TURBULENCE AND SEDIMENT RESUSPENSION MODELLING IN
LAKE ERIE

by

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Abstract

Physical and biogeochemical observation in Lake Erie, and a field-scale numerical model was applied to study the turbulent dissipation and mixing, sediment resuspension and bottom stress parameterizations. In Chapter 2, temperature microstructure data were processed and grouped to represent characteristic turbulent kinematic energy (TKE) dissipation rate ($\varepsilon$) and turbulent diffusivity ($K_z$) profiles in distinct regions of Lake Erie. The presence of a seasonal thermocline can increase $\varepsilon$ up to $10^{-6}$ (m$^2$ s$^{-3}$) and decrease $K_z$ to molecular diffusivity in the bottom layer of the western basin and metalimnion of the central and eastern basins. The TKE budget shows efficiency of turbulent energy transfer from the wind decreased with increasing wind stress, and 36 - 54% of energy transferred from wind dissipated beneath the surface mixed layer, which was larger than that observed in smaller lakes. In Chapter 3, a three-dimensional Reynolds-averaged Navier Stokes (RANS) equation model, coupled with a water quality model, was applied to simulate sediment resuspension in Lake Erie. The model was qualitatively validated and able to reproduce the timing of resuspension. The model showed surface wave-induced bottom shear stress dominated sediment resuspension in shallow western basin, and with contributions from up- and down-welling events, and internal Poincaré wave motions, the importance of mean current-induced bottom shear stress increased in central basin. Resuspension in the deeper regions (>25 m) of the central and eastern basins was not modelled. To improve the prediction of sediment resuspension in field and model, bottom shear stress parameterizations, including surface wave stress method ($\tau_w$), quadratic stress method ($\tau_c$), log-law method ($\tau_L$) and turbulent kinetic energy method ($\tau_{TKE}$), were assessed in Chapter 4. In 2008-09 observations, resuspension induced by surface waves, bottom mean currents, and combination of them could be predicted by total bottom shear stress ($\tau_b$), represented by $\tau_w + \tau_c + \tau_L$, and $\tau_{TKE}$. Considering required model
setups (computing power, grid, etc.), \( \tau_b = \tau_w + \tau_c \) is still the most practical parameterization method within field-scale RANS models. In comparison to observed values based on the same parameterization, the optimal algorithms of \( \tau_w \) and \( \tau_c \) in the RANS models showed 0.031 Pa and 0.025 Pa root-mean-square-error (RMSE), respectively.
Co-Authorship

Significant contributions have been made by Leon Boegman and Yingming Zhao by direct supervision of work, interpreting the results and commenting on the original work of Shuqi Lin for journal preparation. In the thesis, “we” refers to the co-authors in each Chapter.

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# List of Abbreviations

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<tr>
<td>ADCP</td>
<td>Acoustic Doppler current profiler</td>
</tr>
<tr>
<td>ADV</td>
<td>Acoustic Doppler velocimeter</td>
</tr>
<tr>
<td>AR</td>
<td>Autoregressive model</td>
</tr>
<tr>
<td>BBL</td>
<td>Bottom boundary layer</td>
</tr>
<tr>
<td>BNL</td>
<td>Bottom nepheloid layer</td>
</tr>
<tr>
<td>ELCD</td>
<td>ELCOM-CAEDYM</td>
</tr>
<tr>
<td>IDM</td>
<td>Inertial dissipation method</td>
</tr>
<tr>
<td>ISW</td>
<td>Internal solitary waves</td>
</tr>
<tr>
<td>NLIW</td>
<td>Nonlinear internal waves</td>
</tr>
<tr>
<td>RANS</td>
<td>Reynolds-averaged Navier-Stokes</td>
</tr>
<tr>
<td>RMSE</td>
<td>Root-mean-square-error</td>
</tr>
<tr>
<td>SFM</td>
<td>Structure function method</td>
</tr>
<tr>
<td>TKE</td>
<td>Turbulent kinematic energy</td>
</tr>
<tr>
<td>TSS</td>
<td>Total suspended solid</td>
</tr>
<tr>
<td>SOD</td>
<td>Sediment oxygen demand</td>
</tr>
</tbody>
</table>
Roman symbols

\( a \)  
Maximal wave amplitude

\( A \)  
Surface area of lake

\( C_d \)  
Drag coefficient at air-water interface

\( C_D \)  
Drag coefficient at sediment-water interface

\( C_{D,1m} \)  
Drag coefficient 1 m above bottom

\( C_{D,ref} \)  
Drag coefficient at reference height above bottom

\( d_{50} \)  
Median grain size of sediment

\( D_s \)  
Dimensionless particle diameter

\( D \)  
Molecular diffusivity

\( f_c \)  
Current friction factor

\( f_w \)  
Wave friction factor

\( g \)  
Gravitational acceleration

\( h_s \)  
Surface layer depth

\( h \)  
Water depth

\( H_s \)  
Significant wave height

\( k \)  
Wavenumber

\( k_B \)  
Batchelor cutoff wavenumber

\( k_s \)  
Bed roughness

\( K_Z \)  
Turbulent diffusivity

\( l \)  
Characteristic length scale of lake

\( L \)  
Wave length

\( N \)  
Buoyancy frequency
<table>
<thead>
<tr>
<th>Symbol</th>
<th>Definition</th>
</tr>
</thead>
<tbody>
<tr>
<td>$P_{10}$</td>
<td>Wind energy flux</td>
</tr>
<tr>
<td>$P_{bias}$</td>
<td>Percentage bias</td>
</tr>
<tr>
<td>$Pr$</td>
<td>Prandtl number</td>
</tr>
<tr>
<td>$R$</td>
<td>Pearson correlation coefficient</td>
</tr>
<tr>
<td>$Re$</td>
<td>Reynolds number</td>
</tr>
<tr>
<td>$Ri$</td>
<td>Gradient Richardson number</td>
</tr>
<tr>
<td>$Re_b$</td>
<td>Buoyancy Reynolds number</td>
</tr>
<tr>
<td>$S^2$</td>
<td>Vertical current shear</td>
</tr>
<tr>
<td>$S(k)$</td>
<td>Batchelor spectrum (in Chapter 2)</td>
</tr>
<tr>
<td>$SS_B$</td>
<td>Suspended solid from lake bed</td>
</tr>
<tr>
<td>$SS_R$</td>
<td>Suspended solid from river</td>
</tr>
<tr>
<td>$SS_{sed}$</td>
<td>Mass of all superficial sediments at bottom (in Chapter 3)</td>
</tr>
<tr>
<td>$T'$</td>
<td>Temperature fluctuations</td>
</tr>
<tr>
<td>$T_t$</td>
<td>Total model time</td>
</tr>
<tr>
<td>$T_R$</td>
<td>Total duration of resuspension events (the time when $\tau_b &gt; \tau_{cr}$)</td>
</tr>
<tr>
<td>$T_s$</td>
<td>Significant wave period</td>
</tr>
<tr>
<td>$u$</td>
<td>Instantaneous horizontal velocity</td>
</tr>
<tr>
<td>$u_*$</td>
<td>Friction velocity at bottom</td>
</tr>
<tr>
<td>$u_{a*}$</td>
<td>Friction velocity at air-water interface</td>
</tr>
<tr>
<td>$u_{*c}$</td>
<td>Current-induced friction velocity</td>
</tr>
<tr>
<td>$u_{*w}$</td>
<td>Wave-induced friction velocity</td>
</tr>
<tr>
<td>$u(z)$</td>
<td>Velocity profile</td>
</tr>
<tr>
<td>$u_{max}$</td>
<td>Maximal instantaneous burst velocity (in Chapter 3)</td>
</tr>
</tbody>
</table>
\( u_{mean} \) 5 min-averaged mean flow velocity (in Chapter 3)
\( u' \) Turbulent horizontal velocity
\( U \) Mean horizontal current velocity
\( U_{bot} \) Modeled current speed in the bottom layer
\( U_{cr} \) Critical value for time-averaged current velocity
\( U_{orb} \) Maximal wave orbital velocity
\( U_x \) eastward component of mean current velocity (in Chapter 2, 3)
\( U_y \) northward component of mean current velocity (in Chapter 2, 3)
\( U_z \) vertical component of mean current velocity (in Chapter 3)
\( U_{1m} \) Mean current velocity 1 m above bottom
\( v_s \) Settling velocity of particles
\( w \) Instantaneous vertical velocity
\( w' \) Turbulent vertical velocity
\( W_{10} \) Wind speeds 10-m height
\( z_0 \) Roughness height of lake bed (i. e. zero velocity level)

**Greek symbols**

\( \alpha \) Resuspension rate
\( \Gamma \) Mixing efficiency
\( \varepsilon \) TKE dissipation rate
\( \varepsilon_B \) TKE dissipation rate from Batchelor fitting
\( \theta \) Shield parameter
\( \theta_{cr} \) Critical value of Shield parameter
\( \kappa \)  
von Karman constant \((=0.4)\)

\( \kappa_x \)  
eastward component of eddy diffusivity

\( \kappa_y \)  
northward component of eddy diffusivity

\( \kappa_z \)  
vertical component of eddy diffusivity

\( \nu \)  
Kinematic viscosity

\( \rho \)  
Density of fluid

\( \rho_a \)  
Density of air

\( \rho_h \)  
Density of hypolimnetic water

\( \rho_s \)  
Density of particles

\( \rho_w \)  
Density of water

\( \rho_0 \)  
Reference density

\( \tau_b \)  
Bottom shear stress

\( \tau_c \)  
Bottom current-induced stress (quadratic stress)

\( \tau_{cr} \)  
Critical shear stress

\( \tau_L \)  
Bottom stress estimated by log-law method

\( \tau_{max} \)  
Maximal shear stress (in Chapter 3)

\( \tau_{ref} \)  
Reference shear stress \((= 1 \text{ Pa})\)

\( \overline{\tau_R} \)  
Reynolds stress

\( \tau_{TKE} \)  
Bottom stress estimated by turbulent kinematic energy method

\( \tau_v \)  
Viscous stress

\( \tau_w \)  
Surface wave-induced stress

\( \chi_\theta \)  
Temperature variance dissipation rate
Chapter 1

Introduction

1.1 Background and motivation

1.1.1 Study area and current situation

The Laurentian Great Lakes are one of the world’s most important inland water system, and collectively contain around one fifth of the world’s fresh surface water. As the source of fresh water for urban, agricultural, industrial and recreational purposes, the Great Lakes require sustainable lake management, and deterioration of the water quality in the Great Lakes has been the focus of study in recent decades.

Located on the Canada-US border, Lake Erie is the shallowest of the Great Lakes. One-third of the population living in the Great Lakes catchment area is distributed within the watershed of Lake Erie. Thus, water quality in Lake Erie has received great attention. As reported, Lake Erie has experienced harmful algal blooms, hypoxia, and fish kills reflecting the effects of significant eutrophication induced by anthropogenic activities since the middle of last century (Rosa and Burns, 1987; Hawley et al., 2006; Rao et al., 2008; Stumpf et al., 2012; Rao et al., 2014). Studies about biogeochemical processes and relative hydrodynamic processes are fundamental for lake management. Also, for comprehensive understanding of internal mechanisms and target-oriented management, the research on Lake Erie, which is a complex multi-basin lake system, should extend to lake-scale using multiple observation sites and computer models.

