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Estimating the Mean Ocean-Bay Exchange Rate of the Chesapeake Bay

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[1]   A model of the salt balance in the Chesapeake Bay is discussed, which takes into account only time-dependent riverine input and mean ocean-bay exchange. Estimates of (spatial) mean bay salinity are made using two different data sources: a 16 year record of surveys taken by the Environmental Protection Agency's Chesapeake Bay Program, and a 10 year record of hydrographic sections taken in the lower bay by the Center for Coastal Physical Oceanography at Old Dominion University. Using United States Geological Survey river flow data to force the model, both data sets are consistent with this simple model and both imply a mean oceanic exchange (*e*-folding) timescale of 90 days, which is equivalent to an effective exchange rate of roughly 8×10^3 m^3 s^{-1}.   *INDEX TERMS:* 4235 Oceanography: General: Estuarine processes; 4223 Oceanography: General: Descriptive and regional oceanography; 4845 Oceanography: Biological and Chemical: Nutrients and nutrient cycling; *KEYWORDS:* estuarine exchange, Chesapeake, long-term observations, hydrography

1.   Introduction

[2]   Estuaries represent one of the major interfaces between terrestrial and oceanic systems. The Chesapeake Bay has a very large watershed, draining several different river basins. The bay has a wide mouth, 18 km across at its narrowest point, so that subinertial circulation in the bay is strongly rotational [*Valle-Levinson et al.*, 1996], and hence exchange processes between the bay and the ocean are quite complicated. Oceanic water tends to enter the bay in the northern portion of the bay mouth and bay water leaves through the southern portion. It is difficult to make even an instantaneous estimate of the oceanic flux into the bay, due to the strong tidal and subinertial circulation at the bay mouth, and to large spatial inhomogeneities in the flow [*Valle-Levinson and Lwiza*, 1997; *Whitford*, 1999]. High spatial resolution measurements over a tidal cycle, such as *Valle-Levinson et al.* [1995] produce instantaneous estimates of the exchange, but are not useful for determining the mean oceanic exchange, since the subinertial variations can be quite strong [*Goodrich*, 1988].

[3]   Recent work [*Wong and Valle-Levinson*, 2002] at the mouth of the Chesapeake Bay has resulted in long time series of direct measurements of the currents at five locations, which span roughly 75% of the bay mouth. Over a the course of two ~75 day deployments, they observed a mean net outflow of 3×10^3 m^3 s^{-1} and a mean net inflow of 2×10^3 m^3 s^{-1}. This results in a net exchange a factor of four smaller than the value which will be estimated in this paper. The discrepancy is discussed in the analysis section.

[4]   Using multiyear time series of bay salinities allows a different, indirect approach to determining the effective exchange rate. For instance, *Pritchard* [1960] used 5 years of salinity data in Chincoteague Bay to estimate exchange rates there. This paper uses a similar approach. *Gibson and Najjar* [2000] used a recursive model to describe variation in the Chesapeake Bay using the EPA data and USGS freshwater flux data used in this paper. Their approach provides a good empirical predictive ability for salinity in different portions of the bay.

[5]   *Pritchard* [1960] used a simple box model to estimate the exchange rate of the ocean with Chincoteague Bay, a small coastal bay which straddles the Maryland/Virginia border. He derived an expression that relates the change in mean salinity to the river influx, bay salinity, and oceanic salinity. He then used mean values of these to estimate the mean exchange rate. However, this method is sensitive to the values chosen for the mean bay and oceanic salinities, and it is not clear how to estimate appropriate values for these. In this analysis, we still compute mean bay salinities, but avoid the issue of the sensitivity of the result to how the mean is estimated by studying not the difference between the oceanic and bay salinities, but by examining the temporal variation in a large set of estimates of mean bay salinity. This timescale is determined by comparing salinity observations with results from a simple box model, and optimizing the model by varying the exchange rate. The determination of this timescale is the focus and major contribution of this paper; no attempt is made to explain the dynamics of the exchange process. However, the technique outlined here could be used in other estuaries to develop a better understanding of what processes set exchange rates.

[6] This analysis is possible because of the existence of two long-term monitoring efforts. The Environmental Protection Agency's (EPA) Chesapeake Bay Program (CBP) has been making extensive surveys of the entire bay and its tributaries for the last 16 years, which can be used to produce an estimate of mean bay salinity at biweekly to monthly intervals. Also, data from a 10 year effort by Old Dominion University's (ODU) Center for Coastal Physical Oceanography (CCPO) are used for estimates of lower-bay mean salinities. Because of the levels of biological activity found in estuaries and the fouling it tends to lead to, few long moored time series of salinity exist in estuaries, making shipboard survey programs like these the best way to collect reliable long term salinity time series. In addition, continuous records of daily freshwater flux into the bay through its many tributaries are available from the United States Geological Survey (USGS).

