Hydrodynamics of eastern Lake Ontario and the upper St. Lawrence River

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Abstract

Eastern Lake Ontario and the upper St. Lawrence River provide drinking water for approximately 175,000 people. To understand the flow dynamics surrounding the eight drinking water intakes in this region, the hydrodynamics were simulated using the Estuary and Lake Computer Model (ELCOM) for the period April–October 2006. Model simulated water levels, temperatures, and current velocities were compared with observations. Root-mean-square errors in temperature and current simulation were ~2 °C and ~5–8 cm s−1, respectively. Normalized Fourier norms ranged from 0.8 to 1.2. These errors are consistent with other applications of Reynolds-averaged models to the Great Lakes. ELCOM thus reasonably captures the dynamics of the flow regimes in the nearshore region. The flow was found to be predominantly wind induced in the southwestern lacustrine portion of the domain, with observed but not modeled weak near-inertial oscillations, and hydraulically driven in the northeastern riverine portion. Diurnal and semi-diurnal forcing influenced the flow throughout the domain. Flow reversal of the St. Lawrence River near Kingston occurred during storm events. The model results were applied to delineate Intake Protection Zones surrounding the intake protection areas.

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We applied the Estuary and Lake Computer Model (ELCOM; Hodges et al., 2000). It is a three-dimensional, hydrodynamic, two-coordinate model used for predicting the velocity, temperature, and salinity distribution in natural water bodies subjected to external environmental forcing such as wind stress, surface thermodynamics, and inflows and outflows. It is designed to facilitate modeling studies of aquatic systems over seasonal time scales, although the limit of computational feasibility depends on the size of the lake, the resolution requirements, and available computational resources. The model solves the unsteady, viscous Navier–Stokes equations for incompressible flow using the hydrostatic assumption for pressure (Dallimore and Hodges, 2000). Modeled and simulated processes include baroclinic and barotropic responses, rotational effects, tidal forcing, wind stresses, surface thermal forcing, inflows, outflows, and transport of salt, heat, and passive scalars. The hydrodynamic algorithms in the model are based on the Euler–Lagrange method for advection of momentum with a conjugate-gradient solution for the free-surface height (Hodges, 2000). ELCOM has an eddy-viscosity/diffusivity closure scheme for horizontal turbulence correlations and employs the ULTIMATE QUICKEST advection scheme for scalars. Vertical mixing was computed using an explicit turbulent kinetic energy budget closure scheme that was applied to each individual water column of the three-dimensional (3D) flow matrix during each model time step (Hodges et al., 2000). Turbulent kinetic energy was introduced at the surface from surface wind stress and at the lake bed through a bottom shear drag coefficient parameterization. The extent of mixing performed in a single model time step was limited by a mixing time scale, allowing for partial mixing of vertically adjacent cells (Laval et al., 2003). Momentum was introduced into the water column by wind stress at the surface and was distributed vertically by the 3D mixed-layer model.

### Model setup

The model grid had a uniform horizontal resolution of 300 m x 300 m. The vertical grid had 73 layers with varying resolution, chosen to reproduce the observed vertical temperature stratification with reasonable computational effort (see Hall, 2008; Boegman and Rao, 2010). The top three layers were at 0.1 m spacing and the next 40 layers (through the epilimnion and metalimnion) at 0.5 m spacing. The lower layer spacing (in the hypolimnion) ranged from 1 m to 5 m, at the maximum depth of 70 m. The model grid covered the region from the Sandhurst Shores intake in the southwest to the Brockville intake in the northeast (Fig. 1). A grid sensitivity study showed that this could be due to differences in the parameterization of vertical mixing. ELCOM provided better results in the mixed layer depth compared to the other models in the offshore areas. All the models showed substantial error in simulating subsurface currents. The model has not been comprehensively verified in high-resolution application to nearshore regions with an open boundary to the main lake.

