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A NUMERICAL STUDY OF EXCHANGE AND MIXING OVER A SILL AT THE MOUTH OF THE SAINT JOHN RIVER ESTUARY

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ABSTRACT: The Saint John River discharges into the Bay of Fundy at the city of Saint John, New Brunswick. Although the Bay of Fundy at that location has tidal ranges of up to 8m, tides in the Saint John River Estuary are damped by the Reversing Falls and have a range of approximately 50cm. The tidal effects on the Saint John River are nevertheless significant. The mixing and exchange of salt and fresh water between the main Saint John River, the Kennebecasis Fjord and the exit to the open sea is strongly controlled by the presence of a shallow sill. The area is one of extreme spatial variability in salinity and temperature and acts as a laboratory for testing sound speed interpolation algorithms.

The balance between the fresh water discharge of the Saint John River and the upstream intrusion of salt waters from the Bay of Fundy is examined in a numerical model of the estuary. QUODDY, a three-dimensional finite element baroclinic shelf circulation model, is used. The model has upstream and downstream open boundaries. These boundaries are forced with observations made over a tidal cycle in July 2001. Additional current and CTD data within the bounds of the model were obtained in June 2003. The observations are used to examine the performance of the model in various regions of the estuary. The gross circulation is adequately reproduced along the main Saint John River stem. However, due to the extremely thin and localized nature of the exchange, some detail is lost as a result of compromises in array size and computational speed. Problems were encountered in modelling the flow from the Saint John River Estuary into the Kennebecasis Fjord. The combination of particularly high density gradients together with strong topography found in this area pose challenges in the implementation of a sigma-coordinate model.

1. INTRODUCTION

The exchange and mixing of salt and fresh water in an estuary are dependent on a multitude of factors including discharge, tidal range, and morphologic restrictions (Dyer 1997). The Saint John River discharge is large (2.83×10^2 to 31.1×10^2 m^3/s), and with a pronounced seasonal variability due to a spring freshet (Trites 1960). The ultimate receiving basin is the Bay of Fundy where local tides range in excess of 7m and are dominated by the M2 tidal period (12.42 hours). Because of the Reversing Falls Sill, however, it has long been known that the inflow is restricted, resulting in reduced tidal amplitudes of 0.5 to 0.7 m immediately upstream. The intense mixing of that sill results in injection of blended salt-fresh water at ~23 ppt (Trites 1960). The fate of that injected mixed water, however, is then controlled by a second sill in Grand Bay which governs the advection of the mixed water both up the main river axis and laterally into the Kennebecasis Fjord. Exchange over that second sill (herein termed the Boars Head Sill), is the focus of this paper.

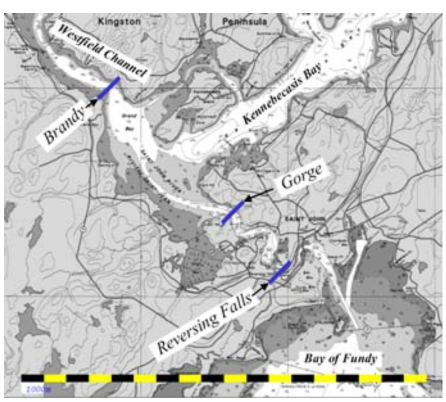
The presence of the Boars Head Sill has been inferred to restrict the replacement of the bottom waters of the Kennebecasis Fjord (Trites 1960). Based on existing knowledge of the gross sill depths the exchange was assumed to be limited to spring tides and low river discharge periods. Geomorphic observations from new multibeam sonar mapping (Hughes Clarke and Haigh, this volume), have allowed us, for the first time, to see the detailed scour evidence of this exchange. As part of this new mapping, new ADCP and CTD profiling over tidal cycles have provided spatially and temporally dense oceanographic observations of the exchange mechanism during a spring tide, low-discharge period. This paper tries to reproduce those observations using a three-dimensional baroclinic hydrodynamic model of the area in order to better understand the physical mechanisms and ultimately to be able to predict the change in the exchange character as external forcing parameters, such as river discharge, water temperature, and tidal range are altered (as occurs at other times of the year).

2. OBSERVATIONS

2.1 July 2001 Survey – Upstream and Downstream Boundary Conditions

Our model has two open boundaries (Figure 1). The upstream boundary is located at Brandy Point (refered to as Brandy) where the river is roughly 900m wide. The downstream boundary is located at the beginning of the gorge which leads to the Reversing Falls (called Gorge). Here the river has begun to constrict and is approximately 350m wide. On 21 July 2001(Gorge) and 23 July 2001(Brandy) the Ocean Mapping Group (OMG) acquired data at the two boundaries. Observations were taken over a tidal cycle (12.42 hours). At each boundary a series of vertical profiles of velocity, temperature and salinity were measured. Velocity was measured across the boundary whereas temperature and salinity were measured at the center of the channel only. Elevation was measured continuously throughout the survey.