1.1.2 Research focus

1.1.2.1 Small scale hydrodynamic process
In natural water systems, hydrodynamic and biogeochemical processes are tightly coupled. Turbulence, as a small-scale hydrodynamic process that drives diffusion (Imberger, 1998; Kantha and Clayson, 2000), has long been the focus of research. Although the large-scale motions are mainly horizontal, turbulent eddies are associated with random velocity fluctuations in all three directions (Pope, 2000). Thus, they tend to make a significant contribution to vertical diffusivity as well as sediment resuspension, regulating biogeochemical processes in the water column (Ackerman et al., 2001; Lorke and MacIntyre, 2009; Rao et al., 2008). The spatiotemporal change of turbulence intensity can affect the mass transport in water systems to a large extent. The bottom boundary layer (BBL) forming above the sediment surface is a hotspot for transformation of large-scale motions to turbulence and dissipation of turbulent kinetic energy (TKE) (Imberger, 1998; Lorke and MacIntyre, 2009), and the active turbulent vortex motion within BBL, which provides sufficient near-bed vertical velocity, is one of the major triggers of sediment resuspension and contributes to sediment transport (Figure 1.1; Aghsae and Boegman, 2015; Salim et al., 2018; Soulsby, 1983; Yuan et al., 2009). Therefore, characterizing turbulence intensity can help with parameterizing sediment dynamics across the sediment-water interface (Kim et al., 2000; Biron et al., 2004; Bluteau et al., 2016; Zulberti et al., 2018). Given that sediment plays an important role in elemental cycling in the aquatic environment, turbulence potentially modifies the biogeochemical environment of BBL and even the whole water column by inducing sediment resuspension.
Figure 1.1 Sketches showing the typical distributions of turbulent eddies in a medium sized lake and turbulent bursting resuspension mechanisms (from Wuest and Lorke, 2003 and Boegman and Ivey, 2009).

The intensity of turbulence is often expressed by the rate of dissipation of turbulent kinetic energy, $\varepsilon$ [m$^2$s$^{-3}$], the rate by which energy is transferred from the energy containing eddies and eventually dissipated by viscosity (Pope, 2000). Turbulent mixing (or turbulent diffusivity) $K_Z$ [m$^2$s$^{-1}$], which is the diffusion of scalar variables by turbulent eddies, is found to be well-correlated with $\varepsilon$. It is difficult to directly measure $\varepsilon$ due to instantaneous velocity-gradients. Thus, indirect methods have been developed, including Batchelor fitting, (e.g., McCune, 1998; Ruddick et al., 2000; Bouffard and Boegman, 2013), the structure function method, SFM, (e.g., Wiles et al., 2006; Jabbari et al., 2016) and the inertial dissipation method, IDM (e.g., Jabbari et al., 2015). The SFM and IDM are typically applied to time-series data from moored instruments at individual locations; whereas Batchelor fitting is applied to microstructure profile data, which can be collected at multiple sites in a season, providing information on local vertical distributions of $\varepsilon$. As the ratio between buoyancy flux and the vertical density gradient (Osborn, 1980), $K_Z$ is achievable when buoyancy flux is accurately measured, which remains difficult. Thus, a
parameterization of $K_Z$ was developed as a function of buoyancy frequency $N$, kinematic viscosity $\nu$, and $\epsilon$.

Turbulence observations in lakes and oceans showed that $\epsilon$ and $K_Z$ are distinctly different in the surface boundary layer, interior of the stratified water, bottom boundary layer, nearshore and offshore areas (Kantha and Clayson, 2000; Wuest and Lorke, 2003; Bouffard et al., 2012). A review paper (Wuest and Lorke, 2003) characterized the three distinct water bodies: the energetic surface boundary layer, the slightly less turbulent bottom boundary layer, and the strongly stratified and calm interior in a medium-sized lake (Figure 1.1). However, the observations of turbulence levels in large lakes are limited and sporadic, for example, in the Great Lakes, turbulence observations focused more on the thermocline region during seasonal stratification where internal waves (e.g. Poincaré waves) occurred (e.g., Bouffard et al., 2012; Choi et al., 2012).

In Lake Erie, annual heat budgets (Schertzer et al., 1987) and large scale hydrodynamic motions, (e.g., mean circulation patterns (Beletsky et al., 1999), near-inertial internal waves with a period ~17h (Valipour et al., 2015) have been well investigated and characterized at basin-scale. But the research associated with mapping and characterizing the small scale hydrodynamic motions like turbulence levels and vertical mixing, which would benefit research on biogeochemical fluxes in Lake Erie, remains incomplete.

1.1.2.2 Sediment resuspension

Sediment resuspension affects water quality in natural water systems by entraining seabed organic matter into the water column, which can increase remineralization (Aller and Aller, 1998), alter oxygen and nutrient dynamics (Rao et al., 2008; Almroth et al., 2009; Moriarty et al., 2018), and decrease water clarity (Tilzer, 1983; Fréchette et al., 1989; Gloor et al., 1994). High
suspended solid concentrations, induced by resuspension, may influence fish breeding success, egg and larval survival, as well as food availability and feeding efficiency (Bruton, 1985). Resuspension of organic biomass can increase the sediment oxygen demand (SOD) by enhancing the surface area of decaying organic matter (Ackerman et al., 2001; Lorke and MacIntyre, 2009), thereby exacerbating hypoxia, leading to internal loading of phosphorus and consequent nuisance algae blooms (Matisoff et al., 2016; Paytan et al., 2017). Therefore, understanding resuspension mechanisms, predicting and quantifying sediment resuspension events are of great importance in water system management.

Sediment resuspension occurs when the applied total bottom shear stress, $\tau_b$, is greater than the critical shear stress (Van Rijn, 1993). Resuspension of non-cohesive particles is parameterized according to the Shields parameter $\theta$, which is the ratio of total bottom stress to the submerged particle weight (Van Rijn, 1993):

$$\theta = \frac{\rho u_*^2}{(\rho_s - \rho)gd_{50}}$$

(1.1)

where $u_*$ is the friction velocity, $\rho_s$ and $\rho$ are the density of particles and fluid, respectively, $d_{50}$ is the particle diameter, $g$ is acceleration due to gravity. Laboratory-based Shield diagrams give the critical value $\theta_{cr}$ as a function of $d_{50}$, indicating the initiation of sediment motion. However, the existing shields diagram has not been extended to give $\theta_{cr}$ for $d_{50} < \sim 40 \mu m$. Soulsby and Whitehouse (1997) extended the work of Shields by fitting a simple analytical formula to the traditional hand-drawn Shields curve for critical shear stress, and applying a correction to make the curve fit for very fine grain sizes.

$$\theta_{cr} = \frac{0.30}{1 + 1.2D_e} + 0.055(1 - e^{-0.02D_e})$$

(1.2)
\[ D_* = g \left( \frac{\rho_s - \rho}{\rho \nu^2} \right)^{1/3} d_{50} \]  
\[ \tau_{cr} = \theta_{cr} g (\rho_s - \rho) d_{50} \]  

where \( D_* \) is dimensionless particle diameter and \( \nu \) is the kinematic viscosity.

Different from laboratory environment, realistic conditions in natural water systems not only have complex hydrodynamic forcing, but contain a wide range of sediment sizes. In the Great Lakes, researches have shown that resuspension was initiated by the combined effects of currents, surface waves and high-frequency internal waves, among which the surfaces waves predominate in shallow regions (e.g., Lou et al., 2000; Hawley et al., 2004; Hawley and Eadie, 2007; Valipour et al., 2017).

In Lake Erie, which is the shallowest of the Laurentian Great Lakes, the hydrodynamic energy is higher than the other Great Lakes (Thomas et al., 1976). In general, the shallowness of Lake Erie, especially the western basin and nearshore areas, makes its lakebed very susceptible to storm events (Bedford and Abdelrhman, 1987; Sheng and Lick, 1979; Lick et al., 1994; Valipour et al., 2017). Haltuch et al. (2000) mapped the composition of lakebed sediment; mud (<0.063μm), which is fine enough to be resuspended, was found to be the prevalent bottom substrate in all three basins, such that resuspension can be expected. Western and central basins of Lake Erie, which have flatter topography, are more susceptible to heightened storm conditions and lake water level variations than the deep and bowl-shaped eastern basin. Resuspension events resuspend and deposit the same material over and over again with little net accumulation or erosion in central basin, while in eastern basin, the deposition of advected material is more continuous, with little resuspension taking place (Lick et al., 1994; Hawley and Eadie, 2007).
Although a lot of effort has been put into observations of sediment resuspension in lakes, *in situ* observational approaches are often limited by technological and/or cost constraints, and so only the observations from limited sites and periods are available (e.g. Hawley et al., 1996; Hawley and Lee, 1999; Hawley and Eadie, 2007; Valipour et al., 2017). Given the complicated morphology of natural water systems, like Lake Erie, a comprehensive understanding of basin-scale sediment resuspension and the driving mechanisms, dynamics remains elusive. Recently, increasing concern over water quality of Great Lakes has stimulated interest in applying numerical sediment models to address the aquatic environmental issues, which are more efficient for interpolating and extrapolating results in space and time.

Coupled with hydrodynamic models and/or wind wave models, the sediment resuspension models, based on surface waves (Hawley and Lesht, 1992; Hawley et al., 2004; Dusini et al., 2009), nonlinear wave-current interactions (Lou et al., 2000), and the combined effect of surface waves and currents (Lick et al., 1994; Cardenas et al., 2005; Niu et al., 2018) were developed and validated against the observations in Lake St. Clair, Lake Michigan and Lake Erie. Although these models varied from zero to three dimensions and simulated uniform or multiple sediment classes, most of models simulated the resuspension according to difference between bottom stress and critical shear stress, sediment grain size, and resuspension rate (or so called ‘entrainment rate’). A key factor in deciding whether resuspension occurs or not, bottom shear stress can be parameterized in different ways, and in most of cases, considering the temporal resolution of most field-scale numerical models, it is correlated with surface wave orbitals and mean current velocities. A resulting issue is that setting the values of time-averaged critical shear stress becomes little more than guesswork with few measurements available. Given the wide sediment diameter range and complex flow conditions in a real lake, the threshold provided
by the Shield diagram is not as useful as expected. Thus, a better parameterization of bottom shea
stress and a more accurate prediction of sediment resuspension remains to be explored.

