Three-dimensional modeling of sediment resuspension in a large shallow lake

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Abstract

We simulated bottom resuspension events in Lake Erie, using a coupled three-dimensional hydrodynamic and water quality model. Key parameters in the model, including critical bottom shear stress ($\tau_{cr}$) and resuspension rate ($\alpha$) were calibrated and validated by comparing the model output to observations. These included total suspended solid (TSS) concentrations in the bottom boundary layer (RMSE = 0.74 mg L$^{-1}$) and water column (RMSE = 0.81 mg L$^{-1}$), and to time series of acoustic backscatter signal ($R^2 > 0.8$) and turbidity ($R^2 \approx 0.4$) from long-term moorings near the lakebed in 2008-09 and 2013. Signals from phytoplankton, in spring and summer, caused discrepancies between modeled TSS and the observed turbidity data. Although common practice, we show that literature-based or field-observed critical shear stress should not be directly applied in large-scale Reynolds-averaged sediment model as this will likely underestimate resuspension. In agreement with the literature, the model reproduced more frequent and intensive surface-wave driven resuspension in the shallow regions ($< \sim 20$ m); particularly in the western basin, compared to the deeper central and eastern basins, where bottom stresses induced by mean currents ($\tau_c$) were comparable with those due to surface waves ($\tau_w$). However, on the north-shore of the eastern basin, $\tau_c$ often predominated over $\tau_w$. We simulated thermocline motion, including up- and down-welling events and swashing of the internal Poincaré wave, to contribute to $\tau_c$ in the central basin and form nepheloid layers.

Keywords: Sediment resuspension; hydrodynamic and water quality model; Lake Erie; Surface wave- and mean current-induced bottom stress
1. Introduction

Sediments and total suspended solids (TSS) are important biogeochemical components of aquatic systems, and their abundance influences water quality (Donohue and Molinos, 2009). For example, high TSS concentrations can impact breeding success, egg and larval survival, food availability, feeding efficiency, and recruitment through transport and burial processes (e.g. Bruton (1985); Donohue and Molinos (2009); Schallenberg and Burns (2004); Tuttle-Raycraft et al. (2017)). Increased TSS can reduce light penetration affecting photosynthesis (Bergmann et al., 2004; Fréchette et al., 1989; Gloor et al., 1994; Ji et al., 2002; Tilzer, 1983), and resuspension of sediments can increase sediment oxygen demand (Bruton, 1985). Sources of TSS vary depending on external (river inflows, shoreline erosion) and internal (sediment resuspension) loads. In deep lakes and offshore regions, TSS induced by resuspension is often ignored, but resuspension in shallow lakes and in nearshore regions can contribute significantly to TSS (Graham et al., 2016; Liu and Huang, 2009; Niu et al., 2018; Valipour et al., 2017).

Resuspension events, which are often driven by bottom currents and surface waves, are primarily observed during fall and winter storms in the Great Lakes (e.g. Lake Superior (Churchill et al., 2004; Hawley, 2000), Lake Michigan (Lesht, 1989; Lou et al., 2000), and Lake Erie (Lick et al., 1994). During storms, orbital velocities from wind waves extend to the bed, creating strong wave-induced bottom shear stress that resuspends sediment (Bedford and Abdelrhman, 1987; Valipour et al., 2017).

Whereas field observations provide an opportunity to understand local processes, they are impractical at large spatial scales (i.e., basin scale). Computational models that parameterize lake-wide sediment resuspension provide insight with greater spatial resolution (Beletsky et al., 2003; Lou et al., 2000; Stroud et al., 2009). Recently, Niu et al., (2018) simulated sediment
resuspension and associated enhanced turbidity events in the western Lake Erie, which were consistent with measurements and satellite imagery. Their model, however, lacked near-bed observations needed for the validation of resuspension algorithms and so they employed literature-based model parameters (see also Marti and Imberger (2008); Morales-Marin et al. (2018)). The validity of this approach is questionable, because Reynolds-averaged equation models average resuspension dynamics over large grid cells (~ km), whereas some literature-based model parameters come from observed resuspension events caused by localized turbulent bursts at much smaller scales (~ cm; (Boegman and Ivey, 2009)).

In this study, we applied a three-dimensional hydrodynamic model ELCOM (Estuary and Lake Computer Model) coupled to a biogeochemical model CAEDYM (Computational Aquatic Ecosystem Dynamics Model; currently distributed together as AEM3D; www.hydronumerics.com.au) to model sediment resuspension in Lake Erie. ELCOM has been used to simulate the thermal structure (Liu et al., 2014; Valipour et al., 2019), and basin-scale wave-induced currents in Lake Erie (León et al., 2005; Valipour et al., 2015). ELCOM-CAEDYM has also been applied to simulate temporal and spatial variability of phytoplankton, nutrients, and water quality in Lake Erie (León et al., 2011; Bocaniov and Scavia, 2016; Valipour et al., 2021). Given that the ability of ELCOM-CAEDYM to accurately simulate sediment resuspension has not been examined, in Lake Erie or elsewhere, the primary objective of this study is to validate the model against field, especially near-bed observations, test the validity of directly applying literature-based or field-observed model parameters in low-resolution resuspension models. The secondary objective is to apply the validated model to provide a better understanding of the spatial patterns and mechanisms involved in basin-scale sediment resuspension dynamics in the lake.
2. Methods

a. Study area

Lake Erie is comprised of western, central and eastern basins, with maximum depths of 16 m, 25 m, and 64 m, respectively (Fig. 1). The shallowness of the western and central basins makes them susceptible to resuspension of sediments by wind-induced surface waves (Hawley and Eadie, 2007; Sheng and Lick, 1979; Valipour et al., 2017). A thermocline may form in the western basin on a diel basis (Loewen et al., 2007; Jabarri et al., 2019). In contrast, a seasonal thermocline forms in the central and eastern basins, supporting near-inertial internal (Poincaré) waves with a period ~17 h. These are the dominant wind-induced baroclinic motions during the stratified period (Rao et al., 2008; Valipour et al., 2015).

Sediment type and grain size varies among the three basins of Lake Erie. Haltuch et al. (2000) mapped the spatial distribution of the substrates in each basin, and noted that resuspendible sediment with a grain size less than 63 μm was the most prevalent substrate, at 51%, 61% and 60% in the western, central and eastern basins, respectively.

b. Field measurements

Field measurements were carried out in Lake Erie in 2008 to collect water quality data (see also, Bouffard and Boegman (2013); Bouffard et al. (2012); Liu et al. (2014); Valipour et al. (2017)), which obtained TSS concentration data at various depths from filtered water samples pumped at stations in the western basin (stations 357, 1227), central basin (stations 1228, 341, 1231, 84), and eastern basin (station 452) (Fig. 1; Table 1) during days 203-221 in 2008. During the spring and summer of 2008 and 2009, a tripod was deployed on the lakebed of station 341 (Valipour et al., 2015), equipped with a Nortek Vector acoustic Doppler velocimeter (ADV) at 1 m above the bottom (mab), and multi-parameter water quality sondes (XR-620 and XR-420;
RBR Ltd Canada) included autoranging Seapoint turbidity and fluorescence sensors (accuracy ±2% measured values) at 1.5 mab (XR-620) and 5 mab (XR-420), respectively (Table 1). Additional YSI 6600 sondes with wiped turbidity sensors (accuracy ±2% measured values; Yellow Springs Instruments, USA) were deployed 1 mab at station 1228, in 2008, and stations 67, 475, 591, in 2013 (Table 1; Fig. 1). Wave height and period for 2008 were obtained from a lake buoy located 15 km to the south-west of station 341 (National Data Buoy Center (NDBC)-45005; Fig. 1).