The processed elevation data is shown in Figure 1. The M2 residual elevation relative to the geoid at Brandy is 4.8cm higher than the M2 residual elevation at Gorge. This implies a net downstream flow from Brandy to Gorge. The M2 amplitude and phase are 23.8cm and 284.6° at Brandy and 21.2cm and 268.3° at Gorge. The M2 tide signal at Brandy lags that at Gorge by 34 minutes but the tides are not symmetric. Figure 1 shows that high tide at Brandy lags high tide at Gorge by only 8 minutes whereas low tide at Brandy lags low tide at Gorge by 51 minutes. This implies a steepening of the flood rise (increased M4 component) probably resulting from the propagation of the tidal wave across the shallow sill.



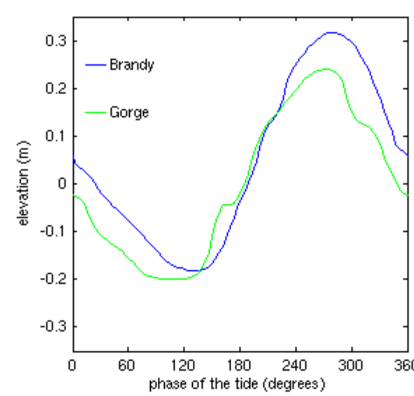


Figure 1. Overview of Saint John River Estuary (left) and measured elevations at the boundaries (right).

The velocity profiles at the open boundaries were measured using an ADCP. The combination of the mount depth of the ADCP and the blanking distance result in no velocity data being available in the top 2.61m of the surface waters. In addition, the ADCP does not provide data in the bottom 6% of depth (RD Instruments 1996). We assume that the velocity is constant in the upper 2.61m of surface waters and in the bottom 6% of depth. Upon examination of the processed ADCP data, we observe that boundary effects are present along the sides and bottom of the river cross-sections; otherwise, there is very little crosswise variation in the velocity structure. We feel that using a velocity field that varies vertically but not across the boundary is a reasonable representation of the velocity field. At present, we use a velocity profile that is averaged over the middle third of the river's cross-section at the boundaries. Time series of measured temperature (T), salinity (S), and normal velocity (v_n) profiles at Brandy and Gorge are shown in Figure 2.

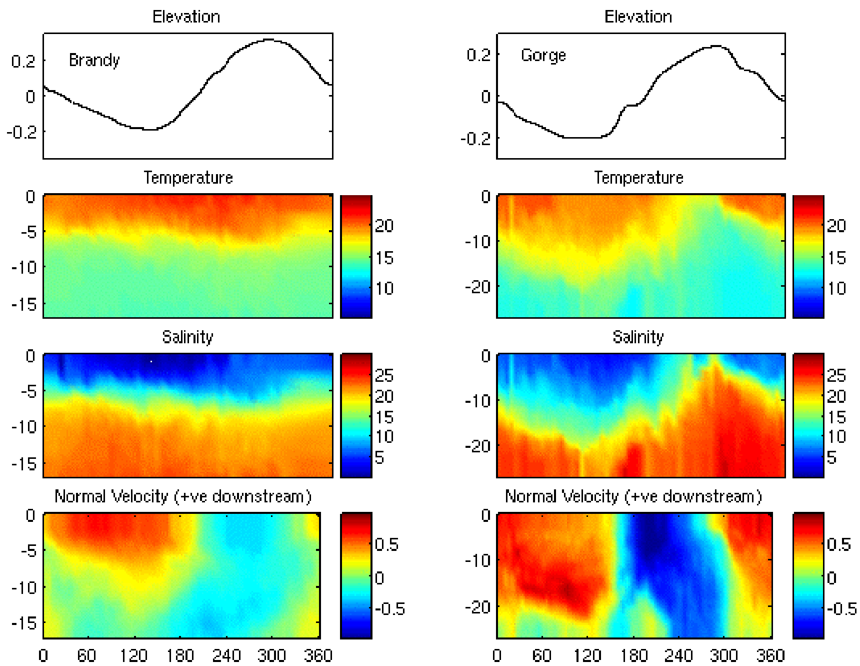


Figure 2. Measured elevation, T, S and v_n at Brandy (left column) and Gorge (right column) from July 2001 surveys. All fields are shown as a function of the phase of the M2 tide (in degrees).

During the ebb tide, the river is flowing downstream at both boundaries, with a much stronger flow at Gorge. Gorge is located at the constriction of the Saint John River just above the Reversing Falls. As the river cross-section is smaller at Gorge than at Brandy, it is expected that the velocity magnitude be greater at Gorge. In addition, a fresh warm layer of water originating from the Saint John River is lying over a salty cool layer of water. The lower layer is a combination of water from the Saint John River and the Bay of Fundy which has been well mixed in the Reversing Falls. During the flood tide, the currents at both boundaries reverse and head upstream. At Brandy the temperature and salinity structures do not change significantly throughout the tidal cycle. At Gorge, there is an influx of cool salty water from the Reversing Falls at high tide and the fresh water layer disappears briefly.