1.1.2.3 Bottom shear stress estimates and parameterization

In a steady isotropic turbulent bottom boundary layer (BBL), $\tau_b$ over a flat bottom is the sum of the viscous stress ($\tau_v$) and Reynolds stress ($\overline{\tau_R}$). In most natural water systems, the Reynolds number ($\textit{Re}$) is sufficiently high to sustain a turbulent boundary layer above the sediment surface. Under such conditions, the $\tau_v$ may be negligible, and $\overline{\tau_R}$ dominate the shear stress computation. However, $\overline{\tau_R}$ is computed by turbulent velocity fluctuations, which usually cannot be resolved by most field-scale numerical models (based on Reynolds-averaged Navier-Stokes [RANS] equations) given the available computational power. It can be assumed that $\tau_b$ is constant within the boundary layer (constant stress layer), and is related to the mean current velocity at a certain height above sediment by the drag coefficient $C_D$ (so-called quadratic stress law) (Johns, 1983).

$$\tau_b = \rho C_D U^2$$ (1.5)

Within the boundary layer, velocity distribution $u(z)$ follows the “law of the wall” (van Rijn, 1990),

$$u(z) = \frac{u_* \ln \frac{z}{z_0}}{\kappa}$$ (1.6)

where $\kappa \approx 0.4$ is von Karman’s constant, $z_0$ is the roughness length, which is often related to sediment diameter $d_{50}$, and $u_*$ is friction velocity, which is a turbulent velocity scale defined by $\tau_b$,

$$u_* = \sqrt{\frac{\tau_b}{\rho}}$$ (1.7)
Thus, $\tau_b$ can be modeled based on Eq. (1.5) if a model can simulate the mean current velocity 1 m above bottom ($U_{1m}$), and the drag coefficient 1 m above bottom can be derived as (Johns, 1983),

$$C_{D,1m} = \left( \frac{\kappa}{\ln(1m) - \ln(z_0)} \right)^2$$  \hspace{1cm} (1.8) 

A more reliable way is to determine $u_*$ by fitting the profiles to Eq. (1.6), if velocity profiles $u(z)$ are available, and get $\tau_b$ by Eq. (1.7).

The simplified parameterizations described above are suitable for current-dominated environments, and surface waves-induced stress $\tau_w$ and/or stress induced by current-wave interaction should be added, depending on the existence of surface wave forcing (e.g., Lou et al., 2000; Hawley et al., 2004). The assumption of constant stress and choice of drag coefficients needs to be modified depending on hydrodynamics, presence of bedforms and sediment properties (Soulsby, 1997; Li and Amos, 2001). In the existing field-scale numerical models, the quadratic stress-based (with or without $\tau_w$) parameterization of $\tau_b$ is usually applied with the concept that resuspension occurs when $\tau_b$ exceeds time-averaged threshold $\tau_{cr}$ (van Rijn, 1993), and this concept has been tested in multiple aquatic systems through field-scale numerical simulations (e.g., Blaas et al., 2007; Warner et al., 2008; Hu et al., 2009; Morales-Marin et al., 2018; Niu et al., 2018). However, there are two issues within this parameterization and resuspension concept.

One is the accuracy of computed $C_D$ in models, which can determine the magnitude of quadratic stress and the occurrence of resuspension. Another issue is whether a time-averaged threshold concept can predict sediment resuspension accurately in complex water systems. Recent laboratory experiments and field observations suggest that near-bed intermittent turbulent eddies, which are likely to be filtered out by time and space-averaging in RANS models, control
sediment resuspension rather a single time-averaged critical velocity (e.g., Yuan et al., 2009; Aghsaee and Boegman, 2015; Yang et al., 2016; Salim et al., 2018). Furthermore, the quadratic stress law is also an inappropriate parameterization for predicting resuspension or high TSS in BBL generated by nonlinear internal waves (NLIW) (Quaresma et al., 2007), and vortex shedding beneath internal solitary waves (ISW) (Aghsaee and Boegman, 2015).

Effort has focused on research that assess τb parameterizations based on turbulence measurements in simple and complex flows under laboratory conditions (Biron et al., 2004), river estuaries (Kim et al., 2000), and the continental shelf (Bluteau et al., 2016; Zulberti et al., 2018), and the adaptability of these τb parameterizations may be variable because of heterogeneity in resuspension mechanisms and topographic features. In the Great Lakes, such assessment of τb parameterizations based on in situ mean current and turbulence measurement has not occurred. Although the bottom mean currents and surface waves have been shown to be the major mechanisms driving sediment resuspension in the Great Lakes, shear instability induced by high-frequency internal waves (HFIWs) carried by basin-scale internal waves, which are a predominant hydrodynamic feature in these large lakes, may also induce sediment resuspension or maintain the turbid BBL (Valipour, 2012; Valipour et al., 2017).

To better predict sediment resuspension in these large lakes, τb parameterizations based on observations and models should be assessed, and the algorithms of existing τb parameterizations in RANS models should be validated against observed values so that the accuracy of the model results is known and required model setups (computing power, grid, etc.) are acceptable.

1.1.3 Objectives of thesis
The specific objectives of this thesis are:
1. To map the spatiotemporal distributions of turbulence characteristics ($\varepsilon$ and $K_Z$) in Lake Erie and investigate the effects of typical hydrodynamic processes (e.g., storms, near-inertial waves, upwelling, etc.) on the variability of $\varepsilon$ and $K_Z$.

2. To validate the ability of a well-used RANS model to simulate sediment resuspension in Lake Erie and extrapolate model results from field sites to understand the hydrodynamic processes that control sediment dynamics at the basin-scale.

3. To improve parameterizations for sediment resuspension, from both observations and RANS models, by assessing methods to compute $\tau_b$ from in situ observations and within a computational framework. Explore practical and optimal improvements in $\tau_b$ parameterizations for various sediment resuspension mechanisms.

1.2 Layout of the thesis

In Chapter 2, over 600 temperature microstructure casts from multiple stations across Lake Erie were processed and grouped to map typical values of $\varepsilon$ and $K_Z$ in different regions. Hydrodynamic and meteorological data were used to study significant variations of $\varepsilon$ and $K_Z$ during hydrodynamic events (e.g., storms, Poincaré waves, entrainment, etc). TKE budget and $K_Z$ within Lake Erie were also compared with those in other lakes. In Chapter 3, the three-dimensional RANS equation model ELCOM, coupled with water quality model CAEDYM, was validated qualitatively against in situ turbidity and backscatter signal intensity data for simulating basin-scale sediment resuspension in Lake Erie. The contribution of surface wave-induced and mean current-induced bottom stress to resuspension events was compared in different basins. The hotspots of sediment resuspension were revealed in the model results, suggesting the regions that potentially have enhanced sediment-water-interface nutrient exchange and sediment oxygen demand. In Chapter 4, four $\tau_b$ parameterizations, including the surface wave stress method,
quadratic stress method, log-law method and turbulent kinetic energy method, were assessed in observations and models. The algorithms for surface wave-induced and current-induced stress applied in RANS models were validated against the observed surface wave stress and quadratic stress. The possibility to compute $\tau_b$ based on model output $\varepsilon$ was explored for future sediment resuspension model. In Chapter 5, general conclusions are drawn and suggestions for future research are given.
Chapter 2

Mixing and dissipation of turbulent kinetic energy in Lake Erie

2.1 Introduction

Small-scale hydrodynamic processes in lakes and specifically the diapycnal mixing associated with turbulence transports mass and momentum contribute to regulate biogeochemical processes. In Lake Erie, these include algal biomass delivery to benthic filter feeders (e.g., Ackerman et al., 2001; Boegman et al., 2008), sediment resuspension (e.g., Bedford and Abdelrhman, 1987; Valipour et al., 2017), transport of oxygen (e.g., Bouffard et al., 2013; Bouffard et al., 2014) and nutrients (e.g., Schwab et al., 2009).

The seasonal thermal cycle (Schertzer et al., 1987; Boegman et al., 2001), water-level fluctuations (Platzman and Rao, 1964), mean current patterns (Saylor and Miller, 1987; Beletsky et al., 1999) and internal wave dynamics (Bouffard et al., 2012; Valipour et al., 2015) in Lake Erie have been extensively studied and documented in the literature. However, spatial and temporal distributions of small-scale processes, like turbulent diffusivity and dissipation, remain comparatively unexplored in Lake Erie and the Great Lakes in general.

This is because direct measurement of the rate of dissipation of turbulent kinetic energy ($\varepsilon [m^2s^{-3}]$) and the turbulent diffusivity ($K_z [m^2s^{-1}]$) remains difficult (e.g., Ivey et al., 2008). The turbulent kinetic energy (TKE) introduced to a lake by the surface wind stress (Wuest et al., 2000; Valipour et al., 2015) is ultimately dissipated into heat by viscous friction at a rate given by $\varepsilon$ (e.g., Imberger 1998; Bouffard et al., 2012), with the ratio between irreversible mixing and dissipation that can be characterized by the buoyancy Reynolds number $Re_b = \varepsilon / \nu N^2$ (Imberger and Ivey, 1991; Shih et al., 2005; Bouffard and Boegman, 2013; Monismith et al., 2018), where $\nu$ is viscous diffusivity [m²s⁻¹], and $N$ is buoyancy frequency [s⁻¹].
This paper summarizes the distribution of $\varepsilon$ and $K_Z$ in Lake Erie, to 1) better understand spatial and temporal differences in turbulent processes and associated vertical biogeochemical fluxes; 2) complete the TKE budget; and 3) provide a dataset to enable validation of closure schemes in three-dimensional hydrodynamic models (Ackerman et al., 2001; Rao et al., 2008).

Wuest and Lorke (2003), Ivey et al. (2008), and Monismith and MacIntyre (2009) showed that lakes have stronger dissipation and mixing in surface and bottom mixed layers due to friction. In comparison to the significantly increased turbulence and mixing in boundary layers, the interior turbulence of medium-sized stratified water systems was extremely weak, even though baroclinic basin-scale internal seiches and near-inertial internal waves which are active beneath the surface layer carry a large amount of mechanical energy. However, in some large lakes (e.g., Lake Erie [Bouffard et al., 2012], Lake Michigan [Choi et al., 2012]) self-induced shear of high-frequency internal waves (HFIW) generated from near inertial internal waves have also been considered to make a contribution to dissipation and mixing in the pelagic thermocline.

In Lake Erie, Poincaré waves are the dominant wind-induced response beneath the surface layer (Rao et al., 2008; Valipour et al., 2015), but can become unstable if the lake is shallow and subjected to strong wind forcing. And associated enhanced vertical shear at the crests and troughs of the Poincaré waves can be the source of increased $\varepsilon$. Therefore, it is reasonable to expect the high level of $\varepsilon$ in the surface, bottom layer and thermocline region of Lake Erie, but what are the ranges and mean values of $\varepsilon$ in these various regions? Considering the unique topography of Lake Erie (multi-basin), what are the regional characteristic values of $\varepsilon$ and $K_Z$? Do the variabilities of $\varepsilon$ and $K_Z$ depend on temporally and spatially thermal and hydrodynamic features? And does TKE balance in Lake Erie agree with that in other lakes of varying size; if not, what are the causes of discrepancies?
2.2 Methods

2.2.1 SCAMP casts

During the summers of 2008-2009, intensive field campaigns were carried out in Lake Erie. These campaigns included the collection of more than 600 temperature microstructure casts using a Self-Contained Autonomous Microstructure Profiler (SCAMP; PME Inc.), which is a hand-held instrument designed to measure small-scale fluctuations in temperature. Free-falling through the water column at ~ 0.1 m s\(^{-1}\) and sampling at 100 Hz, temperature microstructure fluctuations as small as 1 mm were recorded with an accuracy of 0.001°C.

