note that the Br\textsuperscript{−} concentration changes observed in the first portion of the incubations are related to the strength of chamber mixing and the buoyancy of the spike injected. That is, while every effort was made to match the density of the spike solution to that of the surrounding bottom waters, tidal changes continually acted to shift the salinity field within the estuary, making it difficult to match density exactly at the start of the incubation. The density of the spike at BECI 3 was slightly less than the surrounding waters. This caused the spike to initially ‘float’ at the top of the chamber until it was fully mixed into the chamber waters. Since the samples are removed through a tube extending into the upper portion of the chamber, the first few samples are, therefore, elevated in Br\textsuperscript{−} concentration. At BECI 6 the densities were reasonably matched, full mixing was quickly achieved and a relatively constant concentration is observed. At BECI 8, the spike was heavier than the chamber waters and rested on the sediment surface at the bottom of the chamber until mixing finally homogenized the chamber waters.

Two important points can be drawn from the tracer results. First, mixing of the chamber waters is sufficient to homogenize the chamber waters as indicated by the nearly constant chamber concentrations measured at least by the end of each incubation. Second, a very gentle mixing process was achieved; small variations in density delayed the full mixing of the chamber waters. As displayed in Fig. 5, significant gradients in salinity and hence density, were measured in the sedimentary pore waters. Given the tracer results, it is very unlikely that the chamber stirring was vigorous enough to significantly enhance pore water exchange even at sites comprised of highly permeable, sandy sediments.

What is important about the group 2 results is that despite the unsteady initial values, there is no evidence to suggest rapid loss of the spike as was observed with the first group of results (Fig. 7). Furthermore, salinities are constant throughout the deployment, which is also in contrast to the increase in salinity observed in the group 1 results.

4. Discussion

We hypothesize that the individual results presented above can be explained by localized, advective transport of deep pore waters into the estuarine channel and that this process exerts a significant influence on estuarine nutrient distributions. We will start by focusing on the group 1 chamber results. In these incubations, chamber water nutrient concentrations increased from low river water values to $>100$ μM while the Br\textsuperscript{−} tracer decreased to less than 1/4 of the starting concentration. Some loss of the tracer is expected due to diffusion of the tracer into the sediment pore waters and dilution of tracer by replacement waters that are drawn into the chamber as each sample is removed. These processes should cause the tracer concentration to decrease by about 6% during these 2 h incubations.

Normally, a dramatic decrease in the chamber water tracer like that observed here would indicate that the
chamber lid had not sealed and that the tracer was exchanging with the surrounding waters. This explanation is not consistent with the nutrient results, however. River water PO$_4^{3-}$ and NH$_4^+$ concentrations are generally less than 2 μM. If the loss of the tracer was explained by enhanced exchange between chamber waters and river waters, nutrient increases in the chamber would be retarded by this process. To cause large positive changes in nutrient concentrations in the presence of significant leakage between the river and chamber waters would require a sea floor flux much larger than implied by the concentration increase. As we will demonstrate later, measured remineralization rates in the surface sediments cannot support the required fluxes and this possible explanation is rejected.

Another possible explanation for these results would be greatly accelerated exchange between the pore waters and chamber waters. This would increase the flux of nutrients into the chamber and dilute the tracer by exchange with the pore waters. However, the magnitude of the exchange required strongly suggests that this is also not the explanation for the observed results. Given the chamber heights (8.9 and 12.1 cm for deployment 1 and 2, respectively), to dilute the tracer by a factor of 4 would require complete mixing between pore waters and chamber waters for the upper 39–51 cm of the sediment column assuming a sediment porosity of 0.7. This homogenization would need to be completed within the 2 h deployment time. Pore water exchange this rapid is extremely unlikely.