2.2 19th June 2003 Survey – Main River Axis

In June 2003, OMG conducted a series of surveys in the Saint John River Estuary and Kennebecasis Bay. Here we focus on two surveys which show the intrusion of salt water into the estuary over a tidal cycle. Data acquired during these surveys are important to this numerical study as they allow us to examine how well our model reproduces important features of the salt water intrusion.

On 19 June 2003 OMG conducted a survey along the main river axis from just North of Gorge to Brandy. ADCP and underway CTD (MVP-30) observations along this transect at high water and low water are shown in Figure 3. At high water the cold salty water from the Reversing Falls is flowing upstream creating a salt water wedge-like intrusion along the bottom. At Gorge, the salt water actually reaches the surface briefly. Figure 3 indicates that this intrusion does not competely make it over the sill to replenish the salt water layer at Brandy. This suggests that the salt water at Brandy is not necessarily replenished at each tidal cycle (we are now 4 days away from the highest spring tide). At low tide, the warm fresh upper layer is flowing strongly downstream. The observations indicate that, on the ebb, the intrusion at high tide has now been reduced to a near stationary thin layer (< 2m thick) over the sill which never disappears. The bulk of the salt water is actually pushed back into the deeper sections of the Gorge, but, based on other sections not reported here, is not completely flushed out over the Reversing Falls. The Gorge thus acts as a resevoir for the saline intrusion until the next tide.

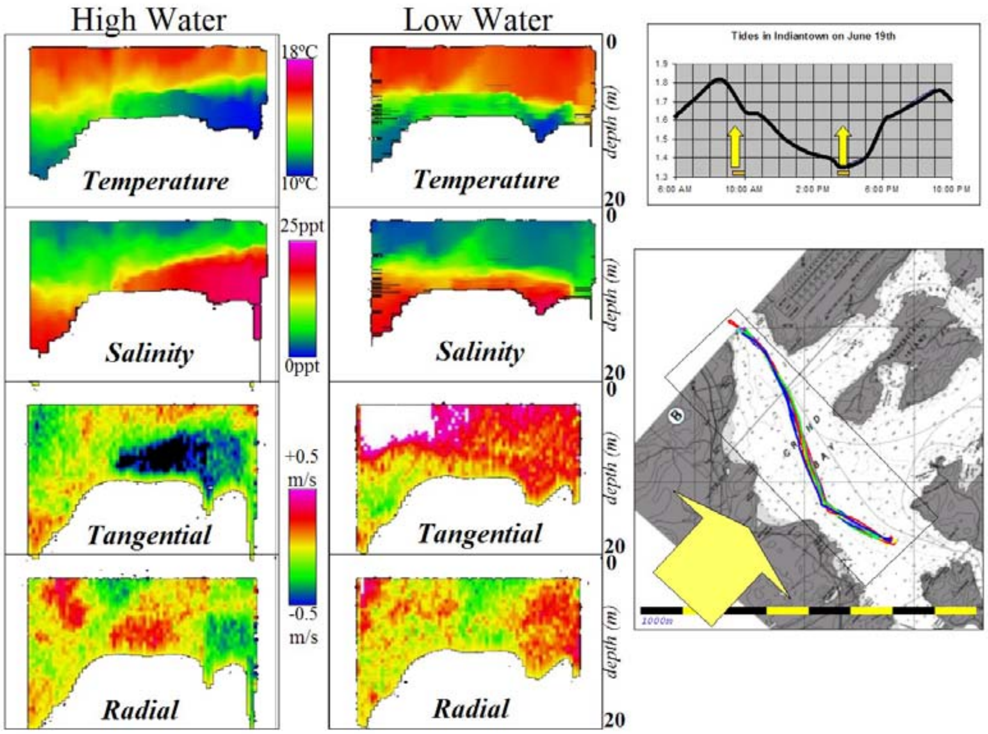


Figure 3. Observations made in the Saint John River Estuary on 19 June 2003. The tidal phases at which the observations were made are shown with yellow arrows on the elevation plot (top right). The locations of the observations are shown on the chart of the river (bottom right). T and S are shown along the transect with Brandy to the left and Gorge on the right. Tangential and radial velocities are positive downstream (to the right) and eastward (into the page), respectively. Currents over -0.5ms are black and over +0.5 m/s are white .