[7] This rest of this note is as follows: section 2 will detail the sources of data, section 3 discusses the model that will be applied, and section 4 will provide an analysis of the data in the framework of the model. A discussion and summary follow.

2. Data

2.1. CCPO Bay Mouth Hydrographic Sections

[8] Numerous volunteers from CCPO have participated in an ongoing program over the past 10 years to collect detailed hydrographic information in the lower bay ([*Valle-Levinson et al.*, 1995]; data are also available online at http://www.ccpo.odu.edu/~jay/home.html). Once a month, at spring high tide, a survey is made across a section about 20 km inside the bay mouth (Figure 1). The section is sufficiently far inside the bay mouth that it appears to respond to freshwater events on roughly the same timescale and at the same magnitude as does the rest of the bay. The hydrographic section consists of 20 CTD stations along the survey line, which is about 40 km long. This survey includes casts in the middle of three channels which lie parallel to the local bay axis. As of December 2000, 104 surveys have been made. Mean salinities were estimated from each of the cruises using a simple horizontal and vertical average (Figure 2a). This time series will be referred to as S_{CCPO}.

2.2. EPA Data

[9] The Chesapeake Bay Program of the EPA has been making water quality measurements of the bay since 1984, and all of the data are publicly available (EPA, Chesapeake Bay Program Data Retrieval: EPA data available on the World-Wide Web, accessed April 10, 2001, at URL http://www.chesapeakebay.net/data/index.htm). It consists of salinity, temperature, and many biological and chemical measurements at various depths. There are 39 stations that are sampled on a regular basis (Figure 1) as part of the "main stem" survey. Only cruises which consisted of at least 30 casts and took less than five days to complete were considered in this study, ensuring reasonably synoptic and bay-wide coverage. Over the 16 year period, 207 cruises meet these criteria (on average, every 28 days). To estimate the average salinity, a simple (vertical and horizontal) mean of these data are taken (Figure 2a)-no attempt to "weight"
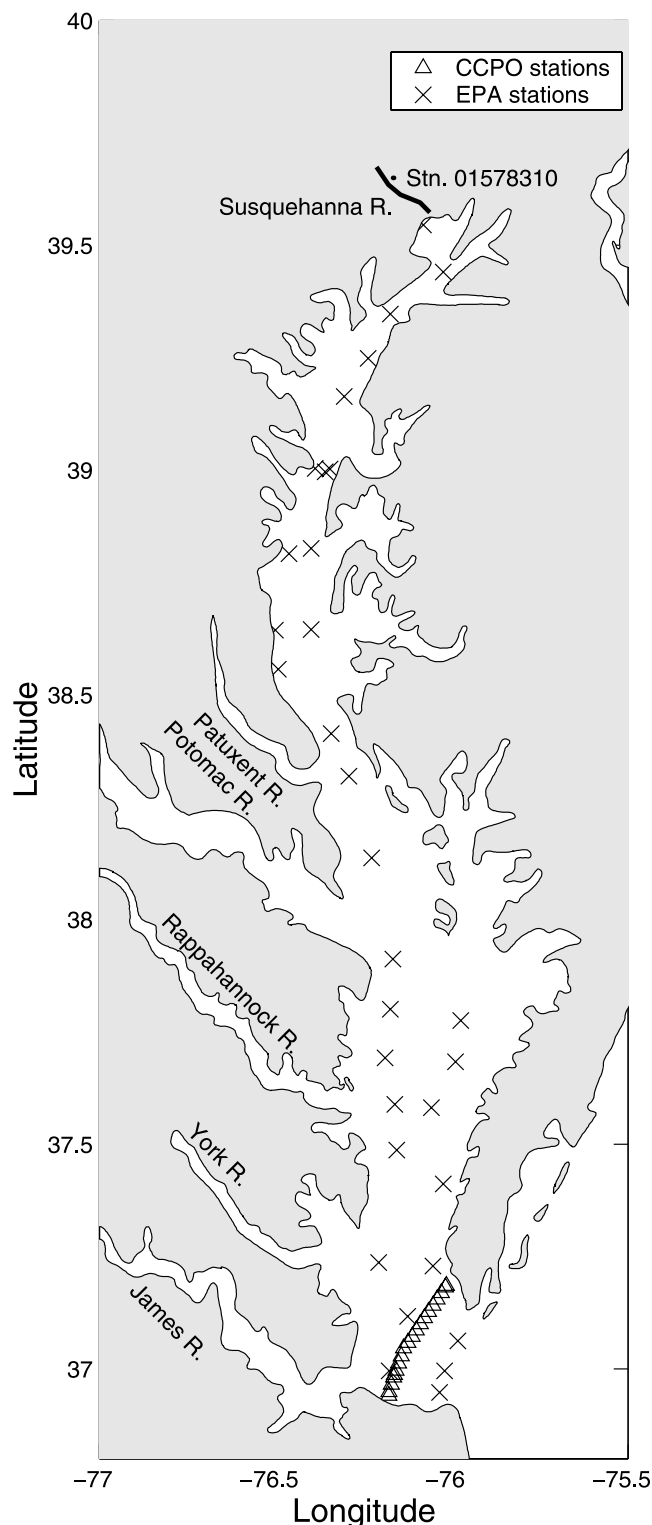