Superficial bed sediment samples were collected at stations 341, 1231 (Table 1) from PONAR box core sampler. The measured particle diameter was $d_{50} = 10 \, \mu m$ (Valipour et al., 2017), which agreed with previous work in central Lake Erie (Fukuda and Lick, 1980; Hawley and Eadie, 2007). According to the particle diameter analysis and Wentworth grain size chart, the bed sediment has 1% fine sand (125 – 250 µm), 4% very fine sand (63 - 125 µm), 75% silt (3.9 – 63 µm), and 20% clay (0.06 - 4 µm).

c. Model description

ELCOM is an unsteady three-dimensional nonlinear z-level Reynolds-Averaged Navier-Stokes (RANS) equation model for incompressible flow using the Boussinesq and hydrostatic approximations. The horizontal advection scheme is based on the Tidal, Residual, Intertidal Mudflat (TRIM) model of Casulli and Cheng (1992). Scalar transport (e.g., temperature, salinity, or tracer) uses a conservative ULTIMATE QUICKEST approach (Leonard, 1991) with a constant eddy viscosity for turbulent closure in the horizontal direction and a mixed layer model, based on a turbulent kinetic energy balance, for turbulent closure in the vertical (Hodges et al., 2000). The free-surface evolution is governed by vertical integration of the continuity equation.
for incompressible flow through the water column applied to the kinematic boundary condition

(Kowalik and Murty, 1993).

CAEDYM is a process-based water quality model (Robson, 2004). When coupled with
ELCOM, biogeochemical variables are updated by CAEDYM after each ELCOM time step
(Romero, 2004). Inorganic particles (e.g., TSS) are modeled with the CAEDYM module, by
accounting for settling and resuspension, with a three-stage numerical algorithm in ELCOM for
scalar transport: (1) vertical mixing by the vertical mixed layer model (Reynolds stress term); (2)
advection of the scalar field by the resolved flow field; (3) horizontal diffusion by turbulent
motions. The scalar variables transport approach in the ELCOM model has been described in
(Hodges et al., 2000) and the science manual (Hodges and Dallimore, 2015) in detail, and the
governing equations are reproduced in the Supplementary information (Table S1).

The TSS equation used in CAEDYM is:

$$\frac{\partial TSS}{\partial t} = \frac{v_s}{\Delta z} \frac{TSS}{TSS} + \alpha \frac{\tau_t - \tau_{cr}}{\tau_{ref}} \frac{\rho_s A}{K_T + \rho_s A}$$

where $TSS$ is the total suspended solid concentration (mg L$^{-1}$), $v_s$ is the settling velocity (m s$^{-1}$,
calculated according to Stokes’ Law based on the user defined particle density ($\rho_s = 2650$ kg m$^{-3}$)
and median sediment diameter ($d_{50}$, m), $\Delta z$ (m) is the height of local z-level layer, and $\alpha$ is the
resuspension rate (g m$^{-2}$ d$^{-1}$). $K_T$ (g m$^{-1}$) controls sediment abundance and $A$ is the area of the
bottom cell (m$^2$). The model assumed an infinite sediment pool in each bottom grid cell. The first
term on the right-hand side of Eq. 1 represents settling process which occurs throughout the water
column, and the second term parameterizes resuspension within the bottom-most layer (Gal, 2009),
which occurs when the total bottom shear stress ($\tau_t$) exceeds the critical shear stress ($\tau_{cr}$). Within
this term, $\tau_{ref}$ is a reference shear stress (set to 1 Pa) and to makes $(\tau_t - \tau_{cr})/\tau_{ref}$ dimensionless.
The total bottom shear stress was computed from the ELCOM simulated mean-flow currents and CAEDYM simulated surface waves:

$$\tau_t = \tau_c + \tau_w = \rho_f (u_{*c}^2 + u_{*w}^2)$$  \hfill (2)$$

where $\tau_c$ and $\tau_w$ are current-induced and surface wave-induced shear stresses (Pa), which were derived, respectively, from the current-induced and surface wave-induced shear velocity $u_{*c}$ and $u_{*w}$ (m s$^{-1}$), and water density $\rho_f$ (kg m$^{-3}$) computed internally in the model. This formulation neglects the effect of wave-current interaction on the bottom shear stress (Grant and Madsen, 1979; Lou et al., 2000). Both $u_{*c}$ and $u_{*w}$ were computed from formulas given in van Rijn (1993):

$$u_{*c} = \sqrt{\frac{f_c \bar{U}^2}{8}}$$  \hfill (3)$$

$$u_{*w} = \sqrt{0.5 f_w U_{orb}^2}$$  \hfill (4)$$

where $\bar{U}$ is the mean current velocity in bottom layer (m s$^{-1}$) provided by ELCOM and $f_c$ is the current friction factor. Here, $f_w$ is the wave friction factor and $U_{orb}$ is the maximum orbital velocity (m s$^{-1}$) computed by CAEDYM. The wave-field parameters are also computed within CAEDYM including wave period, wave height, wave length, and $U_{orb}$ from wind speed, fetch and water depth via linear wave theory (Dean and Dalrymple, 1984), with fetch being computed by the pre-processor at each grid cell along 8 compass directions (Supplementary material, Table S1).

Both $f_c$ and $f_w$ are defined by $d_{50}$ (Swart, 1974; van Rijn, 1993), which is related to bed roughness ($k_s$) through $k_s = 2.5 d_{50}$ (Engelund and Hansen, 1972):
\[ f_c = \frac{0.24}{[\log\left(\frac{12\Delta z_{bot}}{k_s}\right)]^2} \]

\[ f_w = \exp\left[5.213\left(\frac{k_s}{a}\right)^{0.194} - 5.977\right] \]

where \( \Delta z_{bot} \) (m) is the height of the bottom water column grid cell, and \( a \) is the maximum wave amplitude (m) (Supplementary material, Table S1). In Eq. 5, which was derived for hydraulically rough channel flow, \( \Delta z_{bot} \) has been used to replace the channel depth \( h \), to allow for current-induced resuspension in deep systems, such as lakes, with baroclinic flow. This form of Eq. 6 assumes the near-bed water environment is fully rough turbulent flow.

d. Model setup

The model was calibrated for 2008 (days 203-305) and validated for 2009 (days 118-280) and 2013 (days 250-310) using a hydrodynamic configuration based on Liu et al. (2014), including meteorological forcing, inflows, outflows, and the bathymetric grid. A horizontal grid of 2 km × 2 km was used in the lake with 45 vertical layers, including finer 0.5 m grid near the surface and through the thermocline, and coarser 5 m grid in the deep (~ 65 m) eastern basin (Fig. 1c). ELCOM is unconditionally stable for barotropic flows, and so the time step was set at 300 s to satisfy the internal wave Courant-Friedrichs-Lewy condition, \( CFL = \sqrt{2} \) (Hodges and Dallimore, 2015; Hodges et al., 2000).

We initialized the model using observed water temperature profiles throughout the lake from a spring-summer survey (Environment and Climate Change Canada, ECCC); whereas, the initial velocity field was quiescent (‘cold’ start). Spin-up of this shallow wind driven system should be within a 17 h inertial period (Stevens and Imberger, 1996; Valipour et al., 2015). The TSS field was initialized in 2008 with pumped profile observations collected on 21 and 22 July (days 203 and 204) at stations 341, 1227, 1228, 1231 (Table 1), in 2013 with four sets of surface
TSS observations collected on 29 May 2013 (day 149) in the western basin (Fig. 1) by the National Oceanic and Atmospheric Administration (NOAA), and in 2009 with a uniform concentration of the average TSS (1.5 mg L\(^{-1}\)) used to initialize the model in 2008; because there was no observed data. Each initial temperature and TSS profile were first linearly interpolated on to the vertical layers and then horizontally interpolated throughout the domain using the inverse distance weighting method.