2.3 13th June 2003 Survey - Sill to Kennebecasis Bay Intrusion

On 13 June 2003, OMG conducted a survey between the Saint John River and Kennebecasis Bay. The survey took place over an entire tidal cycle. Figure 4 shows the observations along the transect from Grand Bay over the Boars Head Sill and into Kennebecasis Bay at two hours after low water and two hours after high water. Kennebecasis Bay has a permanent pycnocline. The temperature and salinity of the deep water remain at approximatly 4.9°C and 21.8ppt, respectively. The location of the pycnocline moves up and down throughout the tidal cycle, rising to a minimum depth of approximately 9m at the midpoint of the ebb tide and reaching a maximum depth of approximatly 15m at the midpoint of the flood tide. There is also an internal wave at the density interface which appears to be triggered by water spilling over the sill into Kennebecasis Bay during high tide. The wave travels along the density interface into Kennebecasis Bay. The formation of this wave can be seen in the T and S profiles at two hours past high water in Figure 4. An internal density wave in Kennebecasis Bay of the tidal period was also observed by Page (1979). Of interest is the temperature and salinity of the injected water. It has the same temperature and salinity of the water that occurs at the density interface. The tangential velocity profile in Figure 4 shows water flowing along the density interface into Kennebecasis Bay. This is in contrast to what Trites (1960) saw in his summer survey: the new water entering Kennebecasis Bay moved over the sill and towards the bottom implying that it was saltier and colder. As OMG's survey was conducted in early summer, there was likely a larger discharge of fresh water from the Saint John River which would prevent the salt water intrusion from progressing far enough upstream to spill into Kennnebecasis Bay.

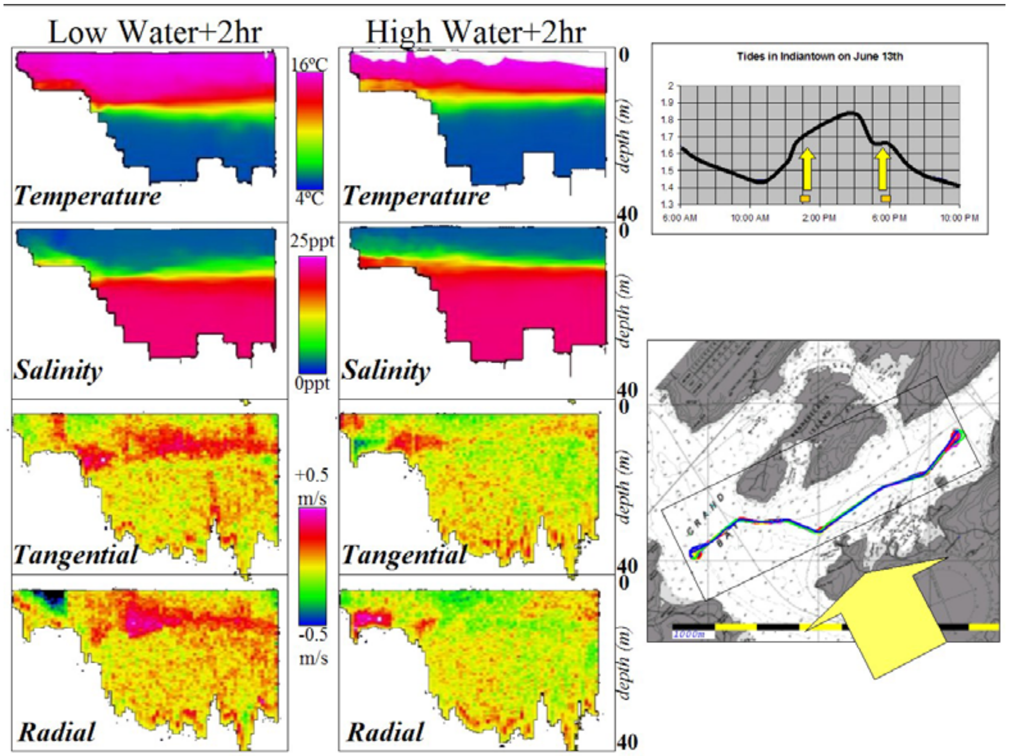


Figure 4. Observations made from the Saint John River into Kennebecasis Bay on 13 June 2003. See Figure 3 for details. T and S are shown with Grand Bay to the left and Kennebecasis Bay to the right. Tangential velocity is positive to the right and radial velocity is positive into the page.

3. NUMERICAL MODEL

We are using the three-dimensional finite element shelf circulation model QUODDY (Lynch et al. 1996). QUODDY is a nonlinear, free-surface, tide-resolving model. It uses general terrain-following vertical coordinates (sigma-coordinates) which provide tracking of the free surface. QUODDY is driven by rotation, tide, wind, and barotropic and baroclinic pressure gradients. For the present study, the effects of wind are ignored.