ELCOM has been previously applied to coupled lake–river systems. Morillo et al. (2008) investigated the interaction of two rivers flowing into Couer d’Alene Lake and found that basin morphology and wind speed and direction have influence on the fate and transport of inflowing water from rivers to the lake. Vidal et al. (2005) applied ELCOM to Sau reservoir, which was formed by damming the Ter River, and found that period of the prevailing wind coincided with periods of the third vertical mode. Other applications of ELCOM to reservoirs and reservoirs connected to a river can be found in the literature (e.g., Botelho and Imberger, 2007). ELCOM has not been applied to a system consisting of the headwaters of a river emanating from a large lake, where the water level is controlled by a downstream hydraulic structure — as in our application.
The surface boundary conditions were specified, uniformly across the domain, using measured overlap meteorological parameters (described in the next section) from Station 1263 in the Kingston basin. While spatial variability of winds is likely over the region, two-dimensional hydraulic models have shown flows to not significantly differ when spatially variable winds were applied as opposed to uniform winds (Aaron Thompson, personal communication). This is likely because the hydrodynamic response was predominantly wind driven in the eastern portion of the model domain (near Station 1263) and hydraulically driven in the northern section (away from Station 1263). Moreover, application of spatially variable meteorology was not computationally feasible with the number of surface grid points used in this application.

The open boundary to Lake Ontario was forced with observed water levels from the Kingston gauge station (www.meds-sdmm.dfo-mpo.gc.ca) and the observed temperature profile data from Station 1263. Flows at the boundary were computed internally within the control structure on water levels was included by specifying the flow (Tsanis et al., 1991). Moreover, application of spatially variable meteorology was not computationally feasible with the number of surface grid points used in this application.

The forcing data for the model included measured shortwave radiation (W m\(^{-2}\)), net longwave radiation (W m\(^{-2}\)), surface air temperature (°C), wind speed (m s\(^{-1}\)) and wind direction (degrees clockwise from north), relative humidity, atmospheric pressure (Pa), and rainfall amount (mm d\(^{-1}\)) and were input at 10 min intervals. These were measured by the meteorological buoy at Station 1263 mounted 3.3 m above the water surface. The predominant wind direction is southwesterly (Fig. 2). Four storm events (winds > 20 m s\(^{-1}\)) were observed during days 168, 178, 180, and 192.

Water temperature data were recorded using Onset Tidbit temperature loggers at four thermister chain locations (Stations 1262, 1263, 1264, and 1265; Fig. 1, Table 1). At Station 1262, water current speed and direction were recorded using a two-axis MAV ultrasonic current meters. Vertical profiles of currents were obtained at Stations 1263 and 1264 using an upward looking RDI Workhorse Acoustic Doppler Current Profiler (ADCP) with 1 m vertical bins. Hourly water levels measured at Kingston and Brockville gauges were also available.

Quantification of model error

Time series of simulated water levels, temperature, and currents were compared to measured data to assess the model performance. The ability of the model to reproduce the dominant oscillatory barotropic (surface) and baroclinic (internal) processes was determined by calculating energy spectra. Root Mean Square (RMS) error was used to quantify the agreement between model simulated and observed water levels, temperature, and currents:

\[
RMS = \left( \frac{1}{M} \sum_{i=1}^{M} (f_m^o - f_m^i)^2 \right)^{1/2},
\]

where \(f_m^o\) and \(f_m^i\) are modeled and observed water levels, temperature and currents for sample case \(i\) (out of \(M\) sample cases).