Following other published applications of ELCOM to Lake Erie, differing hourly meteorological forcing data (air temperature, wind speed/direction, shortwave solar radiation, relative humidity, and cloud cover) were applied uniformly to model domain sub-sections across the lake (Fig. 1). Sensor heights are internally adjusted to a common reference height of 10 m. In 2008 data were from four station located in the eastern (Port Colborne, ECCC station C45142), east-central (Port Stanley, ECCC station C45132), west-central (station 341), and western (NOAA station SBIO1) basins (Liu et al. 2014). In 2009 and 2013 additional forcing data was used to further subdivide the central basin into four domains by adding northwest (ECCC-ErieAU) southeast (NDBC-GELO1) station data. In 2008, flow, water temperature and TSS concentration were specified for 11 inflows, including the Detroit, Maumee, Raisin, Sandusky, Vermilion, Rocky, Cuyahoga, Grand [Ohio], Cattaraugus, Buffalo and Grand [Ontario] rivers, and only one outflow, the Niagara River. In 2009 and 2013, only data for the five major inflows were available (Detroit, Maumee, Sandusky, Cuyahoga and Grand Rivers). Daily flow rates and water temperatures were from the U.S. Geological Survey (USGS) National Water Information System, the U.S. Environmental Protection Agency database, the Water Survey of Canada (ECCC), the Grand River Conservation Authority, and the Heidelberg College Water Quality Laboratory. The inflow water temperature for the Detroit river was obtained through a 5-day
average of the Windsor air temperature (León et al., 2005; Valipour et al., 2016; Valipour et al.,
2019).

To track the source of sediment loads, we separated TSS from rivers (SSR) and the lakebed
(SSb) into different sediment pools (TSS = SSR + SSb). For SSR, d50 = 3 µm (Fukuda and Lick,
1980); whereas, SSb was divided into two classes: SSb1 (d50 = 1 µm) represents clay-like
superficial (nepheloid) sediments (Lick and Lick, 1988; Lick et al., 1994), and SSb2 (d50 = 10 µm)
represents the silt-like sediments below (Hawley and Eadie, 2007; Valipour et al., 2017). The
sediment classes were proportioned at 20% (clay) and 75% (silt), according to observations from
the PONAR box core sampler. Sediment load from the Maumee River (Supplementary material,
Fig. S1; data from USGS; https://cida.usgs.gov/sediment/) was derived from SSR by multiplying
by flow rate. Due to a lack of observations, the SSR in the Detroit River was set by apportioning
the annual sediment load according to the daily flow rate (Kemp et al., 1977).

We define Tt as the entire simulation time (102, 182 and 59 days in 2008, 2009 and 2013,
respectively) and Tr as the duration of resuspension events (i.e., the time when \( \tau_t > \tau_{cr} \)) so that the
frequency of resuspension events in different parts of the lake can be studied.

3. Results

a. Hydrodynamic model validation

ELCOM simulated temperatures and currents have been previously calibrated and validated
using the same setup as in the present study on 2 km and 500 m horizontal grids for 1994, 2001,
2002, and 2003 (León et al., 2005), 2004 (Oveisy et al., 2014), 2008 (Liu et al., 2014; Valipour et
al., 2015; Bocaniov et al., 2016) and 2013 (Valipour et al., 2019, 2021). We further validate
modeled bottom current velocities, as the key hydrodynamic driver of sediment resuspension (Eq.
2, 3, 4), which showed good agreement with observations in the bottom layer at station 341 in
234 2008 (Fig. 2). Using a bottom drag coefficient of 0.0045 (Valipour et al., 2015), ELCOM was able to reproduce the current velocities with a north-south root-mean-square-error (RMSE) of 0.043 m s\(^{-1}\), and an east-west RMSE of 0.044 m s\(^{-1}\). Using default parameters in the wave model (Table S1), the CAEDYM simulated wave heights, which have not previously been compared to observed data in Lake Erie, underestimated observed wave heights >1.5 m on days 240, 259, and 271. This may be associated with the low temporal resolution of the wind forcing data (1 h). The RMSE for wave height, and wave period were 0.22 m and 1.33 s, respectively. Further validation of current velocities and wave properties in 2009 is shown in the Supplementary Material (Fig. S2).
b. Calibration of the TSS model

The TSS profiles collected at 7 field stations (Table 1, Fig. 1) during days 205-221 in 2008, were used to calibrate the modeled background TSS concentration. These data were infrequent and did not capture the timing of resuspension events, limiting their use for quantitative model validation. Continuous timeseries data, from the turbidity loggers (e.g., Hawley and Zyren (1990); Valipour et al. (2017)) and the beam amplitude of the ADV backscattering signal (hereafter, ADV-amp, unit Count) were, therefore, used to calibrate resuspension events qualitatively. We were, however, unable to correlate turbidity or ADV-amp to TSS, which is only possible if the suspended particles in the water column are homogeneous and within a specific calibration range (Churchill et al., 2004).

The model was calibrated by systematically varying the primary resuspension-related parameters ($\tau_{cr}, K_T$ and $\alpha$) within the expected ranges found in the literature, which vary greatly (Table 2), and using the in situ observed $d_{50}$. Unfortunately, there is no measured nor literature values for $K_T$ and so the value was chosen to ensure that $\alpha$ was within the range reported in Table 2, and the amount of resuspended sediment was consistent with the observed TSS from pumped water samples. Depending on the sediment composition and water content, $\tau_{cr}$ could vary by 1 order of magnitude between freshly deposited and older sediments and $\alpha$ could vary by over 2 orders of magnitude (Fukuda and Lick, 1980; Lick et al., 1994). Thus, the expected ranges for $\tau_{cr}$ and $\alpha$ were set at 0.01-0.3 Pa, and 31.1-864 g m$^{-2}$ d$^{-1}$, respectively (Table 2). Valipour et al. (2017) provided an in situ estimate of $\tau_{cr} = 0.28$ Pa from the dataset at station 341, and found high turbidity events at station 341 when the maximum instantaneous flow velocity (maximum value in each ADV burst; Table 1) $u_{max} = 0.25$ m s$^{-1}$ or $\tau_{max} = \rho C_D u_{max}^2 = 0.28$ Pa, where $C_D = 0.0045$ (Valipour et al., 2015). This corresponds to a 5 min or burst-averaged flow velocity $u_{mean}$
14 m s\(^{-1}\) (Valipour et al., 2017). Given that the field observations were instantaneous point measurements and the model was averaged over 300 s time steps and 2 km \(\times\) 2 km grids, it was not necessarily appropriate to use the observed \(\tau_{cr} = \tau_{max} = 0.28\) Pa as the threshold for resuspension. Indeed, a sensitivity analysis confirmed that resuspension did not occur in the model using \(\tau_{cr} = 0.28\) Pa. Consequently, we began model calibration using the observed burst-averaged value \(\tau_{cr} = \tau_{mean} = \rho C_D u_{mean}^2 = 0.045\) Pa.