The SCAMP was deployed from the Fast Rescue Craft of the CCGS Limnos, a Canadian Coast Guard Research and Survey vessel, during lake-wide cruises and from the Keenosay, an Ontario Ministry of Natural Resources and Forestry vessel departed from the Wheatley Harbour, during daily cruises east of Pelee Island (Figure 2.1). Microstructure casts were collected at 11 sites in Lake Erie (Table 2.1) in a downward-looking profiling mode. Bouffard et al. (2012) and Bouffard and Boegman (2013) published part of these data in a methods paper, but did not characterize how \( \varepsilon \) and \( K_Z \) vary with depth, time and location throughout the lake.

The downward-looking SCAMP profiles could not measure data within ~ 2 m of the surface. To account for this, the 2008-09 data were supplemented with upward profiling SCAMP casts (McCune, 1998) in central Lake Erie, ~8 km off the coast (hereafter, Sta. M). The data was collected in May 1997, as the stratification just began, and in August 1997 after strong seasonal stratification had developed (Table 2.1). These data fill the near-surface gap (see Bouffard et al. 2012), and also provide casts both before and after a storm event (Table 2.1 and Figure 2.1).
<table>
<thead>
<tr>
<th>Station</th>
<th>Location</th>
<th>Date (number of casts)</th>
<th>Number of profiles</th>
<th>Group</th>
</tr>
</thead>
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<td>5</td>
<td>1</td>
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<tr>
<td></td>
<td></td>
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<td>1</td>
</tr>
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<tr>
<td></td>
<td></td>
<td>Aug 5, 6, 7 2008</td>
<td>14/14/1</td>
<td>1</td>
</tr>
<tr>
<td></td>
<td></td>
<td>May 28 2009</td>
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<tr>
<td></td>
<td></td>
<td>Jul 1, 16 2009</td>
<td>3/6</td>
<td>1</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Aug 4, 5, 6, 7, 25, 26, 27 2009</td>
<td>10/9/11/4/12/7/7</td>
<td>1</td>
</tr>
<tr>
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<td>Aug 6 2008</td>
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<td></td>
<td></td>
<td>Jun 30 2009</td>
<td>1</td>
<td>1</td>
</tr>
<tr>
<td>SE Shoal</td>
<td>82° 55' 9&quot;W, 41° 33' 32&quot;N</td>
<td>Aug 7 2008</td>
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<td>1</td>
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<td>5/2/9</td>
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<td>82° 16' 59&quot;W, 41° 47' 29&quot;N</td>
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<td>May 28 2009</td>
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<td></td>
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<tr>
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<td>Jul 1, 2 2009</td>
<td>6/12</td>
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<td>82° 17' 33&quot;W, 41° 43' 48&quot;N</td>
<td>Jul 31 2008</td>
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<td>452</td>
<td>79° 55' 17&quot;W, 42° 34' 52&quot;N</td>
<td>Jul 29 2008</td>
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<td>Aug 19 2009</td>
<td>3</td>
<td>3 (seasonally stratified)</td>
</tr>
</tbody>
</table>
Figure 2.1 Map of Lake Erie showing observation stations (red dots) and meteorological data stations (black dots). Blue dash line indicated the rough cruise route followed by the research vessel CCGS Limnos during Jul 13 – 16, 2009. Blue dash circle shows the rough area where SCAMP casts were collected from the Keenosay.

2.2.2 Currents, temperatures and wind speed

To measure the ambient conditions, thermistors (RBR TR-1060, 0.001°C accuracy, 0.1-Hz sampling) were deployed over the water column depth with ~ 0.5-1 m vertical spacing (at Sta. 357, 341, 1227, 1228, 1231, 84, 452), and a tripod equipped with upward and downward-looking acoustic Doppler current profilers (600 kHz ADCP and 2 MHz HR-ADCP; Nortek Aquadop, accuracy ±1% of measured values > 0.005 m s⁻¹), was deployed 1.8 m above the bottom at Sta. 341, with 1 Hz sampling frequency and 15 min sampling interval. Details of these moorings can be found in Bouffard et al. (2012), Bouffard et al. (2013), and Valipour et al. (2017).

The wind speed and direction in the central basin were obtained at 1 h intervals, 4 m above the water surface from National Data Buoy Center (NDBC) Sta. 45005 and in the eastern basin were obtained from Environment and Climate Change Canada (ECCC) Sta. 45142 (Port
Colborne) at 1 h intervals, and the anemometer was 5 m above the water surface. Wind speeds were logarithmically rescaled to the standard 10 m height $W_{10}$ based on Garratt (1977).

2.2.3 Calculation of $\varepsilon$ and $K_z$

TKE dissipation was computed from the SCAMP cast data by fitting to the theoretical Batchelor spectrum. In order to calculate $\varepsilon$, we need stationary profiles of the turbulence, which freeze turbulent motions in time, but the temperature signals from the SCAMP are nonstationary. McCune (1998) used an autoregressive (AR) model to identify stationary turbulent patches (see also Imberger and Ivey (1991)). Here, we followed Bouffard et al. (2012) and Bouffard et al. (2013), who employed uniform fixed 25 cm segments. Our own comparisons (e.g., Bouffard and Boegman, 2013; unpublished data) and those of others (Pernica et al., 2014; R. Botero, personal communication) showed no significant differences between variable AR and fixed uniform segments.

After stationary turbulent patches were determined, each segment was converted to a temperature gradient $\left(\frac{\partial T'}{\partial z}\right)$ spectrum. $\varepsilon$ was obtained by fitting these observed spectra to the theoretical Batchelor spectra, whose cutoff wavenumber ($k_B$) depends on $\varepsilon$. The quality of the fit was tested using the maximum likelihood spectral fitting (Ruddick et al., 2000). The Batchelor fitting equations (Eq. A1-4) can be found in Appendix A.

Measuring $K_z$ is challenging in the field, therefore by following Ivey et al. (2008) and Osborn (1980), we use $\varepsilon$ and mixing efficiency ($\Gamma$) to parameterize $K_z$

$$K_z = \Gamma \frac{\varepsilon}{N^2}$$

(2.1)

, where $N = \sqrt{\frac{-g \frac{\partial \rho}{\partial z}}{\rho_0 \frac{\partial \rho}{\partial z}}}$ is the buoyancy frequency, $\rho_0 = 1000$ kg $m^{-3}$ is the reference density.

Here, we followed Bouffard and Boegman (2013) to calculate $K_z$ and $\Gamma$ (Table A1) as a function
of $Re_b$, which also can be interpreted as turbulence intensity, and Prandtl number $Pr$ (ratio of the viscosity over the molecular diffusivity). See also Shih et al. (2005) and Monismith et al. (2018).

2.2.4 Calculation of wind input energy

The wind energy flux at 10-m above the surface was estimated from (Lombardo and Gregg, 1989; Wu, 1980),

$$P_{10} = \rho_a C_d W_{10}^3$$  \hspace{1cm} (2.2)

where $\rho_a = 1.2$ kg m$^{-3}$ is the air density, and $C_d = (0.8 + 0.065W) \times 10^{-3}$ is the drag coefficient for most frequently occurring wind velocity ($W_{10} > 1$ m s$^{-1}$).

TKE budget – We investigated the TKE balance by computing the integral of $\varepsilon$ in both the surface layer and the total lake interior (Antenucci et al., 2000; Wuest et al., 2000; Bouffard et al., 2012),

$$\overline{P_{suf}} = \int_0^{h_s} \rho(z)\varepsilon(z) \, dz$$ \hspace{1cm} (2.3)

$$\overline{P_{tot}} = \int_0^{h} \rho(z)\varepsilon(z) \, dz$$ \hspace{1cm} (2.4)

where $\rho(z)$ is the water density as a function of water depth, $h_s$ is surface layer depth and $h$ is water depth. Wuest et al. (2000) showed that $\varepsilon$ was a function of depth, and decreased sharply below surface layer. We visually identified the sudden decrease in $\varepsilon$ profiles, and defined the layer above as the surface layer.

Flow stability – The gradient Richardson number, $Ri = \frac{N^2}{S^2}$, was used to identify locations of instability to current shear in the flow. $S^2 = \left( \frac{\partial \sqrt{(U_x^2+U_y^2)}}{\partial z} \right)^2$ is the vertical current shear, where $U_x$ and $U_y$ are the eastward and northward components of the mean current velocity,
respectively. This parameter has been assumed to play an intermediary role in the transfer of energy from the internal wave field to turbulence (Polzin, 1996). The Miles–Howard criterion stipulates that the sufficient condition for stability to small perturbations is a local Ri > 0.25 everywhere in the flow. Unstable modes can grow into finite-amplitude perturbations, leading to turbulence, when this canonical limit is not satisfied (e.g., Turner, 1973). For unsteady flows, Ri << 0.25 may be required due to the shearing timescale being less than the growth period associated with an unstable mode (Troy and Koseff, 2005). Time-series of Ri were computed from 1-m bin ADCP data; because temporal and spatial sampling may alias shear layers (Boegman et al., 2001), we took the stability threshold as Ri = 1 (i.e., Ri < 1 was defined as shear instability) (Horn et al. 1986; Polzin, 1996; Bouffard et al. 2014).

2.3 Results

2.3.1 Characteristic and mean value of ε and Kz

ε and Kz profiles were clustered according to water depth and seasonal stratification. The observations revealed three distinct regions (Western basin group: <12 m [hereafter Group W]; Central basin group: 12-25 m [hereafter Group C]; and Eastern basin group: >25 m [hereafter Group E]) with two distinct periods (seasonally stratified and neutrally stratified). Seasonal stratification was observed in Group C and E; however, because of sampling limitations, comparison between seasonally and neutrally stratified observations were only available for Group C. The isopycnal-averaged ε and Kz of different groups, based on all profiles in a group, were graphically compared (Figure 2.2), with mean ε and Kz for different water layers given in Table 2.2.

Group W (Figure 2.2, yellow region), with observations from the western and west-central basin, was assumed to also be representative of the nearshore regions of the east-central and
eastern basins, based on water depth and proximity to shore. The highest values of $\varepsilon$ and $K_z$ were in the surface layer, which has most direct response to wind energy input and thermodynamic heat transfers. Because of the shallowness of the stations, the mean $\varepsilon$ of all turbulent segments in this group was highest, and the variation with depth was not significant. However, the mean $K_z$ in the bottom layer was lowest amongst all groups. The SCAMP profiles were collected mainly in July and August, when the thermocline was deep (~10-15 m depth; Schertzer et al., 1987), with the proximity of the thermocline to the lake bed possibly leading to the low $K_z$ values (Boegman, 2006) in Group W (see section 2.3.2).