To provide a quantitative comparison between modeled and observed currents, normalized Fourier norms were calculated (e.g., Huang et al., 2010):

\[
Fn = \left( \frac{1}{M} \sum_{i=1}^{M} \left[ V_m^o - V_m^i \right]^2 \right)^{1/2} \left/ \left( \frac{1}{M} \sum_{i=1}^{M} V_m^o \right) \right. \]

where \(V_m^o\) and \(V_m^i\) are modeled and observed currents, respectively. The smaller \(Fn\), the better the model results fit the observations. \(Fn\) can also be thought of as the relative percentage of variance in the

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Fig. 2. Wind rose diagram at Station 1263. The length of each wedge denotes percentage of occurrence from a particular direction and the color indicates the wind speed. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

Please cite this article as: Paturi, S., et al., Hydrodynamics of eastern Lake Ontario and the upper St. Lawrence River, J Great Lakes Res (2011), doi:10.1016/j.jglr.2011.09.008
observed currents that is unexplained by the calculated currents. In the case of perfect prediction, $F_n = 0$.

### Results

#### Water levels

The modeled water levels at Kingston and Brockville were compared to the observed gauge water level data (Fig. 3). The modeled water level at Brockville (Fig. 3a) is slightly larger in amplitude than the observed; the RMS error between the modeled and the observed was 4.5 cm. However, at Kingston (Fig. 3b) the modeled water level amplitude was consistent with the observed and the RMS error was 1.1 cm. Modeled results were consistent in phase. The favorable comparisons are not unexpected because the gauge data used for comparison here, which are in close proximity to the data shown, were also applied to force the open boundaries.

Spectral energy plots of the water levels at Kingston (Fig. 4a) demonstrated statistically significant peaks in the model results and field data at 24 h, 12 h, and at the periods of the first (5.06 h), second (3.21 h), and third (2.32 h) longitudinal surface seiche modes (Hamblin, 1982). However, the modeled spectra showed a peak for the fourth (1.71 h) longitudinal seiche mode. This peak was not seen in the observed spectra as the data sample frequency was 1 h (Hamblin, 1982). At Brockville (Fig. 4b) the significant peaks represented the 24 h, 12 h and the first and the third longitudinal surface seiche modes. The 24 h peak was due to the wind forcing (Fig. 4c). There was also evidence of a broad ~10-day peak in the wind and water level spectra, resulting from frontal storm systems characteristic to the region (Hamblin, 1987). The model underestimated these low frequency (5–10 days) surface oscillations caused by wind–induced set-up of the free-surface, potentially due to the constant surface drag formulation. At much lower energy levels, significant peaks corresponding to the first (45 min) transverse seiche mode

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**Table 1**

Table listing details of deployment at the mooring stations by NWRI, Environment Canada. All the moorings were deployed in 2006. The accuracy of the thermister was ±0.2 °C, and that of the MAV and the ADCP was ±0.3 cm s$^{-1}$ and ±0.25 cm s$^{-1}$. The data measured from MAV was averaged from a 60 min burst of 128 2Hz samples.

<table>
<thead>
<tr>
<th>Station</th>
<th>Parameter</th>
<th>Instrument</th>
<th>Deployment period (times in GMT)</th>
<th>Sampling interval (min)</th>
<th>Depth of measurement (m)</th>
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<td>1262</td>
<td>Temperature</td>
<td>Onset Tidbit temperature logger</td>
<td>12 April (12:00) – 26 July (02:10)</td>
<td>10</td>
<td>[1 3 5 7 9.5 13 15 16.5]</td>
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<td></td>
<td>12 April (12:00) – 26 July (14:00)</td>
<td>12 April (12:00) – 26 July (12:00)</td>
<td>10</td>
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<td>30</td>
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<tr>
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<td></td>
<td></td>
<td>11 April – 26 Dec</td>
<td>11.11</td>
<td>[0:15]</td>
</tr>
<tr>
<td>11.9</td>
<td>Current</td>
<td>ADCP</td>
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<td>12 April (12:00) – 26 July (17:10)</td>
<td>10</td>
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<td>MET buoy</td>
<td>27 July (12:00) – 15 Nov (18:10)</td>
<td>27 July (12:00) – 15 Nov (15:00)</td>
<td>30</td>
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<tr>
<td>11.11</td>
<td></td>
<td></td>
<td>1264 Temperature Onset Tidbit temperature logger</td>
<td>27 July (12:00) – 15 Nov (17:40)</td>
<td>10</td>
</tr>
<tr>
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<td>ADCP</td>
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<td>27 July (12:00) – 15 Nov (15:00)</td>
<td>30</td>
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<td>MET buoy</td>
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<td>30</td>
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<tr>
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<td>11.15</td>
<td>Current</td>
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<td>27 July (12:00) – 15 Nov (15:00)</td>
<td>30</td>
</tr>
</tbody>
</table>