Figure 1. The Chesapeake Bay. The locations of the EPA sample stations are marked with crosses; the CCPO survey is marked with triangles. The approximate location of the USGS river gage is noted.

stations is made. As the stations are reasonably evenly distributed, this appears to be a justifiable simplification. This time series will be referred to as S_{EPA}. The EPA data and the CCPO data, although representing different spatial
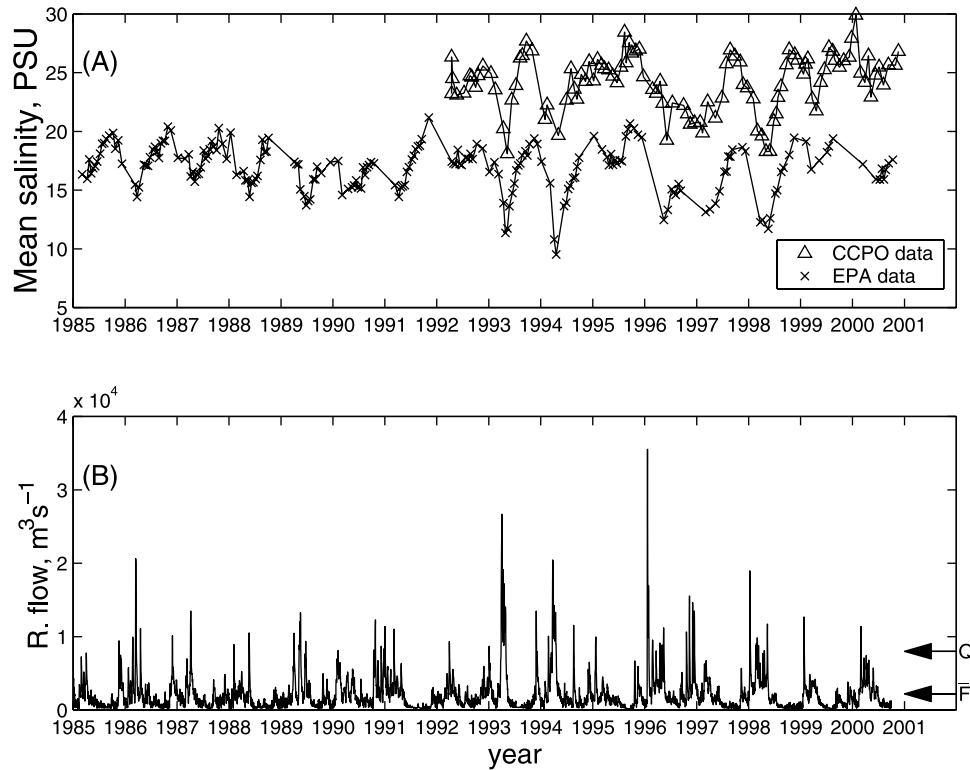
Figure 2. (a) Time series of the mean salinity from the CCPO and EPA data sets. (b) Freshwater flux from the Susquehanna River. The (derived) value of oceanic input (Q) and the mean freshwater input ($\bar{F}$) are indicated with arrows.

domains, display essentially the same temporal variation. The mean CCPO salinities are higher since they are representative of only the lower bay.

2.3. Freshwater Flux

[10] Freshwater flux was determined using USGS stream gage data from the Susquehanna (station 01578310, Figure 1). and the Potomac (station 01646500). The Susquehanna and Potomac are responsible for roughly 48% and 16% of the freshwater flux into the Chesapeake Bay, respectively. [*Schubel and Pritchard*, 1986]. The total flux of freshwater into the bay was estimated by multiplying the sum of the Susquehanna and Potomac fluxes by 1.54, so that it is representative of the total flow into the bay (Figure 2b).

3. Model

[11] The mean salinity of the bay is a product of two competing processes: freshwater flux into the bay through rivers, which freshens the bay, and exchange with adjacent oceanic waters, which raises the salinity. The relationship between freshwater flux and the mean salinity is not clear, at first glance (Figures 2a and 2b). The goal here is to develop a model of the time-dependent "climatic" (spatial) mean salinity.