Observations in Group C (Figure 2.2, red region), were divided into neutrally and seasonally stratified periods. During neutral stratification (Figure 2.2c, d), the distribution of turbulence was similar to that in a medium-sized lake (with wind fetch ~10 km), with the energetic surface layer and the turbulent bottom boundary layer (Wuest and Lorke, 2003). Due to wind, surface wave, and convective cooling induced turbulence, we observed maximum values of $\varepsilon \sim 10^{-6}$ m$^2$ s$^{-3}$ and $K_z \sim 10^{-3}$ m$^2$ s$^{-1}$ in the surface boundary layer (Monismith et al., 2009). During the seasonally stratified period, the isopycnal-averaged $\varepsilon$ profile showed a pronounced metalimnetic increase, with a mean value over $10^{-7}$ m$^2$ s$^{-3}$, but $K_z$ was one order of magnitude lower, during seasonal stratification, than the neutrally stratified period (Table 2.2).
Figure 2.2 Mean (colored bars) and standard deviation (grey shallow) of $\varepsilon$ and $K_z$ in each group (map (i) shows yellow, red and blue are Group W, Group C and Group E, respectively, and black dots represent observation stations). In the left panels (a, c, e, g), primary ordinates are $\log_{10} \varepsilon$, and secondary ordinates are temperature. The temperature profiles are the typical profiles in each group. The right panels (b, d, f, h) are $\log_{10} K_z$. The y-axes of all panels are the normalized depth.
Table 2.2 Mean and standard variation of $\varepsilon$ and $K_z$ in varying regions of each group.

<table>
<thead>
<tr>
<th></th>
<th>Group W</th>
<th>Group C Neutrally stratified period</th>
<th>Group C Seasonally stratified period</th>
<th>Group E</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>$\varepsilon \times 10^{-8}$</td>
<td>$K_z \times 10^{-7}$</td>
<td>$\varepsilon \times 10^{-8}$</td>
<td>$K_z \times 10^{-7}$</td>
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<td>Surface layer</td>
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<td>epilimnion</td>
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<td>/</td>
<td>/</td>
<td>/</td>
</tr>
<tr>
<td>Interior (Group W and neutrally stratified Group C)</td>
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<td>560</td>
<td>5.7</td>
<td>1700</td>
</tr>
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<td>Hypolimnion</td>
<td>/</td>
<td>/</td>
<td>/</td>
<td>/</td>
</tr>
<tr>
<td>Bottom layer</td>
<td>40</td>
<td>510</td>
<td>7.2</td>
<td>1500</td>
</tr>
<tr>
<td>All depth</td>
<td>21</td>
<td>680</td>
<td>14</td>
<td>2400</td>
</tr>
</tbody>
</table>

* The surface layers and the bottom layers in each group were defined as 2 m from surface and 2m above lakebed (Bedford and Abdelrhman, 1987; Valipour et al., 2015), respectively. Group W and neutrally stratified Group C were separated as surface layer, interior, and bottom layer; rather seasonally stratified Group C and Group E were separated as surface layer, epilimnion, metalimnion, hypolimnion, and bottom layer. Metalimnion was defined where temperature gradient changed sharply.

Group E (Figure 2.2, blue region) only contained 5 profiles collected during mid-summer (seasonally stratified period) at Sta. 452 in the eastern basin ~10 km from Long Point. The mid-depth (interior) water column in this deep basin was expected to be quiescent away from surface and bottom boundary layers (Wuest and Lorke, 2003). However, the mid-depth $\varepsilon$ was similar to that in the surface and bottom boundary layers ($\sim 10^{-7}$ m$^2$ s$^{-3}$, Figure 2.2g), showing similar characteristics to Group C during the seasonally stratified period. The metalimnion was much thicker (more diffuse) than in the central basin (Figure 2.2), as also reported by Schertzer et al. (1987) who observed a 25 m thick metalimnion during July and August. Among 5 profiles collected in Group E station (Sta. 452), 3 profiles (Aug 19$^{th}$ 2009) were collected after a storm event with wind speed over 10 m s$^{-1}$ coming from southwest, which could set-up basin-scale
internal waves in the eastern basin (Boegman, 1999; his pg. 112) and a return flow beneath the thermocline (Bartish, 1987), generating shear-induced mixing that elevated $\varepsilon$, and the relatively weak stratification (compared to the central basin) allowed for enhanced $K_z$. This could be the mechanism that led to the high $\varepsilon$ and $K_z$ in both the metalimnion and hypolimnion of Group E; however, more data is needed in this region. We also investigated the occurrence frequency of $Re_b$, $\Gamma$ and $K_z$ of each group as shown in the Appendix A (Text A2, Figure A1).

Now that the basin-average turbulence and mixing characteristics have been identified, in the following section, the specific processes leading to temporal and spatial variation of $\varepsilon$ and $K_z$ in Groups W and C are investigated.

### 2.3.2 Variation of $\varepsilon$ and $K_z$ in bottom layer of Group W

Orbital currents from wind waves and oscillatory seiche-currents often penetrate to the lake bed, creating a bottom boundary layer (BBL) (Bedford and Abdelrhman, 1987; Valipour et al., 2017) with intense dissipation that is one of the most turbulent zones in a lake (Wuest et al., 1996, Imberger 1998; Wuest and Lorke, 2003). However, at the transition between the western and central basins (Sta. 1227) of Lake Erie, the BBL dynamics were modulated by the seasonal thermocline, which laid close to the lake bed in mid-July (Figure 2.3). In this stratified region, $K_z$ was modulated up and down by high-frequency internal waves (HFIWS, Figures 2.3 and 2.4; Bouffard and Boegman, 2012). Immediately above the stratified region and at the crests of the HFIWs (cast 1, 4; Figure 2.3), $\varepsilon$ ($\sim10^{-7}$ m$^2$s$^{-3}$) and $K_z$ ($\sim10^{-5}$ m$^2$s$^{-1}$) were higher (Figure 2.4f, i), likely due to shear instability at the base of the mixed layer (Antenucci and Imberger, 2001; Boegman et al., 2003). At the troughs and notes of the HFIWs (cast 2, 5; Figure 2.3), $K_z$ ($\sim10^{-7}$ m$^2$s$^{-1}$) was lower (Figure 2.4g, j) because the contribution from turbulent eddies to vertical diffusivity was inhibited near bed.
Figure 2.3 Time-series of thermal structure at Sta. 1227. Isotherms are plotted every 1°C from 19°C. Inset shows the time-series of isotherm for two days. The vertical dashed lines indicate microstructure casts. Red arrows show the depths of temperature loggers.
Figure 2.4 Individual $\varepsilon$ profiles (a–e); turbulent vertical diffusivity (f–j) on July 16 2009 (corresponding with the time indicated in Figure 2.3), at Sta. 1227.
**Figure 2.5** Individual $\varepsilon$ profiles and $K_z$ profiles collected (a) on August 4th 1997, (b) on August 6th 1997, at Sta. M; (c) on May 28th 2009, (d) on July 21st 2008, at Sta. 341; (e) on August 19th 2009, at Sta. 452; (f) on August 6th 2008, at Sta. 1228; (g) on July 31st 2008, at Sta. 341; (h) on July 30th 2008, at Sta. 1228. In left panels, primary x-axis are $\log_{10} \varepsilon$ (gray bars), and secondary x-axis are temperature (black solid lines). In right panels, x-axis are $\log_{10} K_z$ (blue solid lines).

### 2.3.3 Variation of $\varepsilon$ and $K_z$ in Group C

In Group C the turbulent response of $\varepsilon$ and $K_z$ to storm events differed depending on the water column stratification. During seasonal stratification, a storm event on August 5, 1997 (winds $\sim$10 m s$^{-1}$) enhanced $\varepsilon \sim 10^{-6}$ m$^2$ s$^{-3}$ and $K_z \sim 10^{-3}$ m$^2$ s$^{-1}$ in both the surface and bottom mixed layers at Sta. M, with entrainment sharpening the thermocline (Figure 2.5a, b). During neutral stratification, a storm event on May 26, 2009 (winds $\sim$10 m s$^{-1}$; Figure 2.6) at Sta. 341 created sporadic remnant turbulent patches at mid-depth with $\varepsilon$ and $K_z$ spanning 4 orders of magnitude (Figure 2.5c). Both the mean and individual $\varepsilon$ profiles indicated that turbulence in the water column, during the neutrally stratified period, was not always weak (Figure 2.2c), which was distinct from observations in Lake Michigan (Choi et al., 2012), perhaps due to the significant difference in lake depths and corresponding volume of the water column occupied by boundary layers.
During calm conditions, the epilimnion may be weakly stratified (Imberger 1985; Pernica et al., 2014), which can inhibit active vertical mixing in Lake Erie (Boegman et al., 2008). In our observations, weak stratification was observed (Figure 2.5d), where $\varepsilon$ and $K_z$ decreased sharply with depth, from around $10^{-7} \text{m}^2 \text{s}^{-3}$ and $10^{-3} \text{m}^2 \text{s}^{-1}$ in the surface mixed layer (above 2 m) to $10^{-9} \text{m}^2 \text{s}^{-3}$ and $10^{-6} \text{m}^2 \text{s}^{-1}$ below 2 m depth, indicating surface mixing did not penetrate to depth.

Storm events energize internal wave activity and turbulence within the metalimnion (Saggio and Imberger, 1998; MacIntyre et al., 1999; Boegman et al., 2003). In the Great Lakes, it is near-inertial internal waves that generate vertical shear through the thermocline, inducing high $\varepsilon$.
in the lake interior (Bouffard et al., 2012; Choi et al., 2012); however, the strong density gradient through the metalimnion maintains low $K_z$.

Our observations were similar on 22 July, 2008 at Sta. 341, where $\varepsilon$ was elevated through the metalimnion. The thermocline was close to the surface (1-7 m depth), with density overturns; the Poincaré waves (shown in the inset of Figure 2.7a) acting as a waveguide for HFIWs (Figure 2.7a). The SCAMP casts sampled the Poincaré wave crests, which had relatively higher $\varepsilon$ than the nodes (Bouffard et al., 2012). Turbulent patches were concentrated in the thermocline region when the thermocline was thin and near surface, and where overturns were associated with increased $\varepsilon$ (Figure 2.8a-e). When the metalimnion expanded (Figure 2.8b-e), $\varepsilon$ increased at the base of thermocline. Correspondingly, $Ri$ was low (<1; Figure 2.7b) through the metalimnion, where $N^2$ remained high (Figure 2.7c), indicating that shear instability perturbed the stratification. The high $\varepsilon$ and relatively weak stratification, caused $K_z$ to have local peaks at the base of metalimnion, exceeding $10^{-4} \text{m}^2 \text{s}^{-1}$, indicating active vertical mixing.
Figure 2.7 Time-series of (a) 10s resolution thermal structure, arrows on y-axis show the depths of temperature loggers; (b) $R_i$, stars on y-axis show the sampling positions of upward looking ADCP; (c) $N^2$ on July 22 2008 at Sta. 341. Isotherms are plotted every 1°C from 10°C. Inset in (a) shows the time-series of isotherm for two days over two Poincaré wave periods. Yellow dash lines indicate microstructure profiles shown in Figure 2.8.
Typically a calmer hypolimnion was present beneath the thermocline and above the bottom boundary layer, where $\varepsilon$ is reduced by up to 3 orders of magnitude (e.g., Figure 2.5b, f).