Calibration was achieved by maximizing \(R^2\) in comparison to observed ADV-amp/turbidity at Sta. 341 in 2008, and evaluating RMSE using TSS observations at 7 stations across the lake in 2008. Table 3 and Fig 3 show five different parameter combinations (A-E), including the best calibration against the observations (combination C), that illustrate the model response to changes in parameters. For the model runs with \(\tau_{cr} = 0.045\) Pa (combination A, B, E), the resuspension events on days 250 and 259 were captured, but not those on days 224 and 239-242. The modeled TSS concentration, in the bottom layer, was qualitatively inconsistent with ADV-amp and turbidity data, with low \(R^2\) (Table 3). We further decreased \(\tau_{cr}\), based on \textit{in situ} measurements from Lick et al. (1994), Bedford and Abdelrhman (1987) and Hawley (1991) in combination C and D, which captured four significant resuspension events (days 224, 239-242, 250, 259). The value of \(\tau_{cr}\) not only determines whether resuspension occurs or not, but also directly affects the amount of sediment resuspended into the water column (Eq. 1). Therefore, on days 250 and 259, combination C and D reproduced more intense resuspension than combination A, B, E, and had high correlation \((R^2 > 0.8)\) with ADV-amp (Table 3). The amount of sediment resuspended into the water column can also be controlled by adjusting \(\alpha\). The calibrated \(\alpha = 450\) g m\(^{-2}\) d\(^{-1}\), was the median of the observed range (Table 2). The comparison between combination C and D showed lower \(\alpha\) (= 300 g m\(^{-2}\) d\(^{-1}\)) leads to slightly lower
correlation (Table 3), and significant mismatch especially during resuspension events on days 240, 250, and 259.

Overall, combination C achieved a strong association ($R^2 > 0.8$) between modeled $TSS$ concentration and ADV-amp, and moderate association ($R^2 > 0.35$) between modeled $TSS$ concentration and turbidity data (Table 3). Errors in the resuspension simulation could be induced by the underestimation of north-south current velocity and wave height on days 240 and 259 in the model (Fig. 2a, d). The background $TSS$ concentration had an average RMSE = 0.74 mg L$^{-1}$ in the bottom layer and average RMSE = 0.81 mg L$^{-1}$ in the overlying water column over the 7 stations (Table 3). The pumped water samples showed only minor variation in observed $TSS$ concentration; therefore, the RMSE for the above combinations did not vary over the parametric study. The comparisons between modeled and observed $TSS$ at each station are given in the supplementary material (Fig. S3).

Model sensitivity to the horizontal diffusivity and grid size was also investigated (Fig. S4, S5). Although variation in the horizontal grid size (1 km x 1 km vs. 2 km x 2 km) led to small differences (1 km grid model compared against observation with north-south RMSE = 0.04 m s$^{-1}$, and east-west RMSE = 0.03 m s$^{-1}$) in the resolved flow field (Fig. S4a, b) and further affected the advection of $TSS$, it had only a minor effect on the timing of resuspension events (Fig. S4c). The peak $TSS$ values, during simulated resuspension events, were insensitive to the horizontal diffusivity within the range 0 – 10 m$^2$ s$^{-1}$, with 1 m$^2$ s$^{-1}$ being typical and default for a lake or ocean model (Fig. S5a). Since $K_T$ and $\rho_sA$ control the amount of resuspendible sediment at the bed (Eq. 1), $K_T$ was proportionally adjusted according to the area of the bottom cell. When the horizontal grid was changed from 2 km to 1 km, $K_T$ was divided by 4 ($\times 10^9$ g m$^{-1}$). We also investigated variation in $\Delta z_{bot}$ from the 0.5 m calibration value (to 0.75, 1,
and 1.25 m) in Eq. 5, which had a negligible effect on the timing of resuspension events (Fig. S5b).

c. Validation of the TSS model and effects of phytoplankton and particle aggregation

The calibrated TSS model was validated against ADV-amp data at station 341 in 2009, and turbidity data at stations 1228, 591, 67, and 475 in 2008 and 2013 (Fig. 4). In 2008, the model reproduced the high turbidity events at station 1228 on days 233, 240, 250, and 260, but the observed turbidity was continuously high, for more than 10 days, after day 260, which was not simulated by the model (Fig. 4a). Settling of phytoplankton during summer in 2008 and 2009 could be the source of the sustained high turbidity (Valipour et al., 2017). Wynne et al. (2010) reported bloom events in 2008 began on day 243 and persisted for over 2 months in the western basin, with the bloom area extending to the west-central basin (stations 341 and 1228), possibly causing high turbidity after day 260 (Fig. 5c, d). This argument is supported by a spike in chlorophyll-α concentration 5 mab (>2.5 µg L⁻¹) at station 341 on days 255-280 (Fig. 3f), with a corresponding increase in turbidity 5 mab, which is located ~10 km to the east of station 1228.

The fluorescence sensors deployed in 2009 at station 341 and analysis of satellite images by Valipour et al. (2017) could be used to classify some high turbidity events, that were not modeled as originating from phytoplankton (days 117-196; Fig. 4f). The modeled TSS concentration reproduced peaks in ADV-amp on days 120, 130, 135, 146, 240, 260, and 270 in 2009. The turbidity observations at station 341 also captured the resuspension events on days 120, 260, and 270 (Fig. 4f).

It was difficult to distinguish sediment resuspension from phytoplankton in turbidity data at the nearshore eastern basin stations (591, 67, and 475) in 2013. Published water quality
simulations, using CAEDYM (Valipour et al., 2016), suggest high nearshore total chlorophyll-\(a\) concentration (~ 5 \(\mu\text{g L}^{-1}\)) in July through September. The model did reproduce high turbidity events, likely resulting from bottom sediment resuspension, on days 290-310 at stations 591 and 67, but the comparison at station 475 was poor (Fig. 4b, c, d).

Fine particles, like clay, which tend to form aggregates, occupy around 20% of sediment on the lakebed. Unfortunately, CAEDYM is unable to account for sediment cohesiveness and aggregation. Particle aggregation increases the settling velocity (Hawley, 1982), but also limits the quantity of sediment that can be entrained (MacIntyre et al., 1990). Therefore, the computed settling velocity, based on Stokes’ Law, may underestimate sedimentation of cohesive particles, and the uniform entrainment rate may overestimate the amount of sediment resuspended, possibly causing inconsistencies between the observed ADV-amp and modeled TSS (e.g., days 242-250, 260-265 of year 2009 at station 341, Fig. 4f).
4. Discussion

Although there were some discrepancies between the model results and the observations, the model reproduced the timing of most large resuspension events and simulated both the background TSS concentration and the qualitative (Fig. 3, 4) nature of the sediment resuspension. Coupled with the hydrodynamic information output by model, we further explored sediment resuspension and transport at the basin-scale, beyond the constrains of point measurements, and through comparison of our results to published studies on Lake Erie sediment dynamics.

a. Spatial variation and seasonal change in TSS concentration

We selected four stations (357, 341, 84 and 452) from the western, west central, central and eastern basins, respectively, to explore the spatial and seasonal variation in TSS concentration. The modeled temperature profiles have been previously validated at these sites for the present 2008 setup (Liu et al., 2014; Valipour et al., 2015) and for other years (e.g., León et al. (2005); Oveisy et al. (2014)).

In the western basin (station 357; Fig. 5 a, b), the water column was well mixed and the modeled TSS concentration had a relatively homogeneous vertical distribution. After day 240, the TSS concentration became elevated, first from the bottom toward the surface, indicating resuspension.

In the west central basin (station 341; Fig. 5 c, d), the TSS concentration remained low in the stratified period (day 203 to 235) (Boegman et al., 2008). The significant wave height did not exceed 1.2 m until day 235 (Fig. 2 d), and the associated surface wave energy could not extend down to or below the thermocline. Around day 250, a wind event generated wave orbital velocity (Fig. 2c, d) and a down-welling event (Fig. 5c), resulting in sediment resuspension (Fig. 5d). This resuspended sediment remained in the hypolimnion when the water column re-
stratified through up-welling (days 255-270). The locally resuspended sediments formed a nepheloid layer in the hypolimnion at station 341, showing the effect of stratification on the vertical distribution of sediments (Fig. 5c, d). Stratification weakened after day 270 and high TSS concentrations, resulting from the bottom resuspension, were transported into the surface layer on days 291 and 300-303 due to the mixing of the water column.