3.1 Finite Element Grid

A triangular grid was generated using BatTri (Bilgili et al. 2004). The model's bathymetry was computed by interpolating bathymetric data from Canadian Hydrographic Service (CHS) Chart 4141. The model's coastline was created from the 2m contour line in the river portion of the model and from the 15m contour in Kennebecasis Bay. The coastline in Kennebecasis Bay is very steep which results in contour lines being very close together. The deeper contour was chosen as sigma-coordinate models perform better if the change in depth over an element is minimized. The current grid has 5375 nodes and 9674 elements. The two-dimensional elements range in size from 814m^2 to 24,013m^2. In the vertical 21 levels are used. From the surface down to 15m of depth vertical nodes are equally spaced at intervals of 1m. The remaining 5 layers are equally spaced amoung the remaining depth. If the depth is less than 20m, then all 21 layers are equally spaced throughout the entire water column. This vertical grid was chosen to resolve the pycnocline in Kennebecasis Bay.

3.2 Initial Conditions

Properly initialization of temperature and salinity is particularly important in Kennebecasis Bay as there is very little water movement in the bay itself. In the Saint John River portion of the model the effects of the tides are strong and it is felt that accurate initial conditions are not as crucial since any errors would be advected out of the domain. For these reasons the model is initialized with data from the 13 June 2003 survey.

3.3 Boundary Conditions

The two open boundaries are forced as follows. At each time step, the elevation is specified using the observed values described above. QUODDY calculates the resulting velocity, temperature and salinity fields. Before proceeding to the next time step, we adjust the computed velocity, temperature and salinity fields by relaxing towards the observed normal velocity, temperature, and salinity fields at each of the boundaries. For the run described here, we used a relaxation time of 5 minutes. These boundary conditions imply that our modelled boundary conditions should not stray too far from the observed conditions. In particular, they guarantee that water advected into our model domain inherits T and S of that depth from the observed data.

3.4 Results

We are running QUODDY in the baroclinic mode. The model starts at rest with zero elevation everywhere. The elevation and normal velocities are slowly ramped up over the first tidal cycle. We run the model for a total of six tidal cycles and examine the results of the last tidal cycle. There are two areas in which we are interested: the Saint John River portion of the model and the entrance to Kennebecasis Bay corresponding to the surveys of 19 June 2003 and 13 June 2003 respectively.

The dynamics of the Saint John River Estuary are controlled by the forcing at the open boundaries Brandy and Gorge, representing the influence of the river dischage and tidal inflow respectively. Thus it is important to know how well the model reproduces the boundary conditions at the two open boundaries. As we are relaxing towards temperature, salinity and normal velocity at the boundary, we expect the model to reproduce the measured values reasonably well. Figure 5 shows the modelled elevation, temperature, salinity and normal velocity at the Brandy and Gorge boundaries. Comparing the computed values in Figure 5 with the observed values in Figure 2 we notice that the model predicts reasonable temperature and salinity structures when water is flowing into the domain. This occurs during the ebb tide at Brandy and the flood tide at Gorge. At both Brandy and Gorge, the salinity of the bottom water in the model is less salty than those observed. The modelled bottom temperatures are higher than those observed. When water is flowing out of the model boundary, temperature and salinity structures are influenced by the temperature and salinity inside the model domain and do not match the observed values as well. In particular, the lower salty layer at Gorge almost dissappears during the ebb tide. Normal velocities are reproduced reasonably well in terms of direction and timing throughout the tidal cycle. At Gorge, normal velocities are weaker at all phases of tidal cycle. At Brandy, the upstream flow during the flood tide is stronger in the model than observed.

As noted above, there is a difference between modelled temperature and salinity and the observed values. These differences lead us to question how well the model conserves the total salt and heat. Ideally the model would be in equilibrium and there would be no net loss or gain of heat or salt over a tidal cycle. Figure 6 shows the total salt and heat content in our model throughout the simulation. It is clear that we are losing salt and, at a much slower rate, gaining heat. There are several possible reasons for this. As we initialized the temperature and salinity fields with data from the June 2003 data and are forcing the boundaries with data from the July 2001 survey, there is a mis-match between the initial conditions and the boundary conditions. As the model adjusts to the proper boundary conditions, this can cause a change in the total heat and salt budget. In addition, the boundary conditions themselves could be contributing to the changes in the heat and salt content. The boundary conditions were measured over a tidal cycle and we have assumed that the flow is periodic with a period of 12.42 hours. In reality, the flow in the Saint John River estuary is much more complicated with the tides and fresh water discharge changing continuously. Likely both of the above factors play a role in the modelled heat and salt budget.