However, the hypolimnion is not always quiescent in large shallow lakes. Our observations showed that wind seiche-induced currents (Rao et al., 2008, Bouffard et al., 2012) led to enhanced turbulence within the hypolimnion, with two mechanisms being observed:

1. Entrainment events: When the Lake Erie thermocline expanded under forcing from strong hypolimnetic currents, fluid parcels from the metalimnion could be entrained into the hypolimnion (Ivey and Boyce, 1982, Schertzer et al., 1987). We observed these events at Sta. 341 and 1228 during late July (Figure 2.5g, h), with a thermocline thickness ~ 4-5m and $\varepsilon$ at the top of the hypolimnion increased to $10^{-7}$ m$^2$s$^{-3}$; 2-3 orders of magnitude larger than the quiescent
hypolimnion (Figure 2.5b, f). These events also elevated $K_z$ at the interface between the metalimnion and hypolimnion to $10^{-4} \text{m}^2\text{s}^{-1}$. Concurrent data from thermistors and ADCPs was not available during the observed entrainment events; thus, we were unable to investigate the evolution of thermocline and current velocities during this time. Valipour et al. (2015b) reported baroclinic motions during July 28-30, 2009, which may have forced these events.

(2). Basin-scale seiche events: A thermocline compression event (Figure 2.9b, c) resulted from a southwest wind (days 178-180; Figure 2.9a) accelerating current velocities in the hypolimnion, generating strong shear (Figure 2.9f,g) and mixing and dissipation near the bottom (Figure 2.9c, e) at Sta. 341 and 1228. While this was likely a remnant Poincaré wave, reported by Valipour et al. (2015b) at Sta. 341 on days 177.5-180, a similar vertical mode-two basin-scale wave event, generating shear instabilities was observed by Beogman et al. (2003) in Lake Kinneret and Lake Biwa. Seiching motions are well known to expand the turbulent bottom boundary layer through mixing (Gloor, 1994). A secondary wind event (Figure 2.9b, day 182), increased westward currents ($\sim0.15 \text{ m s}^{-1}$; Figure 2.9g) diffusing the thermocline at Sta. 341 and generating an internal thermocline jump at Sta. 1228. The residual momentum from these near inertial waves persisted for days, generating high $\varepsilon$ (Valipour et al., 2015b).
Figure 2.9 Time-series of (a) wind velocity and direction from buoy, and thermal structure at (b) Sta. 1228; (d) Sta. 341, the arrows on y-axis are the positions of temperature sensors. Isotherms are plotted every 1 °C from 13 °C. The vertical dash lines represent the times of microstructure profiles. (c) and (e) are profiles of $\varepsilon$ and $K_z$ corresponding to the dash line within isotherms. (f) and (g) are north-southward (u) and east-westward (v) current velocity at Sta. 341.

In general, the typical characteristics of entrainment events include high $\varepsilon$ in the metalimnion, a thickened thermocline and strong hypolimnetic currents. For wind induced basin-scale seiche events, we observed a compression of the thermocline forced by a wind event and an accelerating hypolimnion leading to enhanced and expanded turbulence from bottom mixed layer.
2.4 Discussion

2.4.1 Internal wave degeneration

Mixing and dissipation in the interior of lakes is considered negligible when compared with that in the turbulent bottom boundary layers at the lake perimeter (Imberger, 1998; Wuest et al., 2000; Wuest and Lorke, 2003). Horn et al. (1999) proposed four mechanisms for the degeneration of internal seiches in long narrow lakes: viscous damping; the formation of shear instabilities in the interior; the production of nonlinear internal waves that will break on sloping topography; and the formation of internal hydraulic jumps.

Viscous damping can be seen in Figure 2.9, where near-inertial seiches caused $\epsilon > 10^{-7} \text{m}^2\text{s}^{-3}$ in the hypolimnion and $K_z > 10^{-4} \text{m}^2\text{s}^{-1}$ close to the bottom. Shear instabilities, resulting from Poincaré wave-induced baroclinic shear has been observed in the lake interior (Bouffard et al., 2012). We observed these occurrences (Figure 2.7, 2.8) to cause high $\epsilon (\sim 10^{-6} \text{m}^2\text{s}^{-3})$ in the central basin (Group C) during the seasonally stratified period.

Nonlinear internal (solitary) waves have not been observed in Lake Erie, despite both observational- (Valipour et al., 2015b) and modelling- (Liu and Lamb, 2011) based investigations. Unlike planar internal seiches that steepen producing internal solitary waves, it is expected that an internal Kelvin wave will rather leak energy to radiating internal Poincaré waves (Fuente et al., 2008; Mortimer, 2004). Internal hydraulic jumps have been modeled and/or observed in several lakes (Geneva, Thorpe et al., 1996; Biwa, Saggio and Imberger, 1998; Cayuga, Dorostkar and Boegman, 2013), with those in Lake Geneva and Lake Biwa being similar to those observed during the thermocline compression event in Lake Erie (Figure 2.9b).
2.4.2 Turbulent Kinetic Energy Budget

A turbulent kinetic energy budget has been applied to study the energy flux paths in lakes (Raven et al., 2000; Wuest et al., 2000; Bouffard et al., 2012). Bouffard et al. (2012) did not estimate $\varepsilon$ in the surface layer because downward SCAMP profiles were not able to measure near the free surface and consequently recommended analysis of upwards casts to better examine surface energy fluxes. Using the upward SCAMP profiles collected by McCune (1998), we were able to extend the budget to consider the surface layer (Table 2.3). In agreement with Bouffard et al. (2012)’s observations, the energy flux to the surface layer and total water column were shown to be more efficient during weak wind forcing. Both the ratio of energy flux estimated in the surface to wind energy ($\overline{P_{\text{surf}}}: P_{10}$) and the ratio of energy flux estimated in the total water column to wind energy ($\overline{P_{\text{tot}}}: P_{10}$) decreased with increase wind forcing, from 2.2% to 0.23% and 3.5% to 0.49% of $P_{10}$, respectively, which they hypothesized to be caused by energy loss to wind-wave breaking during storms, but may also be a result of less efficient transfer from Coriolis acting on currents in the surface layer (Boegman et al., 2001; Valipour et al., 2015b). Energy flux estimated in the surface mixed layer ($\overline{P_{\text{surf}}}$) did not change a lot with wind forcing conditions, but energy flux in the total water column ($\overline{P_{\text{tot}}}$) increased with wind forcing. Casts representing strong wind forcing conditions were collected after strong wind events, but within one inertial period (~17 h). The surface mixed layer responds very quickly to the wind and also quickly damps TKE, through surface wave generation and breaking, while the lake interior responds more slowly as it takes several days for TKE to be damped (Valipour et al., 2015b). Therefore, the energy flux associated with internal motion was observed, but $\varepsilon$ related to wind-induced surface wave breaking may not have been captured in these data. Comparing the weak and strong wind forcing, the energy flux entering the basin-scale internal wave field increased
with wind forcing, with changes in interior flux potentially resulting from changes in the stability characteristics of the Poincaré wave and internal seiche (Figure 2.9b-e).

Comparison among Lake Erie, Alpnach, and Baikal showed that $\overline{P_{suf}}: P_{10}$ increased with the lake surface area under weak wind forcing conditions. Wuest et al. (2000) concluded the water column below the surface mixed layer dissipated $0.29 \pm 0.12 \%$ of $P_{10}$ based on the data collected in lakes with various size and internal motions. Lake Erie data agreed with their estimates, under strong wind forcing conditions, with $0.26\%$ of $P_{10}$ dissipated under the surface layer as TKE, but under weak wind forcing conditions, there was $1.3\%$ of $P_{10}$ dissipated beneath the surface mixed layer $(\overline{P_{tot}}: P_{10} = 3.5\%) - \overline{P_{suf}}: P_{10} = 2.2\% = 1.3\%)$. Based on the casts collected in 2008 and 2009 under various wind forcing conditions, Bouffard et al. (2012) estimated an average ratio of $0.85\%$, ranging from $1.5\%$ and $0.33\%$ under weak and strong wind forcing conditions, respectively. Therefore, the estimates in Lake Erie based on two datasets are highly consistent, and in conclusion, the morphometry of Lake Erie (higher ratio of surface area to depth) contributes to the higher energy flux estimated beneath the surface mixed layer by increasing TKE dissipated in side and bottom boundary layers and in metalimnion via near-inertial waves and internal seiche currents. Similar in the scale of surface area, wind energy input in Lake Baikal was purely inertial rather than via both inertial motions and wind seiche currents (longitudinal seiches) in Lake Erie (Raven et al., 2000; Rao et al., 2008), leading to less energy flux transferred into the lake interior. Also, with deeper depth, Lake Baikal is expected to have less TKE dissipated in BBL compared to Lake Erie. Overall, the ratio of the energy flux in the lake surface to that in the whole lake $(\overline{P_{suf}}: \overline{P_{tot}})$ in Lake Erie is smallest among the three lakes, especially under the strong wind condition, emphasizing the energetic nature of lake interior of this large and shallow lake.
Table 2.3 Turbulent kinetic energy budget in the surface layer of Lake Erie. The overbar indicates a mean value.

<table>
<thead>
<tr>
<th>Lake</th>
<th>Lake Erie (present study)</th>
<th>Lake Alpnach (Wuest et al 2000)</th>
<th>Lake Baikal (Raven et al 2000)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Surface area $A$ (km$^2$)</td>
<td>25744</td>
<td>4.8</td>
<td>31722</td>
</tr>
<tr>
<td>Average depth (m)</td>
<td>19</td>
<td>22</td>
<td>744</td>
</tr>
<tr>
<td></td>
<td>23</td>
<td>7</td>
<td>130</td>
</tr>
<tr>
<td>Wind forcing</td>
<td>weak</td>
<td>strong</td>
<td>weak</td>
</tr>
<tr>
<td>$P_{10}$ (mWm$^{-2}$)</td>
<td>59*</td>
<td>571*</td>
<td>32</td>
</tr>
<tr>
<td>$\overline{P}_{suf}$ (mWm$^{-2}$)</td>
<td>1.3</td>
<td>1.3</td>
<td>0.47</td>
</tr>
<tr>
<td>$\overline{P}_{tot}$ (mWm$^{-2}$)</td>
<td>2.1</td>
<td>2.8</td>
<td>0.74</td>
</tr>
<tr>
<td>$\overline{P}<em>{suf}$: $P</em>{10}$ (%)</td>
<td>2.2</td>
<td>0.23</td>
<td>1.5</td>
</tr>
<tr>
<td>$\overline{P}<em>{tot}$: $P</em>{10}$ (%)</td>
<td>3.5</td>
<td>0.49</td>
<td>1.9</td>
</tr>
<tr>
<td>$\overline{P}<em>{suf}$: $\overline{P}</em>{tot}$ (%)</td>
<td>64</td>
<td>46</td>
<td>79</td>
</tr>
</tbody>
</table>

* Calculated over a period of 2 d before the casts. Wind speed data have been rescaled logarithmically to standard 10-m height.