[12] The box model consists of a freshwater flux, $F(t)$ into the bay, an oceanic flux Q into the bay, and, assuming the volume of the bay does not change over sufficiently long time periods, a corresponding flux $-Q - F(t)$ out of the bay. Q is assumed to be constant; implications of this will be considered in the discussion. The freshwater flux into the

bay has salinity 0, the oceanic flux into the bay is assumed to have salinity S_a, and the flux out of the bay is assumed to be of mean bay salinity $S(t)$.

[13] Combining these salinity fluxes yields a linear, first order ordinary differential equation for the salinity balance:

$$\frac{dS}{dt} = -\frac{F}{V}S + \frac{Q}{V}(S_a - S),$$ (1)

where $S(t)$ is the modeled mean salinity and V the total volume of the bay (75 km^3). Some easy checks demonstrate the behavior of this equation. Setting $F = 0$ (no freshwater input) yields a solution where S approaches S_a on an (*e*-folding) timescale of $\frac{V}{Q}$. Likewise, setting $Q = 0$ (no exchange with the adjacent ocean) yields a solution where S approaches zero on a timescale of $\frac{V}{F}$, the freshwater flushing timescale of the bay. This is a special case of *Pritchard*'s [1960] formulation, in which the salinity of the water leaving the bay is made up of a fraction n of bay salinity water and a fraction $1 - n$ of oceanic salinity water. In this case, the water leaving the bay is assumed to have salinity characteristic of that in the bay ($n = 1$).

[14] Equation (1) will be solved numerically using values of $F(t)$ from the USGS data, but it is instructive to analytically derive an approximation to the solution. Assuming $S(0) = S_a$, equation 1 has an analytic solution:

$$S(t) = S_a\left(1 - \int_0^t \frac{F(t')}{V}e^{\int_t^{t'}\frac{Q+F(t'')}{V}dt''}dt'\right).$$ (2)
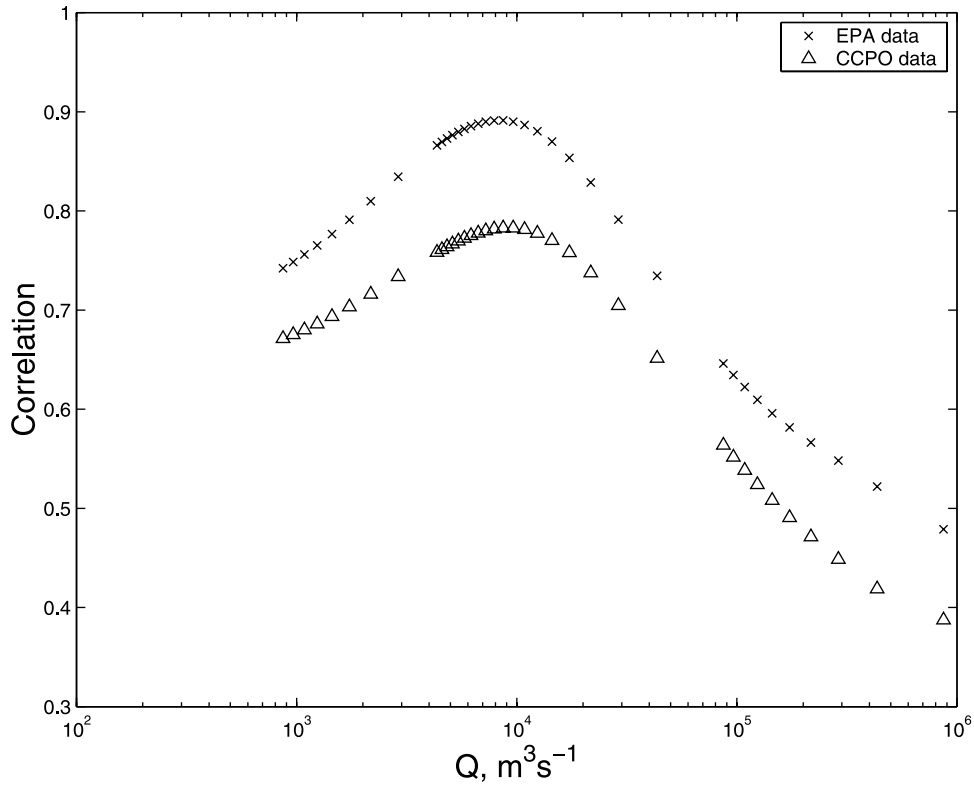
Figure 3. The correlation between the measured mean salinity time series and the modeled salinity, as a function of the exchange flux Q.