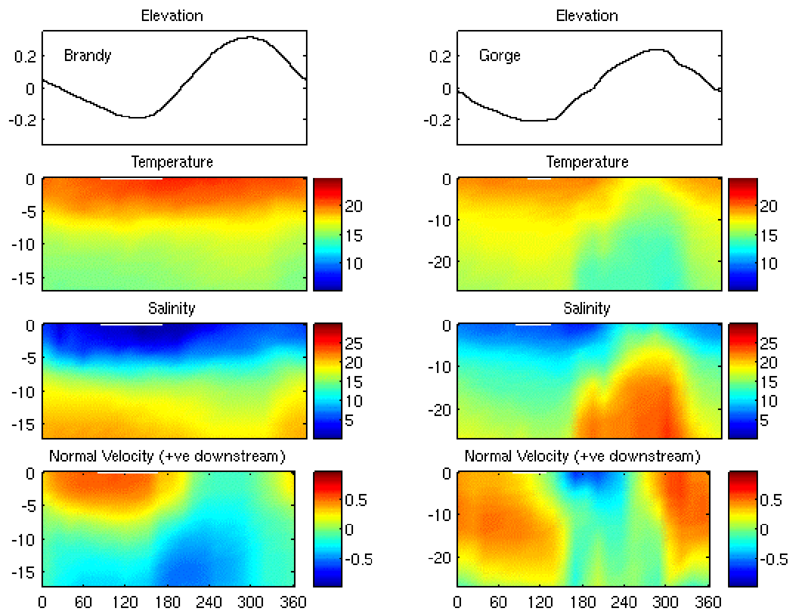


Figure 5. Modelled elevation, T, S and v_n at the Brandy (left column) and Gorge (right column). Results for Brandy are for node 4060 and those for Gorge for node 5014. See Figure 7 for node locations.

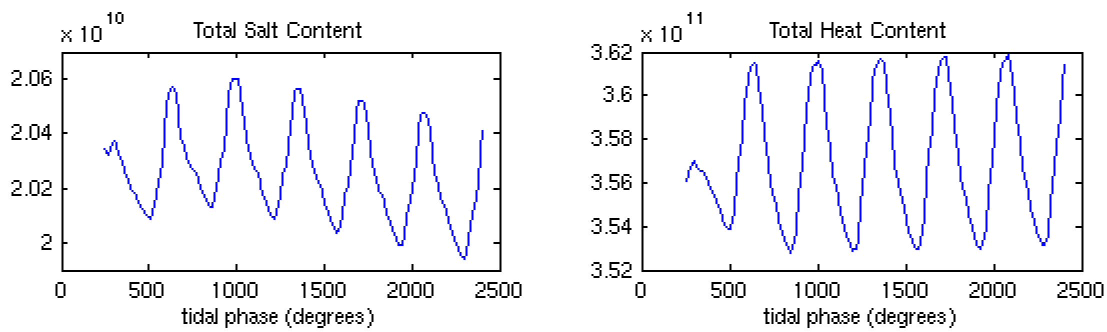


Figure 6. Total salt and heat content of our computational domain as a function of tidal phase.

The model reproduces the gross expected and observed aspects of the flow in the Saint John River Estuary including residual, phase and intrusion characteristics. To illustrate how the model is performing along the main axis of the Saint John River we examine the transect from Brandy to Gorge which corresponds to that of the 19 June 2003 survey (Figure 3). Figure 7 shows the instantaneous behaviour of the model along this transect for the two phases of the tide shown in Figure 3. At high water, the salt intrusion has a similar behaviour as seen in the observations with the cool salty water coming in from the Reversing Falls not making it far enough upstream to replenish the salt water layer at Brandy. The observations show that the salt wedge intrusion is still flowing upstream with the surface waters having already reversed and flowing downstream. The model results are similar but the amplitude of the salt wedge's upstream velocity is much smaller. At low tide there is reasonably good agreement between the observations and the model. The main difference is that the model produces a thinner fresh layer and a thicker salt layer than was observed.

Finally we examine the model results from the Saint John River into Kennebecasis Bay corresponding to the 13 June 2003 survey (Figure 4). Kennebecasis Bay is a fjord-like estuary. It is deep with depths of up to 60m and has a permanent pycnocline with warm fresh water lying over cold salty water. The water in

Kennebecasis Bay has very little movement though Trites (1960) observed speeds exceeding 30cm/s at some depths with the flow confined to the upper 30m. Trites (1960) found that over a tidal cycle there is a slight seaward movement of the surface layer and a flow into Kennebecasis Bay below this. It is also known that cold salty water from the Reversing Falls can be injected into Kennebecasis Bay during the flood tide as observed by Trites (1960) in his summer survey and as well as in the OMG's 13 June 2003 survey. Figure 8 shows the instantaneous flow in Kennebecasis Bay along a transect from the Saint John River into the bay. The instantaneous flow is shown at the same phases of the tidal cycle as shown in Figure 4 : two hours past low water when the pycnocline is at its lowest and two hours past high water when the pycnocline is at its highest. Comparing Figures 4 and 8 the first thing that we notice is that the modelled temperatures and salinity are warmer and fresher than the observed temperature and salinity. This is consistent with the findings in Figure 6 where we observe a net decrease in the salt content and increase in the heat content in our model domain.