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**Fig. 3.** Comparison of hourly water level data measured at (a) Brockville and (b) Kingston gauges with hourly modeled output of water surface height at Brockville and Kingston, respectively.
Comparisons between modeled and observed temperatures north of Amherst and Wolfe Islands at Stations 1262 (Fig. 5a,e) and 1264 (Fig. 5c,g), respectively showed the modeled mixed layer depth to be shallower than observed. This resulted in a cold bias during the summer of ~2 °C. Overall the model captures the seasonal evolution of the temperature profile.

Stations 1263 (Fig. 5b,f) and 1265 (Fig. 5d,h) showed a sharp observed thermocline at depth 15 m. The modeled thermocline was not as abrupt as observed, potentially due to numerical diffusion (Laval et al., 2003) and for the discrete nature of the vertical grid resolution. The warming of the water column, temporal evolution of the stratification, erosion of stratification, and fall cooling were also well simulated at these locations.

The RMS error between modeled and observed temperatures was typically 1–2 °C (Table 2) with a maximum of 3 °C (Fig. 6a,b,c,d). A cold bias of ~2 °C in the model result was seen at Stations 1262 (Fig. 6a), 1263 (Fig. 6b) and 1264 (Fig. 6c). The maximum error (2–3 °C) was observed through the thermocline, where the strong temperature gradient and phase differences between observed and modeled internal waves caused small errors in model skill to lead to larger RMS errors (Dorostkar et al., 2010). These results were consistent with the application of ELCOM and other models to temperate lakes (e.g., Rao et al., 2009; Huang et al., 2010), which have shown maximum errors through the thermocline.

Vertical mode one baroclinic oscillations (internal waves) were evident in the modeled and observed temperature profiles. The strength of these motions may be quantified using the integrated potential (IPE) energy of the water column (Antenucci and Imberger, 2001). Fig. 7a,b shows IPE spectra for Stations 1262 and 1263, respectively, where 24 h and 12 h oscillations were evident in the spectra, with the exception of Station 1263 where the significant energy peak was at the inertial period (~17 h). The lack of an inertial peak in the modeled data likely resulted from the model domain being too small to completely resolve internal Poincare waves, and these waves were not propagating to the domain through the open boundary (see Schwab, 1977 and Rao and Schwab, 2007). The IPE spectral peaks at Stations 1264 and 1265 (Fig. 7c,d) also showed peaks at 24 h and 12 h. These two stations were at the confluence of the St. Lawrence River and the lake, and the peaks resulted from diurnal processes (e.g., sea-breeze winds and day-to-night heating and cooling).

A comparison between modeled and observed, east (Fig. 8a) and north (Fig. 8b) components, MAV currents at Station 1262 shows that the current direction was consistent with the field observations. The RMS errors for the east and north component velocities were 5.7 cm s$^{-1}$ and 3.8 cm s$^{-1}$, respectively (Table 2). The model did not appear to be systematically over/under estimating current velocities, and the RMS errors appeared (Fig. 8) to result from error in modeled direction as opposed to speed.

Comparison of the modeled east–west current profiles at Station 1263 (Fig. 9a,e) reveals that the directions were well modeled, but the model occasionally overestimated the depth to which the strong surface currents penetrated (e.g., days 135 and day 190); comparison of the observed and modeled north–south current profiles at Station 1263 (Fig. 9b,f) shows similar results (e.g., days 185 and 190). These differences may be due to the topographic differences between the 300 m bathymetric grid and the actual bathymetry (Hall, 2008). The...
RMS error magnitude for the east-west and north-south velocity components was 4.0 cm s\(^{-1}\) and 5.1 cm s\(^{-1}\) respectively (Table 2).