In the central and eastern basins, the TSS concentrations were much lower compared to the shallower sites. At station 84 (Fig. 5e f), bottom resuspension occurred but with much lower intensity compared to stations shallower than 20 m. Modeled TSS profiles at station 452 indicated that there was no resuspension at this eastern basin site (Fig. 5h). At stations 84 and 452, TSS concentrations were elevated above the thermocline after day 260 (Fig. 5f, h), indicating TSS with low settling velocity ($SS_{B1}$) were advected from shallower parts of the lake, including the west central basin, the Grand River and the internal swash zone at the depth of thermocline (see following Thermocline motion and current-induced resuspension section). This advective mechanism is described in detail in the Supplementary material (Text S1; Fig. S6).

In general, the seasonal pattern showed increased TSS concentrations during fall, at all four stations, suggesting the more frequent storm events drive resuspension (Hawley and Eadie, 2007), which was strongest in the shallow western basin (Fig. 5b). The deeper station in the central basin (station 84) also responded to the large storm events (days 300-303; Fig. 5f). During summer stratification, when the thermocline intersected the bed, in the central basin, resuspension was also modeled (Fig. 5c-f). The bottom sediment response across Lake Erie to the winds during days 300-303 in 2008 can be seen in the supplementary material Movie S1.

b. Bottom shear stress
Storm events transfer momentum to the surface layer, which is the main energy source for resuspension (Chung et al., 2009; Hawley, 2000; Hawley et al., 2004). In Lake Erie, surface waves were the major abiotic factor behind high turbidity events at station 341 during the summers of 2008-09 (Valipour et al., 2017). Thus, the parameterization of $\tau_w$ determines the accuracy of the model. For shallow lakes ($< \sim 20$ m), like large parts of Lake Erie, wave orbital velocities will penetrate to the lake bed and – along with seiche currents – energize a turbulent bottom boundary layer (Lorke and MacIntyre, 2009). This leads to the frequent classification of the near-bed environment as fully rough turbulent flow. However, the decaying oscillatory nature of a lake response to a wind event, results in a quasi-turbulent bottom boundary layer in Lake Erie (e.g., Jabbari et al. 2020). To investigate the validity of a fully rough flow-based $f_w$ equation, we compare the model calculated $\tau_w$ (Eq. 2 - 6), to the value using $f_{w,\text{theory}}$ from wave theory (Jonsson, 1966; van Rijn, 1990)

$$f_{w,\text{theory}} = \left\{ \begin{array}{ll}
2 \left(\frac{a \times U_{orb}}{v}\right)^{-0.5} & (\frac{a \times U_{orb}}{v} < 10^4; \text{laminar flow}) \\
0.09 \left(\frac{a \times U_{orb}}{v}\right)^{-0.2} & (10^5 \frac{a \times U_{orb}}{v} > 10^4; \text{smooth turbulent flow}) \\
\exp \left[-6 + 5.2 \left(\frac{a}{k\tau}\right)^{-0.19}\right] & (\frac{a \times U_{orb}}{v} > 10^5; \text{rough turbulent flow})
\end{array} \right.$$  

(1)

at stations 341 (17.5 m deep) and 591 (14.8 m deep), where the modeled TSS concentrations have been validated. Overall, the modeled $\tau_w$ was lower than theoretical $\tau_w$ at the peak values (Fig. 6). At station 341, obvious discrepancies occurred in 2008 (Fig. 6a, days 233, 240, 263, 270, 290, 300) and 2009 (Fig. 6b), yielding a RMSE = 0.014 Pa. At station 591, the modeled $\tau_w$ was close to the theoretical value on days 280, 300, and 305 in 2013, when both exceed 0.15 Pa, but was lower than the theoretical value during days 289-295, when $\tau_w$ was over the critical stress, but less than 0.1 Pa (Fig. 6c). The model underestimation of peak values in $\tau_w$, during resuspension events, could explain why the critical stress for resuspension in the model...
(0.025 Pa) is lower than the observed burst-averaged value (0.045 Pa, Valipour et al., 2017). This error could induce biases in the quantity of resuspended sediments.

We have shown that to reproduce resuspension, in Lake Erie with a Reynolds-averaged model, we must apply a critical bottom stress (0.025 Pa) that is less than both the observed value using a local mean current (0.045 Pa) and the observed value using a local instantaneous current (0.28 Pa). This, we speculate, is due to both spatial averaging over the large model grid cells, which will filter out local peaks in current, and also inaccuracies in the model formulation (see text on Eq. 7). Therefore, blind application of field-observed or literature-based resuspension parameters, without adjustment based on near-bed validation data is not justified; examples of this in the literature abound. Niu et al. (2018) applied a literature-based critical stress of 0.05 to 0.07 Pa in western Lake Erie using FVCOM, yet only validated their model against surface TSS and satellite data. Marti et al. (2007) applied ELCOM-CAEDYM to simulate intermediate nepheloid layers in Lake Kinneret with literature-based critical stresses of 0.06 and 0.12 Pa and qualitatively compared their simulations the presence of a nepheloid layer in the lake. Morales-Marin et al. (2018) simulated a small Welsh lake (Llyn Conwy) with FVCOM using literature-based critical shear stresses ranging from 0.05 to 0.2 Pa, with no sediment observations. Mirbach and Lang (2018) simulated transport of and resuspension by density currents from Rhine River underflows in Lake Constance using ELCOM-CAEDYM, but they did not report their critical stress, nor did they validate their TSS simulations against observed data. Our results suggest these studies have applied field-observed or literature-based critical stresses that are too large and, therefore, underestimated sediment resuspension.

To better understand the major mechanisms that trigger resuspension, the model was compared to observations (Valipour et al., 2017) showing surface waves with wave periods $T > 5$
s and wave heights, \( H > 1.5 \) m were able to resuspend bed material at station 341. The resuspension events in 2008 (days 240, 250, and 300) and in 2009 (days 120, 130, 135, 146, 240, 260, and 270) when wave heights \( \geq 1.1 \) m and wave periods \( \geq 4.5 \) s (Fig. 2d, S2d), were dominated by increases in \( \tau_w \) and \( \sim 58\% \) of \( \tau_t \) was from surface waves at this station (Table 4).

However, on days 224 and 259 in 2008, the contribution from currents \( \tau_c \) was greater than from waves \( \tau_w \), suggesting that mean currents drove sediment resuspension on these days. Even during the surface wave-dominated resuspension events (e.g., days 250 in 2008, and 135 in 2009), \( \tau_c \) exceeded the critical threshold, showing the combined effect of both surface waves and mean currents on sediment resuspension (Lick et al., 1994).

In the nearshore eastern basin (station 591), contributions from mean currents and surface waves were comparable during sudden intensive resuspension events (e.g., days 280, 300, and 305 in 2013), but during the longer resuspension event in fall (i.e., days 290-300 in 2013), mean currents played a dominant role; the contribution of \( \tau_c \) to \( \tau_t \) was \( \sim 85\% \) (Table 4).

The duration of resuspension \( (T_r) \) was \( \sim 12\% \) of the simulation time \( (T_t) \) at both stations 591 and 341. Although station 591 is shallower, the ratio of modeled \( \tau_c \) to \( \tau_t \) at 341 (42\%) and 591 (85\%) (Table 4) indicated that mean currents were more significant for resuspension at the shallower nearshore eastern basin site, compared to offshore in the west central basin. This could be related to wind fetch, depth-limited breaking or coastal circulation patterns (Lick et al., 1994; Valipour et al., 2019).