In addition, the above explanation would require that the nutrients released to the overlying chamber waters be replaced by remineralization occurring locally, within these surface sediments. When compared to the measured sulphate reduction rates, this is unlikely. For example, NH$_4^+$ fluxes estimated for group 1 range from 0.32 to 0.87 mol m$^{-2}$ d$^{-1}$. If this flux is supported locally by the remineralization of Redfield ratio organic materials (C : N = 106 : 16), this flux would imply a remineralization rate of 2.1–5.8 mol C m$^{-2}$ d$^{-1}$. The integrated SO$_4^{2-}$ reduction for the upper 20 cm from the results displayed in Fig. 4(b) is 0.0073 mol SO$_4^{2-}$ m$^{-2}$ d$^{-1}$. Assuming a ratio of SO$_4^{2-}$ reduced to C oxidized of 1 : 2, this integrated rate suggests that 0.015 mol C m$^{-2}$ d$^{-1}$ of carbon is oxidized by SO$_4^{2-}$ reducers. This is more than two orders of magnitude less than that estimated to be consistent with the rate of NH$_4^+$ release. Although other oxidants, such as O$_2$, MnO$_2$ and FeOOH, may

![Fig. 7. Benthic flux chamber results from deployments 1 and 2 in Crow Harbor Reach from September 1998. Nitrate or nitrite graphs: O, nitrite; ⋄, nitrate.](image-url)
also contribute to the total organic matter remineralization rate in these sediments, $SO_4^{2-}$ reduction is expected to dominate and it is extremely unlikely that the occurrence of other oxidation pathways can explain these results. Note that the disparity of the results increases if the C:N ratio is greater than Redfield as might be expected if terrestrial and/or salt marsh vegetation contributes significantly to sediment remineralization. Thus, it is clear that the group 1 chamber incubation results are not consistent with leaky chambers or enhanced pore water transport supported by local diagenetic processes.

The most straightforward explanation for the observed results is the outward, advective transport of pore waters and solutes. Assuming the chambers remain well mixed throughout the incubation, an outward flow of 18–24 cm$^3$ cm$^{-2}$ h$^{-1}$ of pore water could simultaneously explain the loss of tracer and the dramatic build-up of $PO_4^{3-}$, $NH_4^+$ and $Si(OH)_4$ within the chambers. An influx of pore waters at this rate would also be consistent with the decrease in $NO_3^-$ and increase in salinity observed.

Our working hypothesis is that permeable, sandy sediment layers underlie the fine-grained, organic-rich sediments that characterize the surface of the salt marsh as schematically shown in Fig. 9. Pressure gradients develop that induce the flow of water along these permeable layers out into the river. A variety of processes could contribute to these pressure gradients. The sand layers may extend far enough into the maritime coastal lands to be recharged with rain waters or meteoric waters. Because sea salt may be entrained along this path, the escaping fluids may be essentially of sea water salinity while the forcing pressure may still be caused by fresh water inputs.

Alternatively, tidal variations across the salt marsh sediment surface may actively pump water through the sediment system. That is, at high tide, water fills all of the sediment pore spaces and organism burrows. As the water level drops with the ebbing tide, water is retained in these cavities, eventually becoming several metres above the water level at low tide in the estuary channel. This difference in water level height produces a hydraulic pressure gradient. If the sand layers extend over significant distances, even a very small migration of fluids through the relatively impermeable salt marsh surface sediments would result in a significant flow.

Bottom surveys of sediment grain size (Fig. 2) indicate that the river bottom is very heterogeneous, with surface sediments ranging from fine-grained silts and muds to coarse-grained sands. Further evidence of layering within the salt marsh deposits is provided by side-scan images of the river bottom (Fig. 3). On these images, sandy bed materials show up as lighter-coloured areas and the steep banks of the channel, cut into the salt marsh deposits, show a black character, indicative of a strong return of acoustic energy. The steep banks of the river channels exhibit white and black banding, which represents a ledge-like structure to the banks themselves and indicates either sandy–muddy interbeds within the marsh or partings between depositional units in the muddy salt marsh deposits. As portrayed on this image from Crow Harbor Reach, these subaqueous ledges are ubiquitous throughout the study area. Assuming that this layered structure extends from the river banks back into/under
the marsh, the side-scan sonar imagery provides evidence of a conduit for groundwater inflow.