At Station 1264, the hydraulic flow through the St. Lawrence River was evident with observed and modeled currents having an easterly mean flow component of ~5–10 cm s\(^{-1}\) (Fig. 9c,g). A baroclinic velocity structure in the modeled data, with a flow reversal about the mid-water column (~5 m depth), is not observed. During this time, the water column was fully mixed (Fig. 5c), and so these flow dynamics did not result from an internal wave but were likely due to wind-induced and hydraulic flow being in opposite directions. The modeled north-south velocity (Fig. 9h) was over-estimated in magnitude at mid-depth with a strong southerly component in comparison to the observed velocity (Fig. 9d) but was consistent in direction.

A wind-induced flow reversal of the St. Lawrence River was observed near days 237–247, 255, and 258 (Fig. 10b), resulting from strong (~10 m s\(^{-1}\)) easterly winds (Fig. 10a). A combined sewer overflow occurred around day 245, making this reversal important for source water protection, as City of Kingston Municipal Engineers were not aware of these flow dynamics, which have the potential to...

Table 2

<table>
<thead>
<tr>
<th>Station</th>
<th>Temperature RMS (°C)</th>
<th>Current RMS (cm s(^{-1}))</th>
<th>Fn</th>
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<tr>
<td></td>
<td></td>
<td>East</td>
<td>North</td>
</tr>
<tr>
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<td>5.68</td>
<td>3.80</td>
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<tr>
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<tr>
<td>1265</td>
<td>1.45</td>
<td>–</td>
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Please cite this article as: Paturi, S., et al., Hydrodynamics of eastern Lake Ontario and the upper St. Lawrence River, J Great Lakes Res (2011). doi:10.1016/j.jglr.2011.09.008
rapidly transport waste up-river toward drinking water intakes. The model reproduced the flow reversals, but the observed currents were stronger at depth and the modeled currents were elevated near the surface (Fig. 10b,c). These differences may result from spatial variability in the surface winds, which were not implemented in the surface forcing. Uniform winds have been applied from Station 1263, but these may be sheltered from the east by Simcoe Island. Winds further along the St. Lawrence River are channeled in an east–west direction through the river valley (Fig. 1). Overall, at Station 1264 the model underestimated the north–south current magnitude by ~4 cm s$^{-1}$ in the surface layer (Fig. 9d,h).

The water column RMS error profiles for current magnitude at Stations 1263 (Fig. 11a) and 1264 (Fig. 11b) showed that model currents were much stronger in magnitude than observed, especially at the surface. The RMS errors were between 1 and 10 cm s$^{-1}$. The east component velocity RMS error at Station 1264 was higher near the bottom due to the riverward flow in the bottom layers around Wolfe Island (Tsanis et al., 1991). Computed Fourier norms ranged from 0.8$<\nu<1.2$ (Table 2). The errors were slightly larger in our nearshore simulations compared to lake-wide simulations (Huang et al., 2010) where 0.4$<\nu<0.9$. This could be due to the complicated topography and presence of numerous islands in the present model.
domain (see discussion in Hall, 2008). In Lake Michigan, Beletsky et al. (2006) reported 0.55 < Fn < 1.59.