However, one of the conceptual limitations within model is parameterizing \( \tau_t \) as the summation of \( \tau_w \) and \( \tau_c \) (Eq. 2), and ignoring the effect of wave-current interaction (Grant and Madsen, 1979; Lou et al., 2000). This simplification approach could induce inaccuracy in timing and intensity of resuspension prediction, especially when strong windstorms generate both
intense bottom currents and wave orbitals.

c. Thermocline motion and current-induced resuspension

Bottom currents associated with vertical thermocline motions including upwelling and
down-welling events, and near-inertial internal waves had been found to be associated with near-
bottom high-turbidity events in Lake Michigan (Hawley, 2004; Hawley and Muzzi, 2003).
Observations from Valipour et al. (2017) showed the contribution to resuspension from near-bed
turbulence, induced by high-frequency internal waves in Lake Erie, and this process was
observed when thermocline approached the lakebed. While the RANS model does not
parameterize sub-grid-scale high-frequency internal waves, we found that motions of the
thermocline affected the distribution of TSS and resuspension events. Our simulations revealed
three mechanisms associated with thermocline-related resuspension (Fig. 8).

(1) Resuspension from down-welling

The thermocline downwelled to the lakebed at locations with different depths (Fig. 8 a-c),
leading to a sharp increase of $\tau_c$ and spikes in the modeled TSS (Fig. 8 d-f, D1-6). With the
downwelling motions, the mean current-induced bottom shear stress was enough to drive
resuspension ($\tau_c > \tau_{cr}$). The spatial dynamics of resuspension from down-welling (Fig. 9 c), is
shown for the north shore of the central basin where the thermocline intersects the bottom slope.

(2) Resuspension from upwelling

Thermocline swashing, which was modeled near the northern shoreline of the central basin
(station 2586; Fig. 8 c, f; U1, U2) during storm events (Fig. 2 d, days 250, 259), compressed the
bottom layer. Cold hypolimnion water that was squeezed from this location (Fig. 8 c) upwelled,
after the wind event, under the effect of buoyancy inducing strong $\tau_c$ (Fig. 8 f) and triggering
resuspension. On day 259 (Fig. 9 a), upwelling along the bottom of the northern flank of the
central basin was modeled, corresponding with resuspension in the central basin. This was consistent with both our assumption, that resuspension was caused by upwelling of cold bottom water after a storm, and with observations in Lake Michigan, that high bottom turbidity coincided with periods when the thermocline was elevated (Hawley and Muzzi, 2003).

Interestingly, the north-southward up- and down-welling event only induced resuspension along the north shore, with upwelling events near the surface and downwelling events at depth (Fig. 9, Movie S2). These two mechanisms generated internal swash zones, lifting bottom sediments and forming a turbid intrusion in the metalimnion (Fig. 9a-c). Similar results were observed in Lake Kinneret (Marti and Imberger, 2008). Conversely, the southern flank was relatively quiescent during these events. Bottom sediment was uniform across the lake, and so only differences in fetch, water depth and hydrodynamic forcing may explain these discrepancies. We suspect this pattern results from the predominant southwest winds in summer having a longer fetch to the north shore, which pushes the thermocline down at the windward shore, relative to the rise at the leeward shore (Monismith, 1986), resulting in more intense up- and down-welling events, thus creating stronger bottom currents and turbulence.

(3) Resuspension from Poincaré waves

Wind events produce near-inertial (Poincaré) waves with a near-inertial period of ~17 h in central Lake Erie (Rao et al., 2008; Valipour et al., 2015). Wavelet analysis on modeled temperature at station 1228, showed energy peaks at ~17 h during days 205-240 (Supplementary material, Fig. S7), indicating the existence of Poincaré waves. At station 1228 (Fig. 8b, e, PW1), during the Poincaré wave signals, the baroclinic currents below the thermocline became stronger, by conservation of volume, leading to elevated $\tau_c$ that exceeding $\tau_{cr}$ and resuspended bottom sediments.
Hawley (2004) showed an association between near-inertial waves and the vertical distribution of sediment in Lake Michigan, indicating that internal waves could cause sediment resuspension directly or indirectly. Valipour et al. (2017) examined the ability of the high-frequency internal waves, at the troughs of the Poincaré waves, to resuspend sediment at station 341 in the summer of year 2009. We found the mean currents generated by internal Poincaré waves contributed to the resuspension events. However, resuspension induced by Poincaré waves was less intense, compared to other upwelling and downwelling mechanisms, and resuspension did not persist the entire time when Poincaré waves were energized. Thus, the major effect of these near-inertial internal waves was to maintain the benthic nepheloid layer by keeping particles in suspension close to the bottom, creating the conditions where surface waves and vertical thermocline motion could transport the sediment higher in the water column (Hawley, 2004; Puig et al., 2001). Sediment resuspension induced by thermocline motions has been observed in the field (e.g., Marti and Imberger 2008) and three-dimensional modeling of these processes could help understand biogeochemical cycling at basin-scale.

To investigate spatial distributions of bottom stress among stations, the lake-wide distribution of $\tau_c$ was computed (Eq. 3, 5) and averaged over the simulation in 2008, showing ‘hot spots’ of enhanced mean current-induced resuspension near the Detroit River mouth, the west central basin, the northern shoreline of central basin, and the northwest corner of the eastern basin (Fig. S8). The model was unable to output $\tau_w$ and $\tau_t$ at more than one location per model run, and so plotting distributions of $\tau_w$ and $\tau_t$ was not currently feasible.

d. Importance of river inputs vs. resuspension on the TSS budget

By separating TSS into riverine and lakebed sources, we were able to compare the relative contributions of resuspension (internal loading; $SS_b$) and river inputs (external loading; $SS_R$) to
TSS in the water column (Movie S3). High $SS_R$ (~ 5 mg L\(^{-1}\)) was modeled at the Maumee (Fig. 10 d, e) and Detroit (Fig. 10 a-f) river mouths, indicating that tributary loading was a significant $TSS$ source in the western basin (Binding et al., 2012; Bolsenga and Herdendorf, 1993). Niu et al. (2018) showed that the duration and area of influence, from riverine turbidity, contributed ~90% of turbidity events during summer but was negligible during fall. Our modeled $SS_R$ showed river inputs contributed ~50% (Fig. 10 a-c, g-i) and ~25% (Fig. 10 d-f, j-l) of the near-bed $TSS$ concentration in the western and west central basins during summer and fall, respectively. The differences between these two models may result from different periods of analysis, different model parameters (e.g., critical stress) and/or lower inputs from Detroit River in Niu et al. (2018).

Kemp et al. (1977) reported that the Maumee and Detroit Rivers contributed 12% and 9%, respectively, of the total sediment input to the lake. The Maumee River has a seasonal variation > 2 orders of magnitude (USGS), with the August to October period (present model run) having the lowest discharge and load during the year (Stow et al., 2008). Sediment loading from the Maumee River peaked on day 260 (Fig. S1), but because of the low discharge, only led to a high concentration near the river mouth (Fig. 10 d). The Detroit River discharge does not show significant seasonal variation, with a flow rate 2 orders of magnitude larger than the Maumee River. Thus, during the present simulation (late summer through fall) the highest $SS_R$ concentrations were from the Detroit River, with the strong flow rate spreading $SS_R$ throughout the western basin. Despite having a secondary role in $TSS$ loading during late summer and fall, the Maumee River provides most of the nutrients to the western basin (e.g. Bridgeman et al. (2012); Schwab et al. (2009)).
5. Conclusions

A coupled three-dimensional hydrodynamic and water quality model (ELCOM-CAEDYM) was applied to simulate the background TSS concentration in Lake Erie with average RMSE = 0.74 mg L\(^{-1}\) (in the boundary layer) to 0.81 mg L\(^{-1}\) (in the water column) over 7 stations across the lake. Resuspension events were modeled with \(R^2 = 0.84\) relative to observed acoustic backscatter, and = 0.39 relative to recorded turbidity timeseries at a west central basin station. We were often unable to distinguish phytoplankton from TSS in observed turbidity signals. Due to temporal and spatial averaging over the model timestep and grid, respectively, and limitations in model algorithms, usage of literature-based or field-measured resuspension parameters (e.g., critical stress) is not justified and will likely underestimate resuspension.