This process may be extremely important in controlling the nutrient levels of the estuary because the flux out at the river bottom may represent the integrated remineralization occurring over a wide area of the salt marsh. This may explain why the estimated fluxes greatly exceed those that could be supported locally as estimated from the measured $\text{SO}_4^{2-}$ reduction rate.

Furthermore, this hypothesis may provide one explanation for why evidence for large advective flows is not observed in all of the benthic flux chamber measurements. The flow is expected to be restricted to relatively small areas where permeable, sandy layers outcrop into the river channel. The thickness of these layers is unknown but they may only be a few centimetres to a few tens of centimetres thick. It would be very difficult to routinely deploy a chamber directly on such a layer.

Without knowing the distribution of the advective layers within the river channel, it is not possible to directly estimate the importance of this process in controlling nutrient concentrations in the estuarine waters. To set rough limits, we have developed a simple steady-state advection–diffusion–reaction model of the estuary. The model does not include the effects of stratification or inundation of the salt marsh at high tide and is presented for illustrative purposes only. The average discharge of the river is reported to be $85 \text{ m}^3\text{s}^{-1}$. At an approximate average depth of 10 m and channel width of 600 m (at least for the critical Crow Harbor Reach region), this transport corresponds to a flow velocity of roughly $50 \text{ m} \text{h}^{-1}$. Using this advection velocity, the best fit of the salinity field was achieved with an eddy diffusivity of $5 \times 10^2 \text{ m}^2\text{h}^{-1}$.

The mean fixed N flux for the group 1 deployments is $0.5 \text{ mol m}^{-2}\text{d}^{-1}$. If we assume that advective inputs are restricted to the region between 0 and 12 salinity and that there are no other inputs or losses of fixed N, the resulting distribution can be estimated as a function of seabed area that exhibits the measured fluxes (Fig. 10). The results demonstrate the potential importance of advective nutrient inputs. If only 0.05% ($500 \text{ m}^2\text{km}^{-2}$) of the estuary sea floor exhibits fluxes of the magnitude measured in the group 1 stations, a broad nitrate maximum of the magnitude displayed in Fig. 2(a) is calculated. This calculation demonstrates that this process has the potential to significantly control nutrient

**Fig. 9.** Schematic of hydrostatic advective pore water transport model.

**Fig. 10.** Numerical simulation of nitrate vs. salinity in the Satilla Estuary. In addition to advective and diffusive processes that would conservatively mix the riverine and sea water endmembers, fixed nitrogen is supplied from the seabed in the 0–12 salinity region at the rate measured for the group 1 stations. Different model runs demonstrate the sensitivity of the simulation to variations in the proportion of the seabed exhibiting these fluxes. Legend: ○, 0.10%; ●, 0.05%; △, 0.01%.
concentrations in this estuary. Additionally, because the flux may be derived from such a small portion of the total seabed, this calculation qualitatively supports our conclusion that subsequent deployments may have simply missed the locations of advective transport.

In summary, the combination of benthic flux measurements, pore water nutrient profiles, water column transect nutrient distributions, sediment grain size distributions and side-scan sonar observations support the premise that advective transport of pore waters through permeable sand layers may be an important source of nutrients to the Satilla River Estuary. Locally occurring benthic microbial mineralization measured via $S^{35}$–$SO_4^{-2}$ reduction rate incubations is insufficient to generate the nutrient fluxes observed in 20% of the deployments reported here. A simple steady-state advection–diffusion model based on the measured benthic ammonium fluxes and water column salinity distributions suggests that the observed water column $NO_3^{-}$ concentrations could be maintained if only 0.05% of the estuary seabed consists of permeable sands which exhibit advective exchanges of the magnitude observed for the group 1 deployments.

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