This equation, as it stands, is difficult to interpret, due to the complex form of the integral. In section 4, it will be shown that the value $\frac{F(t)}{V}$ is significantly smaller than $\frac{Q}{V}$, so that the time-varying nature of F is not an important consideration in the exponent. Without significantly changing the nature of the solution, $F(t)$ can be replaced (in the exponent) with $\bar{F}$, where $\bar{F} = 2.2 \times 10^3$ m^3 s^{-1} is the mean value of F ($\frac{V}{\bar{F}} = 397$ days). The time dependence of F will be retained in the first integral. This yields a solution of the form

$$S(t) = S_a\left(1 - \int_0^t \frac{F(t')}{V} e^{(t'-t)\frac{Q+\bar{F}}{V}}dt'\right). \quad (3)$$

This solution is considerably easier to interpret. It shows that the mean bay salinity is a response to recent freshwater fluxes into the bay, and the influence of any given freshwater event decreases over the timescale $V(Q + \bar{F})^{-1}$. It remains to estimate the mean exchange flux Q.

[15] The primary shortcoming of this model is the interpretation of the input salinity S_a. In the case of the EPA data, the mean bay salinity approaches a maximum of about 21 psu during weak flow periods, and the CCPO lower bay data approaches 30 psu, both much fresher than adjacent oceanic waters, which tend to be approximately 32 psu just to the north of the bay mouth. However, considering the form of the solutions (2) and (3), the correlation between the model solution and the salinity observations is independent of the value chosen for S_a. Once an optimal Q is chosen, that value can be used to estimate a value of S_a that best fits the observations. This value is more representative of the salinity that the mean

approaches under low freshwater flow conditions than of the open ocean salinity. Therefore, the values for S_a will be different for the EPA and CCPO data.

4. Analysis

[16] To estimate Q, Equation 1 is numerically integrated for a range of values of Q, resulting in a set of potential solutions referred to here as S_Q. These solutions are then compared to S_{EPA} and S_{CCPO}. The optimal Q is chosen when the correlation between S_Q and S_{EPA} or S_{CCPO} are maximum (Figure 3). For the EPA data, this value is $Q = 8 \times 10^3$ m^3 s^{-1} (maximum correlation of 0.9), and for the CCPO data, it is $Q = 9 \times 10^3$ m^3 s^{-1} (maximum correlation of 0.8). These correspond to exchange time scales $\left(\frac{V}{Q}\right)$ of 90 and 80 days, respectively. These values of Q are consistent with observations made by *Valle-Levinson et al.* [1995, 1996] and *Goodrich* [1988], though these direct estimates were made using relatively short time series.

[17] Now that Q has been estimated, the solution of (1) and the salinity time series can be used to fit S_a for each data set. Since the CCPO data are representative of the lower bay, its respective S_a value is much higher ($S_a^{CCPO} = 30.8$) than that for the EPA data ($S_a^{EPA} = 21.6$), which is more representative of the whole bay. The fit of the model data onto the observations is made with only one free parameter, S_a; it is therefore encouraging that not only the maximum salinity is matched but the magnitude of the salinity variations (Figures 4a–4d). This suggests that the simple conceptual model is consistent with the bulk salinity processes within the bay.
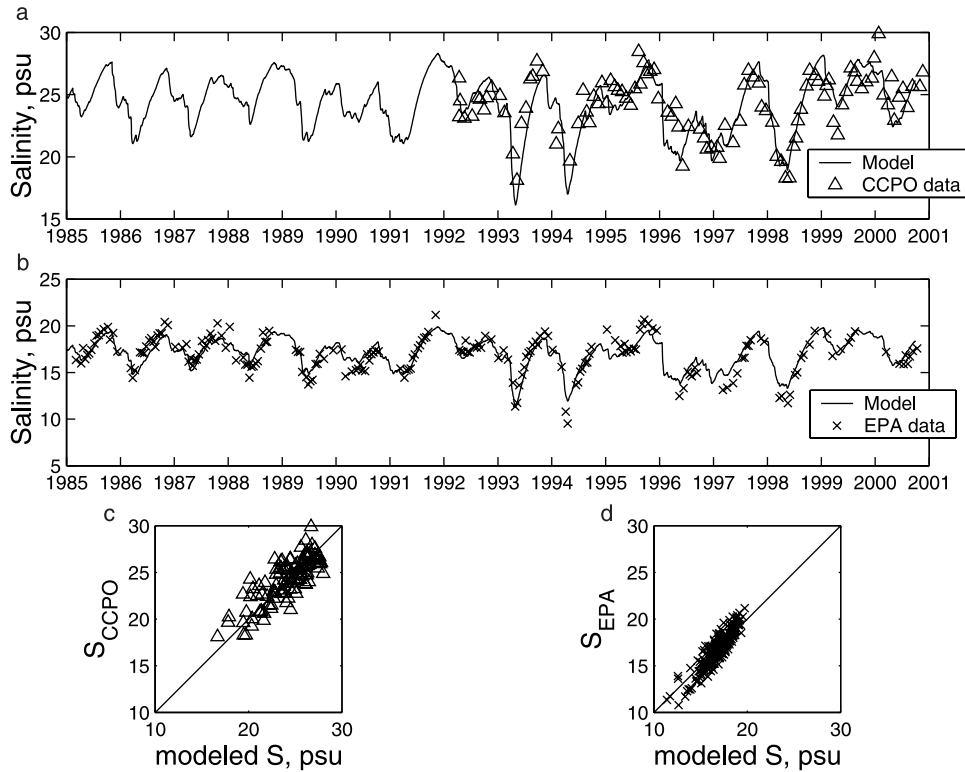
Figure 4. Comparison of modeled and observed mean salinities: (a) CCPO mean data and modeled data with $Q = 8 \times 10^3$ m^3 s^{-1} and $S_a = 30.0$, (b) EPA mean data and modeled data with $Q = 8 \times 10^3$ m^3 s^{-1} and $S_a = 20.9$, (c) point-by point comparison, modeled and CCPO, and (d) point-by-point comparison, modeled and EPA.