The three-dimensional nature of the simulations reproduces the following aspects of sediment resuspension in Lake Erie. Increased TSS concentrations occurred during fall in all three basins, resulting from the increase in resuspension from storm events. The well-known frequent resuspension from surface waves at depths < 20 m, was reproduced by the model. Current-induced resuspension was also modeled to occur in the west central and eastern basins. Thermocline swashing and up- and down-welling events caused resuspension where the thermocline intersected the lakebed in the west central basin and along the north shore of the central. A nepheloid layer was maintained by the currents associated with persistent internal Poincaré wave activity in those regions. Riverine inputs were more significant in summer, compared to fall, when TSS originated from the lakebed during storm events.

The TSS model has several conceptual limitations, including parameterizing the total bottom stress without consideration of wave-current interactions (Eq. 2), fixing the ‘channel’ depth as the height of the bottom grid cell (Eq. 5), and assuming fully rough turbulent flow (Eq. 6).
suitable calibration, these limitations did not have a discernable impact on model performance, although the assumption of fully rough flow led to an underestimation of surface wave-induced stress.

To quantitatively calibrate modeled resuspension during storm events, it would be valuable to obtain TSS observations during storm events. More in situ observations of critical shear stress would enable assessment of the variation in bottom stress within a model grid cell. It is hoped that the results of this study (e.g., maps of bottom stress and near-bed TSS) can be applied to guide future research on the biogeochemistry of Lake Erie and other lakes, such as improved representation of sediments in biogeochemical models of the lake (e.g., Leon et al 2011).
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### 6. Tables

#### Table 1

Details of stations and instrument deployments. See Fig. 1 for a map of the stations.

<table>
<thead>
<tr>
<th>Station</th>
<th>Latitude</th>
<th>Longitude</th>
<th>Depth (m)</th>
<th>Instrument Details</th>
<th>Year</th>
<th>Day of Year</th>
<th>Instrument/Sample Depth (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>341</td>
<td>41°47’29’'</td>
<td>82°16’59’’</td>
<td>17.5</td>
<td>XR-620 (Turbidity) with 1 Hz sampling frequency and 3 min interval</td>
<td>2008</td>
<td>212-288</td>
<td>16</td>
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<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>XR-420 (Turbidity, Fluorescence) with 1 Hz sampling frequency and 3 min interval</td>
<td>2009</td>
<td>118-288</td>
<td>16.5</td>
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<td></td>
<td></td>
<td></td>
<td></td>
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<td>2008</td>
<td>204, 205, 213, 219, 220, 221</td>
<td>12, 15</td>
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<td>1227</td>
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<td></td>
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<td>1228</td>
<td>41°47’53’’</td>
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<td>YSI-6600 Turbidity with 1 Hz sampling frequency and 1 h interval</td>
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<td>155-275</td>
<td>13.5</td>
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<td></td>
<td></td>
<td></td>
<td></td>
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<td>1, 3, 7, 10, 13, 14, 14.5</td>
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<td>1, 5, 8, 12, 15, 18, 19</td>
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<td></td>
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<td>Superficial bed sediment sample (PONAR box core sampler)</td>
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<td>84</td>
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<td>452</td>
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<td>Pumped water sample</td>
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<td>149</td>
<td>Surface (0.5-1)</td>
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Table 2 Comparison of published observations of $d_{50}$, $\tau_{cr}$ and $\alpha$ in Lake Erie and other locations

<table>
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<tr>
<th>Study area</th>
<th>$d_{50}$ (10^{-6} m)</th>
<th>$\tau_{cr}$ (Pa)</th>
<th>Resuspension rate $\alpha$ (g m^{-2} d^{-1})</th>
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<td>Lake Erie[^a]</td>
<td>1, 10</td>
<td>0.01; 0.025</td>
<td>450</td>
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<td>Lake Erie[^b]</td>
<td>10</td>
<td>0.045-0.28*</td>
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<tr>
<td>Lake Erie[^c]</td>
<td>10</td>
<td>0.03-0.3</td>
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<td>Lake Erie[^d]</td>
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<td>Lake Erie[^e]</td>
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<td>0.1</td>
<td>N/A</td>
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<td>Lake Erie[^f]</td>
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<td>0.1-0.2</td>
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<td>Lake Erie[^g]</td>
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<td>30</td>
<td>0.13</td>
<td>N/A</td>
</tr>
<tr>
<td>Lake Michigan[^j]</td>
<td>N/A</td>
<td>0.009-0.134</td>
<td>N/A</td>
</tr>
<tr>
<td>Lake Michigan[^k]</td>
<td>N/A</td>
<td>0.1</td>
<td>N/A</td>
</tr>
<tr>
<td>Lake Ontario[^l]</td>
<td>N/A</td>
<td>0.03-0.06</td>
<td>N/A</td>
</tr>
<tr>
<td>Lake St. Clair[^m]</td>
<td>N/A</td>
<td>0.25</td>
<td>N/A</td>
</tr>
</tbody>
</table>

[^k](Hawley et al., 2004).  [^l](Hawley et al., 1996).  [^m](Tsai and Lick, 1986).

* $\tau_{cr}$ was modified based on a correlation between instantaneous burst currents and burst-average currents (see text for further explanation).*
Table 3 Simulated parameter combinations, with RMSE between modeled TSS concentration in bottom layer and observed TSS from water samples near-bed collected during days 205-221 at station 357, 1227, 1228, 341, 1231, 84, 908 and 452, and association (R²) between modeled TSS concentration in bottom layer and ADV-amp and turbidity data at station 341.

<table>
<thead>
<tr>
<th>Combination</th>
<th>$d_{50}$ (μm)</th>
<th>$\tau_r$ (Pa)</th>
<th>$\alpha$ (g m⁻² d⁻¹)</th>
<th>RMSE in bottom layer (mg L⁻¹)</th>
<th>RMSE in overall water column (mg L⁻¹)</th>
<th>$R^2$ with ADV-amp</th>
<th>$R^2$ with turbidity</th>
</tr>
</thead>
<tbody>
<tr>
<td>(A)</td>
<td>1 (SSB₁); 10(SSB₂)</td>
<td>0.045 (SSB₁); 0.045 (SSB₂)</td>
<td>450</td>
<td>0.74</td>
<td>0.63</td>
<td>0.51</td>
<td>0.06</td>
</tr>
<tr>
<td>(B)</td>
<td>1 (SSB₁); 10(SSB₂)</td>
<td>0.03 (SSB₁); 0.045 (SSB₂)</td>
<td>450</td>
<td>0.74</td>
<td>0.64</td>
<td>0.72</td>
<td>0.16</td>
</tr>
<tr>
<td>(C)</td>
<td>1 (SSB₁); 10(SSB₂)</td>
<td>0.01 (SSB₁); 0.025 (SSB₂)</td>
<td>450</td>
<td>0.74</td>
<td>0.81</td>
<td>0.84</td>
<td>0.39</td>
</tr>
<tr>
<td>(D)</td>
<td>1 (SSB₁); 10(SSB₂)</td>
<td>0.01 (SSB₁); 0.025 (SSB₂)</td>
<td>300</td>
<td>0.74</td>
<td>0.73</td>
<td>0.83</td>
<td>0.25</td>
</tr>
<tr>
<td>(E)</td>
<td>1 (SSB₁); 10(SSB₂)</td>
<td>0.03 (SSB₁); 0.045 (SSB₂)</td>
<td>300</td>
<td>0.74</td>
<td>0.63</td>
<td>0.68</td>
<td>0.15</td>
</tr>
</tbody>
</table>