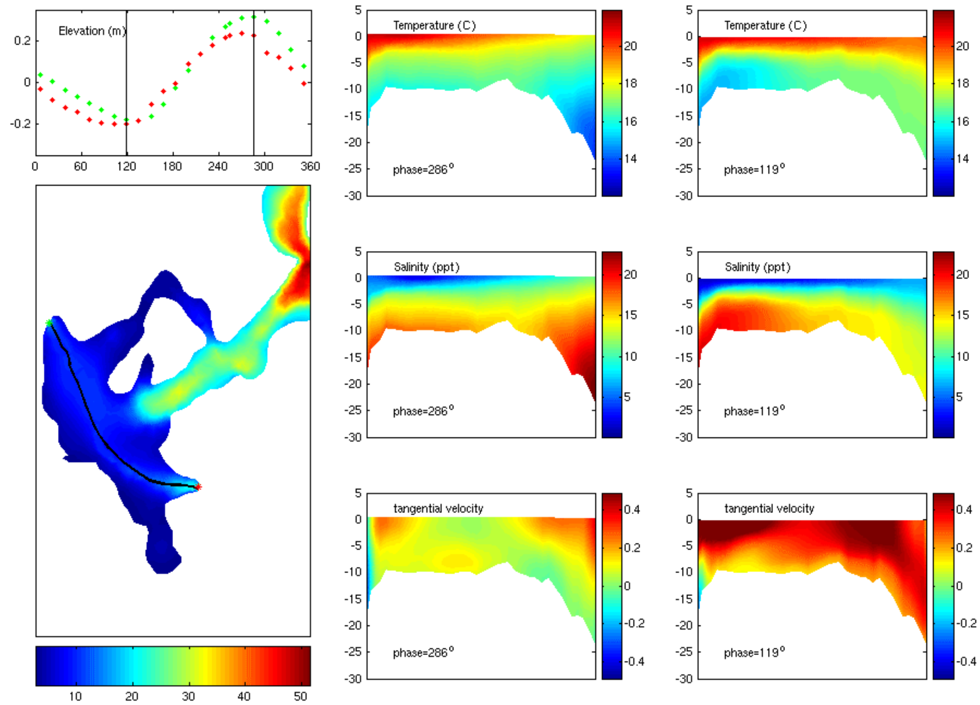


Figure 7. Flow along transect indicated by the black line in bathymetric depth plot (bottom left) at the indicated phases of the tide (top left). T, S, and v_n (+ve to the right) plots have Brandy at left and Gorge on the right. The * is node 4060 and the * is node 5014.

One feature of the numerical model is that it predicts a flow in Kennebecasis Bay itself. Figure 8 shows that there is a flow into Kennebecasis Bay in the surface waters and a return flow in the bottom waters. This is the exact opposite of what Trites (1960) observed. Observations of the 13 June 2003 survey (Figure 4) show very little movement in Kennebecasis Bay other than the injection of water along the density interface. We believe that the flow produced by our simulation is a numerical error in the calculation of the baroclinic pressure gradient. In sigma coordinates, errors occur in the calculation of the baroclinic pressure gradient when there is a sudden change in topography in waters where there are strong density gradients. This type of error is well known, see for example Haney (1991). The steep slope from the sill into Kennebecasis Bay crosses a particularily sharp density gradient. Unfortunately, unlike the main part of the shallow Grand Bay to the west, there is little flow in Kennebecasis Bay to advect these errors out of the system and thus they accumulate, resulting in this residual circulation.

The modelled pycnocline is significantly thicker than the measured pycnocline. In the 13 June 2003 survey, there is a thickening of the pycnocline as a disturbance moves along the density interface but this thickening is transient. In contrast, the thickness of the modelled pycnocline changes very little throughout the tidal cycle. Initially we thought that the thicker pycnocline was likely due to the diffusion coefficient used by the numerical model being larger than the actual physical diffusion in Kennebecasis Bay. As there is little water movement in Kennebecasis Bay, the vertical mixing of temperature and salinity is dominated by the molecular diffusivity of these agents. In pure water the diffusivity of heat and salt (.01 molar) at 15°C is $14\times10^{-8}m^2/s$ (Fischer et al. 1979) and $0.1493\times10^{-8}m^2/s$ (Weast 1988), respectively. In QUODDY, we had originally set the minimum vertical eddy diffusivity for heat and salt to $10^{-4}m^2/s$. The simulation discussed here has a minimum value of $10^{-8}m^2/s$ with little improvement in the thickness of the pycnocline. In Figure 8, the halocline follows closely the tangential velocity distribution indicating that the thickning of the pycnocline is likely due to the flow shear produced by the error in the calculation of the baroclinic pressure gradient.