[18] Similarly, the steady state portion of (1) can be solved using the mean freshwater flux F, the estimated exchange rate Q, and the mean EPA and CCPO salinities to solve for the effective ambient salinity S_a. This method yields the same results as does the least squares fit discussed earlier.

[19] The uncertainty in the effective mean exchange rate Q cannot be objectively estimated given the method used to determine Q. However, an estimate of its uncertainty can be made by relating it to the standard deviation of the error in the predicted salinity. By defining $\Delta Q \approx \frac{\partial Q}{\partial S} \Delta S$, where ΔS is the standard deviation of the difference between the predicted and observed salinities. To estimate $\frac{\partial Q}{\partial S}$, we use the time-averaged version of equation (1). This yields

$$\Delta Q = \frac{\bar{F} S_a}{(S_a - \bar{S})^2} \Delta S. \qquad (4)$$

Given the estimates of S_a and the mean salinity $\bar{S}$ for each of the datasets, the mean freshwater flux $\bar{F}$, as well as the standard errors $\langle S_{model} - S_{observed} \rangle$, $\Delta Q = 2.2 \times 10^3$ m^3 s^{-1} for the EPA data and $\Delta Q = 2.6 \times 10^3$ m^3 s^{-1} for the CCPO data. This represents the uncertainty in the determination of Q, and is not related to the natural temporal variance in Q.

[20] The apparent discrepancy with the work of *Wong and Valle-Levinson* [2002] is interesting: using direct measurements, they estimated a mean net exchange on the order of 2×10^3 m^3 s^{-1}. The current work in essence uses the time rate of change of the total salinity content of the bay to estimate the exchange rate, and finds that the

effective mean exchange is roughly four times greater. Therefore, such a small mean exchange cannot, by itself, account for the rate of change of salinity in the bay. *Wong and Valle-Levinson* [2002] also observed that the variance of the net exchange was very large compared to the mean exchange. It is likely that there is a fluctuating term that actually dominates the mean exchange; i.e. if the salinity difference between the water entering the bay and the water leaving the bay is ΔS, then the rate of change of salinity content should scale as:

$$\frac{d}{dt} \int S \, dV \approx \bar{Q} \overline{\Delta S} + Q' \Delta S', \qquad (5)$$