In all the combinations, $K_T = 3 \times 10^{10}$ g m⁻¹ and $\rho_s = 2650$ kg m⁻³.
Table 4 Contributions of mean currents and surface waves to sediment resuspension at station 341 during days 202-303 of year 2008, and days 118-280 of year 2009, and station 591 during days 250 – 310 of year 2013. The averaged ratio of $\tau_w$ and $\tau_c$ over $\tau_t$ when $\tau_t > \tau_{cr} = 0.025$Pa were used to calculate $\tau_w: \tau_t$ and $\tau_c: \tau_t$. $T_r$ at 341 is ~ (102 +182 = 284) days in total (years 2008 and 2009), and at station 591 in year 2013 is ~ 59 days.

<table>
<thead>
<tr>
<th></th>
<th>Station 341 (17.5 m)</th>
<th>Station 591 (14.8 m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>$\tau_c: \tau_t$</td>
<td>42%</td>
<td>85%</td>
</tr>
<tr>
<td>$\tau_w: \tau_t$</td>
<td>58%</td>
<td>15%</td>
</tr>
<tr>
<td>$T_r$ ($T_r: T_t$)</td>
<td>36.3 days (12.8%)</td>
<td>6.9 days (12%)</td>
</tr>
</tbody>
</table>
7. Figure Captions

Fig. 1 Bathymetry of Lake Erie. Field stations and model output (M2586) are indicated with red dots. Model output curtains are indicated with blue lines through the western and central basin (see Fig. 9 and 11). Depth contours are in meters.

Fig. 2 Time series of (a, b) observed current velocities from the ADV deployed 1 mab (meter above bottom) and modeled current velocities 1 mab (blue lines) at station 341 (red lines) in year 2008. Observed current velocities were 5-min averaged. Time series of (c) observed wind speed (black line) and wind direction (red dashed line), (d) observed and modeled wave height, and (e) observed and modeled wave period in year 2008. Observed wind and wave properties were 10-min averaged from buoy 45005 (NDBC); whereas modeled wave height and period were at station 341.

Fig. 3 Time series at station 341 of (a-e) modeled TSS at 1 mab for parameter combinations (A-E) in Table 3, and ADV-amp 1 mab (blue dash lines). Orange circles are the observed TSS concentration from pumped water samples. Time-series of (f) turbidity data from XR 620 (1 mab), XR 420 (5 mab), and chlorophyll-a concentration from XR 620 at station 341. Black arrows in (c) indicate the identified resuspension events.

Fig. 4 Time series of turbidity data from YSI 6600 sampled at 1 mab and modeled TSS at 1 mab of stations (a) 1228, (b) 591, (c) 67, and (d) 475. Time series of (e) ADV-amp and modeled TSS at 1 mab of station 341, and (f) turbidity data from XR 620 and chlorophyll-a concentration from XR 620 at station 341. Black arrows in (a, b, e) indicate the identified resuspension events.

Fig. 5 Time series of modeled temperature and TSS concentration profiles in the western (station 357), west-central (station 341), central (station 84) and eastern (station 452) basins. Note the change in scale between (b), (d) and (f), (h). The data missing in the surface layer of (a – f) is due to the change of water level.

Fig. 6 Time series of theoretical $\tau_w$ (Eq. 2, 4, 7), modeled $\tau_w$ (Eq. 2, 4, 6) for $SS_B2$ ($d_{50} = 10 \mu m$) at station 341 (-17.5 m) in years (a) 2008, (b) 2009, and (c) station 591 (-14.8 m) in year 2013. The green dashed lines indicate $\tau_{cr} = 0.025$ Pa in the model.

Fig. 7 Time series of modeled $\tau_c$ (current-induced bottom shear stress, Eq. 3, 5; red lines), and $\tau_w$ (surface wave-induced bottom shear stress, Eq. 4, 6; blue lines) for $SS_B2$ ($d_{50} = 10 \mu m$) at station 341 (~17.5 m) in years (a) 2008, (b) 2009, and (c) station 591 in year 2013. Black arrows indicate the identified resuspension events shown in Fig. 3, and 4.

Fig. 8 Panels (a-c) are time series of modeled temperature profiles at station 1227, 1228, and 2586; (b-f) are corresponding time series of modeled TSS concentration in the bottom layer, $\tau_c$ (current-induced bottom shear stress), and $\tau_{cr}$ (critical shear stress). Black rectangles indicate down-welling (D1-6), up-welling (U1-2), and Poincaré waves (PW1).

Fig. 9 Model output central basin curtain (see Fig. 1) showing (a) up-welling (day 259, U1), (b) day 266 and (c) down-welling (day 271, D6) along the north shore of the central basin. The color bar shows sediment concentration, and the black lines are isotherms contours through the metalimnion from 14°C (bottom) and increasing at 2°C intervals. (d) shows $\tau_c$ in these three days along the bottom of this curtain.

Fig. 10 Model output of $SS_B$ and TSS in the bottom layer on days (a, g) 224; (b, h) day 240; (c, i) day 250; (d, j) day 260; (e, k) day 290; (f, l) day 303. The dates were chosen based on resuspension events in Fig. 3a, c and storm events in Fig. 2d. The black lines are isotherms contours increasing from 16 °C at 2 °C intervals toward the lake perimeter in (g-k). These indicate where the thermocline intersected the lakebed; (k) only shows the 16 °C and contour, and the lake was colder than 16 °C in (l).
Figure 1
Figure 2
Figure 3

(a) $\alpha = 450 \text{ g m}^{-2} \text{ d}^{-1}$; $\tau_{\alpha} = 0.045 \text{ 0.045 Pa}$

(b) $\alpha = 450 \text{ g m}^{-2} \text{ d}^{-1}$; $\tau_{\alpha} = 0.03 \text{ 0.045 Pa}$

(c) $\alpha = 450 \text{ g m}^{-2} \text{ d}^{-1}$; $\tau_{\alpha} = 0.01 \text{ 0.025 Pa}$

(d) $\alpha = 300 \text{ g m}^{-2} \text{ d}^{-1}$; $\tau_{\alpha} = 0.01 \text{ 0.025 Pa}$

(e) $\alpha = 300 \text{ g m}^{-2} \text{ d}^{-1}$; $\tau_{\alpha} = 0.03 \text{ 0.045 Pa}$

(f) Chla 1.5 mab XR 420-5 mab XR 820-1 mab
Figure 5

(a) Station 357 Temperature
(b) Station 357 TSS
(c) Station 341 Temperature
(d) Station 341 TSS
(e) Station 84 Temperature
(f) Station 84 TSS
(g) Station 452 Temperature
(h) Station 452 TSS
Figure 6

(a) Sta. 341
- Theoretical $\tau_w$
- Modeled $\tau_w$
RMSE = 0.0139 Pa

(b) Sta. 34F
RMSE = 0.0142 Pa

(c) Sta. 591
RMSE = 0.0256 Pa
Figure 7
Figure 8

![Figure 8](image-url)
Figure 9

(a) 

(b) 

(c) 

(d) 

Sonthing (km)

TSS (mg L⁻¹)

Depth (m)

τ (Pa)

Day 259 — Day 266 — Day 271 — τ_{cr}
Figure 10