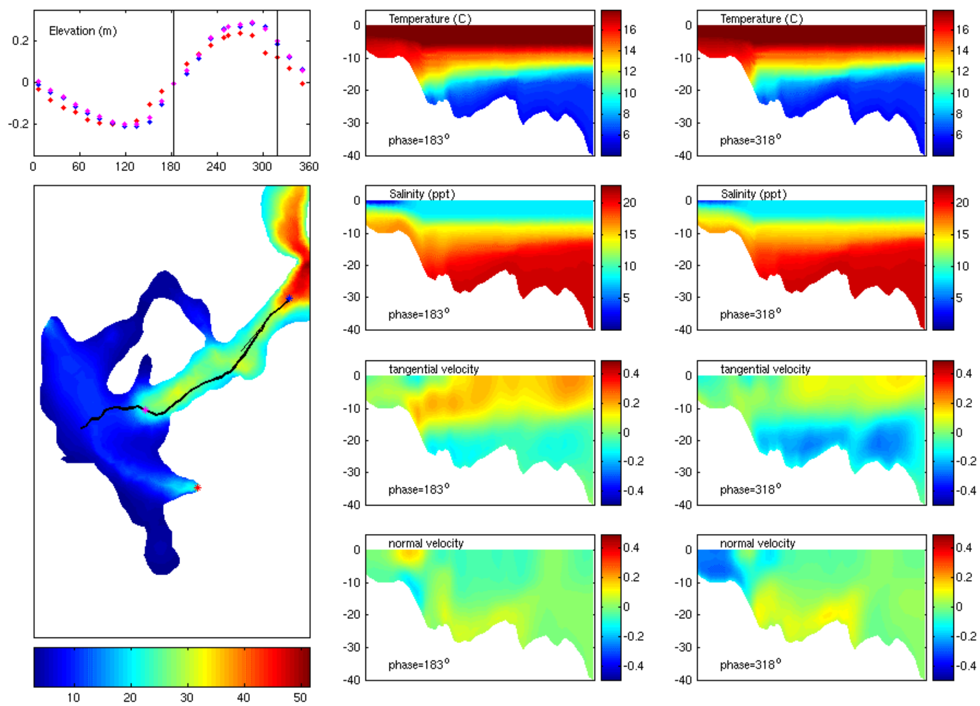


Figure 8. Flow along a transect indicated by the black line in bathymetric depth plot (bottom left) at the indicated phases of the tide (top left). T, S, v_n and tangential velocity (v_t) plots have the St. John River to the left and Kennebecasis Bay to the right. *, *, and * in the bathymetric depth plot indicate the locations at which the elevations are shown. v_t and v_n are positive to the right and into the page, respectively.

Finally, Figure 8 shows salt water of about 16ppt from the Saint John River estuary connected with salt water in the halocline of Kennebecasis Bay. This is similar to the situation shown in Figure 4. However, it is not clear from Figure 8 which direction this water is flowing. In addition, this salt water remains present in the Saint John River estuary throughout the tidal cycle and is always attached to the water in Kennebecasis Bay. Observations on 13 June 2003 show that the salt water layer disappears in the Saint John River Estuary during the ebb tide and thus breaks off from Kennebecasis Bay. The salt water layer does not reappear until high tide. In addition, observations show that the pycnocline does not rise to the level of the sill between the Saint John River and Kennebecasis Bay until high tide. Although the numerical model does predict a rise and fall of the pycnocline with the tide, due to its thickness, a portion of the pycnocline is always level with the sill and thus the 16ppt water in Kennebecasis Bay remain connected to the Saint John River Estuary throughout the entire tidal cycle.

4. CONCLUSIONS

The hydrodynamic model used herein reproduces the circulation along the NNW-SSE channel axis in Grand Bay reasonably well. This section reflects the net upstream passage of saline water into the higher sections of the Saint John River. The model reproduces the observed discontinuous nature of the salt water connection between the Reversing Falls and upstream of Brandy. This has important implications for the known conventional estuarine circulation upstream which requires a reasonably continuous flux of saline water to replenish the lower layer.

The model has more trouble with reproducing the circulation in the Kennebecasis Fjord. Due to limitations of the sigma coordinate system where the vertical nodes are forced to traverse an extreme but mainly level pycnocline on the steeply sloping edges of the fjord, a series of false secondary flow cells have been introduced. This secondary flow generates apparent shear at the pycnocline which causes false turbulent mixing. This mixing accelerates the diffusion, thereby reducing the density gradient and diverging from the observations. The nature of the observed intrusion of mixed water (16 ppt) along the pycnocline rather than into the lower saline watermass, is however, reproduced in the model. This mechanism of injection differs from the bottom hugging flows inferred by earlier work (Trites 1960). This may reflect a mechanism not resolved in the earlier sparse observations, that occurs when the undiluted saline water from the gorge is unable to completely breach the sill.

5. ACKNOWLEDGEMENTS

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