Case study – IPZ delineation

Given the above-mentioned importance of local hydrodynamics on spill transport in the upper St. Lawrence River, we present sample IPZ delineations for the Kingston region. Intake Protection Zone One (IPZ-1) was defined at a 1 km radius around an intake, and an IPZ-2 was defined as the distance from each intake where a fluid parcel would undergo a 2-h travel time to reach the intake during 10-yr storm conditions (Ontario Ministry of the Environment, 2005). The 2-h travel time was requested by the drinking water treatment plant operators so that they could shut down the intake in the case of a spill. To reproduce these conditions, ELCON was run with surface winds scaled such that the maximum gust observed during the spring (21.5 m s⁻¹ on day 168), summer (24.0 m s⁻¹ on day 192), and fall (15.6 m s⁻¹ on day 286) seasons matched 10-yr conditions (23.3 m s⁻¹) as observed at the Kingston Airport (Fig. 1) located nearby on the shore of Lake Ontario (National Building Code of Canada, 2005). Wind events were scaled during each season to account for seasonal differences in wind direction, lake level, and stratification. The above validation of the model hydrodynamics to observed data provides faith in the model results, and these results are used to delineate the IPZs.

The 10-yr, 1-hr average wind gusts at Kingston Airport were measured at 10 m above the surface, and so were scaled from 10 m to the 3.3 m height of the Station 1263 wind anemometer using a power law relation (Schertzer, 1987), giving 19.89 m s⁻¹. The winds were then adjusted over 24 h storm duration (maximum gust plus/minus 12 h) such that the maximum 1-hr average wind gust matched the 10-yr condition corrected to 3.3 m. This procedure was repeated for the spring, summer, and fall storm events. IPZ delineations were then constructed at each intake using the 2-h reverse progressive vector (RPV) diagrams calculated from the modeled surface currents at the eight locations during the 24 h storm events. RPVs were deemed sufficient to capture the flow dynamics surrounding each intake because the modeled velocities did not change dramatically within 1–2 km of each intake.

RPV diagrams are similar in principle to reverse particle tracking, whereby fluid parcels continuously released at the intake are tracked backward in time, such that their position within the flow domain 2 h prior to release may be determined. The RPV diagram for Kingston Central (Fig. 12a) and West (Fig. 12b) intake shows the effect of seasonal variability in wind direction and storm intensity on transport of fluid parcels near the intake. It can be seen that the storm event duration (1.59) as observed at the Kingston Airport (Fig. 1) located nearby on the shore of Lake Ontario (National Building Code of Canada, 2005). The above validation of the model hydrodynamics to observed data provides faith in the model results, and these results are used to delineate the IPZs.

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Reversal of the St. Lawrence flow is seen during easterly storm events that cause the flow of contaminants. Strong interconnectivity between the lacustrine and riverine systems is evident. Wind and lake seiche effects are found in the St. Lawrence River as far as Brockville and the IPZ-2 showed a strong polarization from the hydraulic riverine flow. At Gananoque, the effect of flow reversal during easterly wind events was reduced, relative to offshore Kingston, due to wind sheltering and reduced fetch lengths.

**Discussion and conclusions**

The ELCOM model was applied to simulate the hydrodynamics in eastern Lake Ontario and the upper St. Lawrence River and the results were comprehensively validated against observations collected during April–October, 2006. Spectral analysis revealed a significant wind-induced circulation in the Kingston Basin; consistent with the previous studies (Tsantis et al., 1991; Boegman and Rao, 2010) in the region. Strong interconnectivity between the lacustrine and riverine systems is evident. Wind and lake seiche effects are found in the St. Lawrence River as far as Brockville and flow regulation from the dam at Cornwall is evident at Wolfe Island. Here a riverward flow of the bottom water also occurs. The transient hydrodynamics of the region are thus mostly wind influenced, and storm events result in flow reversals, which can have significant impacts on the transport of contaminants.

To accurately model water levels throughout the domain, the influence of the downstream control structure at Cornwall was accounted for, by specifying both the flow rate and water level at the outflow boundary. This suggests that the dam is regulating the flow pro

![Fig. 10](image-url). (a) Hourly averaged east and north components of wind speed collected at Station 1263. The shaded region is the Combined Sewer Outflow event (b) 30 min east velocity component measured at Station 1264 and (c) 30 min modeled east velocity component at Station 1264.