2.4.3 Comparison to observations in other lakes

Variations in lake morphometry and stratification present a wide range of turbulent responses to wind forcing (Antenucci and Imberger, 2001; Saggio and Imberger, 2001; Bouffard et al., 2012). Maiss et al. (1994) summarized the relation between mean values of $K_Z$ during summer stratification in the hypolimnion of various lakes and the lake surface area (Figure 2.10a).

Adding recent data in metalimnion and hypolimnion, we found a more rapid increase in $K_Z$ with lake surface area than that in Maiss et al. (1994).

To include the baroclinic response to wind, we plotted $K_Z$ against the nondimensional Wedderburn number $W = g' h_s^2 / (u_{a*} l)$ (Imberger and Patterson, 1990), where $g' = \frac{g \Delta \rho}{\rho_a}$ (m s$^{-2}$) is the reduced gravity, $h_s$ is the surface layer depth, $\Delta \rho$ is the density difference across the pycnocline (kg m$^{-3}$), $u_{a*} = (\rho_a / \rho_w C_d W_{10})^{1/2}$ is the friction velocity, and $\rho_w$ is the density of water, respectively. $l = 2(A/\pi)^{1/2}$ is the characteristic length scale (m), computed from the the...
surface area $A$. Beneath the surface mixed layer, $K_z$ decreases with increasing $W$, in the form

$$K_z = 10^{-5 \pm 0.8} W^{-0.5},$$

showing increased stratification or decreased winds lower $K_z$ over a range of lakes with variable bathymetry (Figure 2.10b).

Typical values of $\varepsilon$ and $K_z$ from the present study and those in the literature are shown in Appendix A (Table A2) with corresponding Kolmogorov length, time, and velocity scales. Spatial variability of $\varepsilon$ and $K_z$ exists in each lake, with $\varepsilon$ spanning 2 – 5 orders of magnitude in metalimnion. The $\varepsilon$ in BBL consistently has less variability ($\sim 10^{-7}$ m$^2$ s$^{-3}$), with $K_z$ more variable due to the effects of density stratification on mixing efficiency (Lorke, 2009).

**Figure 2.10** Observed $K_z$ during stratification beneath the surface mixed layer (metalimnion and hypolimnion) of various lakes as function of (a) lake surface area, and (b) Wedderburn number. Lakes are represented using their first letter(s): Er = Erie, Bai = Baikal, On = Ontario, Co = Constance, Si = Simcoe, Ki = Kinneret, Mo = Mono, Al = Alpnach. Bars indicate the range of variation. The approximate fits (blue lines) are provided, where unit of km$^2$ and m$^2$ s$^{-1}$ are used for $A$ and $K_z$. The red line in (a) is the approximate fit provided by Maiss et al., (1994), and blue dash lines in (b) indicate 95% confidence level.
2.5 Conclusions

More than 600 microstructure casts were processed and analyzed to map and characterize $\varepsilon$ and $K_Z$ in Lake Erie. Casts were grouped based on water depth and seasonal stratification. The shallow regions (Group W) had highest average $\varepsilon$, with both $\varepsilon$ and $K_Z$ profiles showing less variation with depth compared to other groups. The presence of a seasonal thermocline during late summer, interacting with the lake bed (Group W) in the transition to the central basin, decreased $K_Z$ in the BBL. For the deeper Group C, there were sporadic turbulent patches, distributed beneath the high $\varepsilon$ and $K_Z$ in the surface layer. Basin-scale internal waves, during seasonal stratification, led to shear instability in metalimnion, increasing $\varepsilon$, but density stratification kept $K_Z$ near molecular diffusivity. The hypolimnion of Group C, during seasonal stratification, had high $\varepsilon$ and $K_Z$ because of thermocline entrainment and wind-induced seiche currents. Although data was limited in Group E, the mean $\varepsilon$ and $K_Z$ in metalimnion and hypolimnion were surprisingly high, possibly due to the Poincaré waves and hypolimnetic return flows set-up by strong wind events.

The lake morphometry, wind stress and baroclinic restoring force have significant effects on the TKE budget and vertical mixing in the lake. A larger surface area can not only increase the proportion of wind energy dissipating in the surface mixed layer, but also increase the vertical mixing in the lake interior. The transfer efficiency of wind energy, to be dissipated as TKE in the lake, decreased with increasing wind forcing. Values ranged between 3.5% and 0.49% under weak and strong wind forcing, respectively. In agreement with published observations in Lake Erie (Bouffard et al., 2012), during weak wind forcing, 1.3% of wind energy dissipated beneath the surface mixed layer, which is much higher than that in median- and small-sized lakes (Wuest et al., 2000). From upward casts after strong wind events, we found
more than half of the wind energy entering the lake was dissipated beneath the surface mixed layer. Integrating published observations, vertical mixing under the surface mixed layer decreases with increasing Wedderburn number, suggesting stratification is inhibiting buoyancy flux.

It is hoped that the data presented in this study can be used by other researchers, for example to compute biogeochemical fluxes in the lake and understand damping mechanisms of basin-scale motions, as well as to validate TKE dissipation in field-scale numerical models. Even though >600 casts were analyzed, many regions of Lake Erie still lacked observation and more measurements during winter time are required to complete year-round turbulence and mixing map of Lake Erie.
Chapter 3

Reynolds-averaged modelling of sediment resuspension in a large shallow lake

3.1 Introduction

Sediments and total suspended solid (TSS) are important biogeochemical components of aquatic systems, and their abundance influences water quality (Donohue and Molinos, 2009). For example, high TSS concentration can impact fish recruitment through egg and larval survival, feeding efficiency and predator avoidance (e.g., Bruton, 1985; Schallenberg and Burns, 2004; Donohue and Molinos, 2009; Tuttle-Raycroft et al., 2017). In shallow areas, an increase of TSS reduces light penetration affecting photosynthesis (Tilzer, 1983; Fréchette et al., 1989; Gloor et al., 1994) 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 (e.g., Liu and Huang, 2009; Graham et al., 2016; Valipour et al., 2017; Niu et al., 2018).

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 [Hawley, 2000; Churchill et al., 2004], Lake Michigan [Lou et al., 2000], and Lake Erie [Lick et al., 1994]). In the western basin and the nearshore zone of Lake Erie, for example, orbital velocities from wind waves extend to the bed, creating a wave boundary layer 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), in which computational models that
parameterize lake-wide sediment resuspension are more practical. Recently, Niu et al., (2018) simulated sediment resuspension and concomitant high surface layer turbidity events, which were consistent with measurements and satellite imagery. Their model, however, lacked near-bed observations needed for the validation of resuspension algorithms, requiring usage of literature-based coefficients (see also Marti and Imberger, 2008; Morales-Marín et al. 2017). The Reynolds-averaged Navier-Stokes (RANS) equation models, used in these studies, average resuspension dynamics over large grid cells (~ km), whereas resuspension is caused by turbulent bursting events at much smaller scales (~ cm to ~ m; Boegman and Ivey, 2009; Valipour et al 2017). Clearly there is a need to test the validity of employing observation-based resuspension model parameters in RANS models.

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, and basin-scale wave-induced currents in Lake Erie (León et al., 2005; Liu et al., 2014; Valipour et al., 2015b), and ELCOM-CAEDYM (hereafter ELCD) has been used to simulate temporal and spatial variability of phytoplankton and nutrients in Lake Erie (León et al., 2011). The ability of ELCD to simulate sediment resuspension, however, has not been examined. When calibrated against field observations, the model can be applied to provide a better understanding of sediment resuspension on a basin-scale.
3.2 Methods

3.2.1 Study area

Lake Erie is divided into three basins: western, central and eastern with maximum depths of about 16 m, 25 m, and 64 m, respectively (Figure 3.1). The shallowness of most of the western and central basins makes them very susceptible to the resuspension of sediments by wind-induced surface waves (Sheng and Lick, 1979; Hawley and Eadie, 2007; Valipour et al., 2017). The central and eastern basins of Lake Erie forms a seasonal thermocline, and in central Lake Erie, near-inertial (Poincaré) waves with a period of ~17 h are the dominant wind-induced internal waves during stratified period (Rao et al., 2008; Valipour et al., 2015b). The topographic features of Lake Erie are complex and the sediment types and grain sizes vary among the basins. Haltuch et al. (2000) presented the spatial distribution of the substrates in each basin, and noted the resuspendable sediment with grain size < 63 μm was the most prevalent substrate with the respective proportions of 51%, 61%, and 60% in the western, central, and eastern basins.

3.2.2 Field measurements

The variable used to validate the model is in situ TSS concentration, which was measured directly from pumped water samples (see Bouffard et al., 2013) and indirectly from a time-series recording using multi-parameter water quality sondes (XR-620 and XR-420; RBR Ltd), equipped with autoranging Seapoint turbidity and fluorescence sensors (accuracy ±2% measured values), YSI 6600 Sondes with wiped turbidity sensors (Yellow Springs Instruments, accuracy ±2% measured values) and beam amplitude (backscatter signal) from a Nortek Vector acoustic Doppler velocimeter (ADV; hereafter, ADV-amp, unit count; see Valipour et al., 2017). Herein, we compared model output to TSS concentration data (Sta. 341, 1227, 1228, 1231) in varying water depths, turbidity data (Sta. 341, 1228) near-bed and ADV-amp (Sta. 341) near-bed
The TSS concentration observations were used for quantitative comparisons, but due to limitations in their temporal and spatial coverage, TSS observations did not capture the timing of resuspension events. For continuous data from turbidity loggers and ADV-amp (e.g., Hawley and Zyren, 1990; Valipour et al., 2017), calibration is required to relate turbidity or ADV-amp to TSS, and unfortunately, this can only be achieved if the suspended particles in the water column are homogeneous and within a specific calibration range (Churchill et al., 2004). Consequently, we used the turbidity data and ADV-amp as qualitative indicators of resuspension events. Observations from Sta. 341 were the main dataset used to adjust the model parameters setup (see 3.3.1. Sensitivity tests).

Superficial bed sediment samples were collected at Sta. 341, 1231 (Figure 3.1b, d; Table B1) from PONAR grabs. 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 $\mu m$), 4% very fine sand (63 - 125 $\mu m$), 75% silt (3.9 – 63 $\mu m$), and 20% clay (0.06 - 4 $\mu m$).
Figure 3.1 (a) 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 Figures 3.8 (Curtain 1) and 3.10 (Curtain 2)). Depth contours are in meters. (c) West-to-east showing vertical grid (z-level) spacing in the model. Superficial sediments collected from PONAR grabs at (b) Sta. 1231 on 22 July, 2008 at 12:27 pm and (d) Sta. 341 on 31 July, 2008 at 8:43 am.