where the barred values are time means and the primed values are fluctuations around those means. Observations at the bay mouth suggest that $\overline{\Delta S} \approx 3$ psu. While [*Wong and Valle-Levinson*, 2002] measured the $\bar{Q}$ and Q' terms, salinity time series were not recovered due to the rough conditions at the mouth of the Chesapeake Bay, which resulted in a low data return. For the fluctuating term to be important, the transport and salinity fluctuations would have to be strongly correlated. The current paper does not distinguish between the time fluctuating term and the mean term in its formulation, and as such actually infers the residence time V/Q due to oceanic exchange. The effective exchange at the mouth is likely a result of both he mean and the fluctuating exchange terms; the timescale estimate is valid because the timescale of fluctuations by *Wong and Valle-Levinson*'s [2002] estimates of $Q(t)$ are short compared to the effective
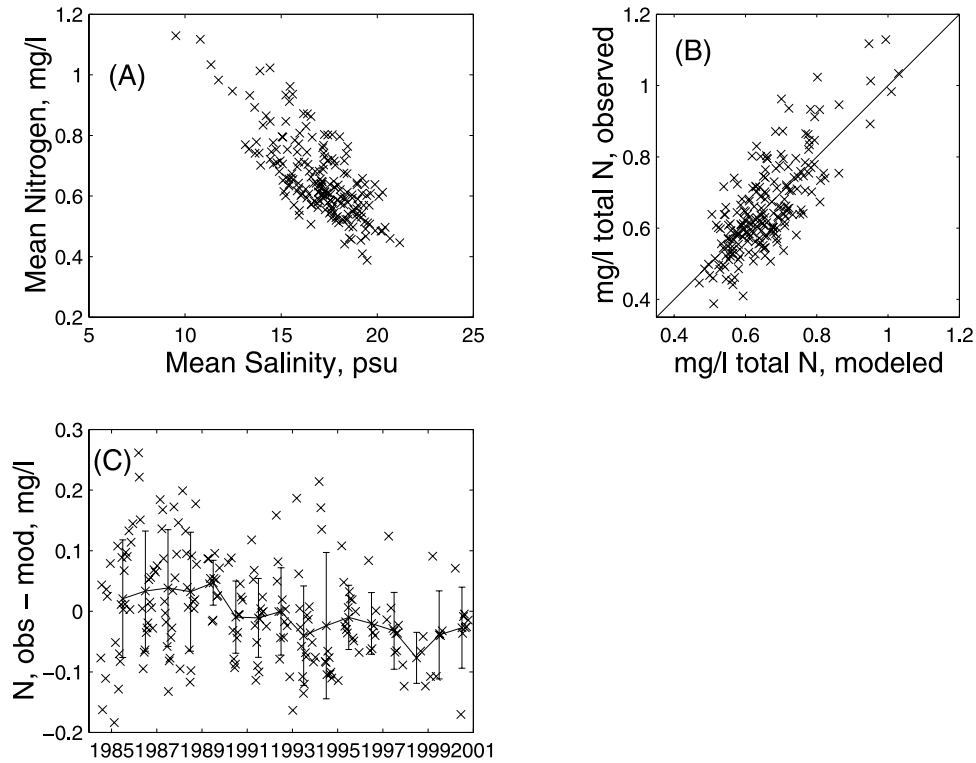
Figure 5. (a) Bay mean nitrogen values as a function of bay mean salinity values, 1985–2000. (b) Observed mean Nitrogen values as a function of the modeled nitrogen concentration using equation (6). (c) The difference, observed – modeled, for the nitrogen concentrations as a function of time. Bars represent yearly means and standard deviations.

exchange timescale, and are essentially "averaged over" by the bay response.

5. Discussion
5.1. Nitrogen Levels

[21] The freshwater flux into the bay is known to significantly influence nutrient levels, which play a major part in determining the rhythms of the bay's ecosystem. Nitro-

gen levels, for instance, are known to be higher during periods of heavy freshwater influence, and are hence negatively correlated with salinity. Levels of mean total nitrogen concentration (total dissolved plus particulate organic [*U.S. EPA CBP*, 1993]) are indeed strongly correlated with the modeled salinity (Figure 5a). Long-term change of nitrogen levels in the bay is of particular concern to regulatory agencies and has been a major focus of Chesapeake Bay remediation programs. Determining these
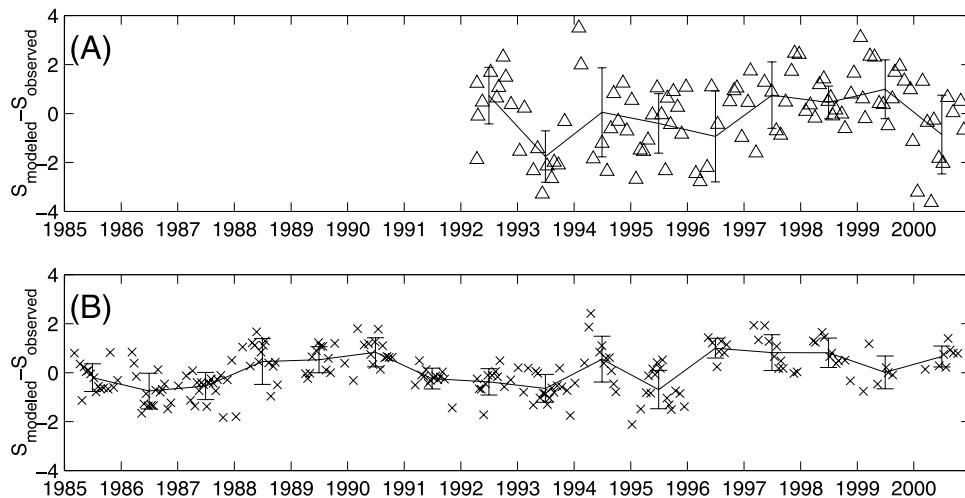

Figure 6. Model – observation difference on seasonal and interannual scales for both the EPA and CCPO data sets. (a) CCPO, interannual and (b) EPA, interannual.

trends and trends in biota is an important part of gauging the health of the bay [*Dauer and Alden*, 1995]. In mean terms, the nutrient budgets of the bay have been carefully quantified [*Boynton et al.*, 1995]; here we study what determines the magnitude of temporal variations in total nitrogen levels. It is difficult to determine what portion of the fluctuations in nitrogen levels are due to seasonal and interannual cycles in river flux and what is due to long-term change. With the framework developed here, and the estimation of the exchange rate Q, a box model similar to that used for salinity can be constructed, and the influence of the time-varying riverine flux determined and eliminated.