![Fig. 11](image-url). RMS error profile for currents (east and north velocity components) at (a) Station 1263 and (b) Station 1264.
Kelvin waves were not observed or modeled within the domain. This was expected because lake-wide simulations (Boegman and Rao, 2010) have shown that the Kelvin wave does not propagate from the main lake-basin into the shallower Kingston Basin over the Duck-Galoo Ridge. A weak near-inertial Poincaré wave signal was observed at Station 1263. This mooring is within the frictional boundary layer, and consequently a strong Poincaré wave signal was not expected (Rao and Schwab, 2007); the inertial circles became shore parallel and strongly damped to satisfy the no-flux boundary condition. Moreover, the single mooring used for the open-lake boundary condition was insufficient to propagate basin-scale inertial oscillations into the model domain (Schwab, 1977).

Previous modeling of the Kingston Basin flow, with FVCOM, has shown a favorable comparison between modeled and observed summer averaged currents during 1986–87 (Tsanis et al., 1991; Shore, 2009). Comparisons between our 2006 results and these data are not in direct agreement, differing in magnitude but agreeing in direction. These differences may easily result from the variability in wind forcing events and river flows for the different simulation years. However, some similarities are evident; for example, both ELCOM and FVCOM simulated a gyre in the Bay of Quinte region (Fig. 13a). Gyres are also simulated by ELCOM in the channel north of Amherst Island, north of Wolfe Island, and south of Kingston Basin (between Amherst Island and Wolfe Island). Tsanis et al. (1991) observed that the 24-h peak in the spectral signal of water currents around Amherst Island was coherent with the wind but was 180° out of phase with the wind direction. We found a similar 24-h peak in the wind and IPE spectra at nearby Station 1264, which could be due to sea-breeze effects. Similarly, the 24-h signals at Station 1265 could result from daily river flow regulation by the hydroelectric dam at Cornwall (Tsanis and Murthy, 1990). The averaged currents at 20 m (Fig. 13b) showed a return flow from Kingston Basin into Lake Ontario, similar in direction but smaller in magnitude, in comparison to 1986–87 data by Tsanis et al. (1991) and from the application of FVCOM by Shore (2009).

Overall, the modeled hydrodynamics agree well, quantitatively and qualitatively with field observations. The RMS error in temperature is approx 1–2 °C through the epilimnion, when the thermocline deepens during summer. The current speed errors are ~5–8 cm s⁻¹ and appear, at times, to be a result of inconsistencies in flow direction as opposed to momentum transfer from the wind. Current velocity errors are mostly found at exposed regions with large fetch lengths, Stations 1263 and 1264, during strong wind events and consequently
may be due to the inability of the open boundary to capture momentum transfer from the 400 km fetch to the southwest. ELCOM thus reasonably captures the dynamics of the flow regimes in the nearshore region.

In agreement with Rao et al. (2009), the present hydrostatic model application with ~300 m grid resolution was unable to reproduce observed high-frequency temperature fluctuations. These processes are likely associated with nonlinear/non-hydrostatic waves and shear instabilities (e.g., Boegman et al., 2003), and their resolution requires a non-hydrostatic pressure solver and 1~10 m grid resolution. Such simulations are not presently feasible over seasonal timescales (e.g., Botelho and Imberger, 2007; Dorostkar et al., 2010).

Future work will apply the validated ELCOM model to investigate the transport of river plumes and industrial and wastewater effluents within the model domain.

Acknowledgments

Funding was provided by the Ontario Ministry of the Environment, Source Water Protection Technical Studies Program, Environment Canada, and Queen's University. Grid sensitivity runs were performed on computing facilities provided by the Canada Foundation for Innovation and the Ontario Innovation Trust. We thank J. Imberger at the Centre for Water and Research (CWR) for providing the ELCOM source code. Field observations and calibrations were coordinated by Bob Rowsell from Environment Canada. We thank Sean Watt and the CRCA for their support of this study.

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