[22] Assuming that nitrogen levels in the bay are affected only by river fluxes and oceanic exchange (likely a poor assumption as surface exchange, burial processes, and denitrification can play a significant role in determining nitrogen levels [*Boynton et al.*, 1995]), the spatial mean total nitrogen may be estimated by

$$\frac{dN}{dt} = \frac{F(t)}{V}(N_R - N(t)) - \frac{Q}{V}(N(t) - N_O), \qquad (6)$$

where N_R is the nitrogen content of the river water entering the bay, N_O is the nitrogen content of the incoming oceanic water, and $N(t)$ is the spatial mean nitrogen in the bay at time t. Q is taken to be 8000 m^3 s^{-1}. The model is numerically integrated using the same freshwater forcing as before. N_R and N_O are determined by regressing the solution to (6) with the measured nitrogen, resulting in $N_R = 1.8$ mg l^{-1} and $N_O = 0.32$ mg l^{-1}, consistent with values found near the head and mouth of the bay, respectively. The model and the observed nitrogen levels are correlated at a level of 0.75 (Figure 5b).

[23] In order to determine if there is a secular, long-term trend in total nitrogen levels in the bay, the model results are subtracted from the measured mean bay nitrogen levels. The difference between the observed mean nitrogen levels and the modeled nitrogen decrease slightly but statistically significantly over the 16 year period of the observations (Figure 5c). A linear regression of the yearly mean differences has a slope of 6 μg l^{-1} yr^{-1}, and correlation of -0.83. Therefore $N(t)$ appears to be decreasing on a multiyear timescale, albeit slowly. That the model works as well as it does suggests that these ideas might be used to better understand the timescales of loading and flushing of total nitrogen in the bay.

5.2. Seasonal and Interannual Variation

[24] Little is known about mean exchange rates of ocean estuary systems; even less is known about the time dependent nature of the exchange on seasonal or interannual scales. The indirect method outlined in this paper is not of particular use for discerning this variability on short timescales; i.e. timescale shorter than the exchange timescale. In addition, the total freshwater flux, which has been shown to dominate the variance of salinity in the bay, has a very strong annual signal with maximum flux in April (4500 m^3 s^{-1}) and minimum flux in September (750 m^3 s^{-1}). Therefore, determining seasonal cycles in Q or S_a (which surely exist) would be exceedingly difficult given the character of the data. However, some insight may be gained into longer timescale variation by examining the difference between the

modeled results (Figure 4) and the observed salinities over the entire temporal span of both data sets.

[25] It is difficult to draw any strong conclusions regarding interannual variability from the model−data comparisons (Figures 6a and 6b), due to the relatively small interannual signal. It appears that in both time series, the model underestimates the salinity in the early 1990s and overestimates it in the late 1990s. If this is due to variations in Q, and the model is overestimating the salinity, this implies that the model is, in some sense, "overexchanging" oceanic and bay waters, suggesting that the exchange rate used in the model is too high in these times. Likewise, the model underestimating the salinity implies that the exchange rate is too weak. This means that the actual exchange rate may have been slightly stronger in the early 1990s and slightly weaker in the late 1990s (and the late 1980s, from the EPA data alone). If it is due to variations in S_a, then the actual S_a was higher in the early 1990s and lower in the late 1990s. Unfortunately, no data exists to distinguish between these hypotheses. The interannual variation does not appear to be tied to the annual mean freshwater flux. In short, determining variation in the effective exchange rate Q or ambient oceanic salinity S_a with any confidence is likely beyond the capability of these data sets.

6. Summary

[26] The mean effective exchange rate between the Chesapeake Bay and the adjacent shelf waters is estimated to be approximately $8 \pm 2.6 \times 10^3$ m^3 s^{-1}, equivalent to an exchange (*e*-folding) timescale of about 90 days. This value is consistent between two independent data sets. Application of a similar model using the estimated exchange rate to values of total nitrogen in the bay reveal a small but significant decline in nitrogen levels over the period 1984−2001. Variation of the model-observation difference on interannual timescales exists but its origin is not clear. It would be interesting to apply this technique to estimate relative flushing rates and timescales for other estuaries; this depends on the existence of other sufficiently long (i.e. many times $\frac{V}{Q}$) time series of high-quality, well-distributed salinity measurements, which are rare. It remains to better understand the dynamic mechanism which provides for the exchange.

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