3.2.3 Model description

The three-dimensional hydrodynamic model ELCOM solves the unsteady RANS equations for incompressible flow using Boussinesq and hydrostatic approximations, and scalar (e.g., temperature, salinity, or tracer) transport equations (Appendix B, Table B2). An eddy-viscosity model is employed, in the horizontal, and a mixed layer model, in the vertical, for turbulence closure (Hodges et al., 2000). The free-surface evolution is governed by vertical integration of the continuity equation for incompressible flow in the water column applied to the kinematic boundary condition (Kowalik and Murty, 1993). ELCOM has been successfully applied to simulate the Lake Erie thermal stratification (Liu et al., 2014) and mean current circulation (León et al., 2005), water quality (León et al., 2011) and baroclinic and barotropic motions in the lake.
(Valipour et al., 2015b). In this study, modelled bottom mean currents, as one of the key drivers of sediment resuspension, showed good agreement with observations from an ADV deployed 1 m above bottom at Sta. 341 (Appendix B, Figure B2).

Inorganic particles (e.g., TSS) are modelled with the CAEDYM sediment module, by accounting for settling and resuspension, with advection and mixing provided by ELCOM (Hodges et al., 2000). In ELCOM, a conservative third-order scalar transport method (ULTIMATE QUICKEST) is applied. The equation for TSS is:

$$\frac{\partial TSS}{\partial t} + \frac{\partial}{\partial x} (U_x TSS) + \frac{\partial}{\partial y} (U_y TSS) + \frac{\partial}{\partial z} (U_z TSS) = \frac{\partial}{\partial x} (\kappa_x \frac{\partial TSS}{\partial x}) + \frac{\partial}{\partial y} (\kappa_y \frac{\partial TSS}{\partial y}) + \frac{\partial}{\partial z} (\kappa_z \frac{\partial TSS}{\partial z}) + S_c$$

where TSS is in mg L\(^{-1}\), \(U_x, U_y, U_z\) are eastward, northward, vertical components of mean current velocities (m s\(^{-1}\)), respectively. \(\kappa_x, \kappa_y, \kappa_z\) are eastward, northward, vertical components of eddy diffusivity (m\(^2\) s\(^{-1}\)), \(S_c\) is the turbulent Schmidt number, \(v_z\) is the settling velocity (m s\(^{-1}\)), calculated according to Stoke’s Law based on the user defined particle density (\(\rho_s, \text{kg m}^{-3}\)). The median sediment diameter is \(d_{50} \text{ (m)}\), \(\Delta z\) and \(\Delta z_{bot}\) are the thickness of the vertical and bottom layers (m), respectively, \(\alpha\) is the resuspension rate (g m\(^{-2}\) d\(^{-1}\)), \(SS_{sed} \text{ (g)}\) is the mass of all superficial sediments at bottom, and \(K_T \text{ (g)}\) is a parameter to control the ratio of resuspendible sediments to all superficial sediments (see also Mirbach and Lang, 2018). Even though the model can only set \(K_T\) as a fixed value, sediment composition varies across the lake. We do not have any measured or literature values for \(K_T\) and so this parameter was chosen such that the \(\alpha\) is within literature range, and the amount of sediment resuspended is also within a reasonable range. The
model assumed an infinite sediment pool in each bottom cell. The first term on the right of Eq. (3.2) represents settling and the second term parameterizes resuspension within the bottom layer, which occurs when the total bottom shear stress ($\tau_b$) exceeds the critical shear stress ($\tau_{cr}$). Within this term, $\tau_{ref}$ is a reference shear stress (usually set to 1 Pa).

The total bottom shear stress

$$\tau_b = \tau_c + \tau_w = \rho_w (u_{sw}^2 + u_{sc}^2)$$  \hspace{1cm} (3.3)

, where $\tau_c$ and $\tau_w$ are current-induced and surface wave-induced bottom shear stresses (Pa), which can be derived, respectively, from the current-induced and surface wave-induced shear velocity $u_{sc}$ and $u_{sw}$ (m s$^{-1}$), and $\rho_w$ is the water density (kg m$^{-3}$). Both $u_{sc}$ and $u_{sw}$ are computed from formulas summarized by van Rijn (1993):

$$u_{sc} = \sqrt{\frac{f_c U_{bot}^2}{8}}$$  \hspace{1cm} (3.4)

$$u_{sw} = \sqrt{0.5 f_w U_{orb}^2}$$  \hspace{1cm} (3.5)

, where $U_{bot} = \sqrt{(U_x^2 + U_y^2)}$ is the mean current velocity in the 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$^{-1}$) given by linear wave theory (Dean and Dalrymple, 1984)

$$U_{orb} = \frac{\pi H_s}{T_s \sinh(\frac{2\pi h}{L})}$$  \hspace{1cm} (3.6)

, where $h$ and $H_s$ are the local water depth (m) and significant wave height (m), $T_s$ is wave period (s), and $L$ is the local wave length (m). These wave field parameters are estimated from wind speed, fetch and depth (van Rijn, 1993; Barua, 2005; Appendix B, Table B3). 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

$$k_s = 2.5 \times d_{50}$$

(Englund and Hansen, 1972):

$$f_c = \frac{0.24}{[\log\left(\frac{12\Delta z_{bot}}{k_s}\right)]^2}$$

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

(3.7)  (3.8)

where the maximum wave amplitude (m), $a = \frac{H_s}{2\sin h\left(\frac{2h}{L}\right)}$ is calculated from linear wave theory

(Dean and Dalrymple, 1984). In Eq. (3.7), derived for hydraulically rough channel flow, $\Delta z_{bot}$ has been used to replace $h$, to allow for current-induced resuspension in deep systems, such as lakes. Herein, we investigated the validity of this assumption.

### 3.2.4 Model setup

The hydrodynamic model was configured based on Liu et al. (2014), including meteorological forcing, inflows, outflows, and the bathymetric grid. Different meteorological conditions were applied uniformly to model domain sub-sections across the lake: eastern (Port Colborne, data from Environment and Climate Change Canada (ECCC) station C45142), east-central (Port Stanley data from ECCC station C45132), west-central (meteorological data from buoy at Sta. 341), and western (NOAA [National Oceanic and Atmospheric Administration] SBIO1) basins. There were 11 inflows, including the Detroit River and Maumee River, and only one outflow, the Niagara River (see Table 2 in León et al. (2011)). To track the source of sediment loads, we separated TSS from rivers ($SS_R$) and the lake bed ($SS_B$) into different sediment pools ($TSS = SS_R + SS_B$). For $SS_R$, $d_{50} = 3 \mu m$ (Fukuda and Lick, 1980); whereas, $SS_B$ was divided into two classes: $SS_{B1}$ ($d_{50} = 1 \mu m$) represented clay-like superficial (nepheloid) sediments (Lick and Lick, 1988; Lick et al., 1994), and $SS_{B2}$ ($d_{50} = 10 \mu m$), represented the silt-like sediments at
depth (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 grabs (Figure 3.1b, d). Daily river flow rates were measured discharges from NOAA (Great Lakes Research Laboratory) or Water Survey of Canada (ECCC) databases (León et al., 2011), and SS$_R$ data for the Maumee River was from the USGS (US Geological Survey; https://cida.usgs.gov/sediment/). Sediment load from the Maumee River was derived from SS$_R$ by multiplying by flow rate (Appendix B, Figure B1). Due to a lack of observations, the SS$_R$ in the Detroit River was set by apportioning the annual sediment load according to the daily flow rate (Kemp et al., 1977).

The horizontal grid was 2 km $\times$ 2 km, and the lake was divided into 45 vertical layers, with a finer 0.5 m grid near the surface and through the thermocline, and coarser 5m grid in the deep (~65 m) eastern basin (Figure 3.1c). The time step was 5 min, to satisfy the Courant-Friedrichs-Levy condition. We initialized the model using observed water temperature profiles throughout the lake from a spring survey conducted by U.S. Environmental Protection Agency (U.S. EPA); 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., 2015b). The first TSS observations (at 7 field stations, Figure 3.1a), which were collected on 21 July 2008 (day 203), were used to initiate the TSS concentration in the model and the remaining data (from field Sta. 341, 1227, 1228, 1231) were used for model quantitative calibration and validation. Available model forcing data (river flows) terminated on day 303, so the model was run from 21 July 2008 (day 203) through 29 October 2008 (day 303), which spans both seasonal stratified and non-stratified periods. We define $T_t$ as the entire simulation time (99.52 d) and $T_R$ as the duration of resuspension events (i.e., the time when $\tau_t > \tau_{cr}$) so that the frequency of resuspension events in varying regions can be studied.
3.3 Model calibration and validation

3.3.1 Sensitivity tests

TSS in the water column, away from tributaries, is determined by a balance between resuspension and settling. In sensitivity tests, we adjusted the relevant model parameters to achieve a best-fit against ADV-amp, turbidity data and the observed TSS concentration. Given that \( \tau_{cr} \), \( d_{50} \), and \( \alpha \) are three primary parameters that determine resuspension (Eq. (3.2), (3.7-8)) and \( d_{50} \) was set to observed in situ values, we adjusted \( \tau_{cr} \) and \( \alpha \) according to ranges found in the literature (Table 3.1) to achieve the best-fit model result. Broad ranges of both \( \tau_{cr} \) and \( \alpha \) have been reported. Lick et al. (1994) suggested that \( \tau_{cr} \) for freshly deposited sediments and older sediments could vary by 1 order of magnitude, depending on the sediment composition and water content, and \( \alpha \) could vary by over 2 orders of magnitude (Fukuda and Lick, 1980). Thus, we expected \( \tau_{cr} \) and \( \alpha \) to be within the range of 0.01 – 0.3 Pa, and 31.1 – 864 g m\(^{-2}\) d\(^{-1}\), respectively, in Lake Erie (Table 3.1).

Our initial estimate of critical shear stress came from Valipour et al. (2017), who analyzed the same observation dataset. They observed near-bed high turbidity events at Sta. 341 when maximum instantaneous burst velocity \( u_{max} = 0.25 \text{ m s}^{-1} \), corresponding to a 5 min-averaged mean flow velocity \( u_{mean} = 0.1 \text{ m s}^{-1} \) \( (\tau_{max} = \rho C_D u_{max}^2 = 0.28 \text{ Pa}; \tau_{mean} = \rho C_D u_{mean}^2 = 0.045 \text{ Pa}, \text{ where } C_D = 4.5 \times 10^{-3} \text{ (Valipour et al., 2015)}) \). But when we applied \( \tau_{cr} = 0.28 \text{ Pa} \), resuspension was not modelled. We hypothesized that bursts events, which lift sediments from the bed, would be filtered through Reynolds-averaging over our 2 km x 2 km x (0.5~5) m grid cells, and consequently the modelled flow velocities would be lower than the instantaneous burst-type ejections that resuspend sediment (Aghsae and Boegman, 2015; Valipour et al.,
Therefore, it was not necessarily appropriate to use the observed 0.28 Pa as the threshold for resuspension. Validation of modeled 5 min-averaged mean flow velocity can be found in Appendix B (Figure B2) and so we set $\tau_{cr} = 0.045$ Pa, as the baseline for model calibration.

**Table 3.1** Comparison of published observations of $d_{50}$, $\tau_{cr}$ and $\alpha$ in Lake Erie and other locations.

<table>
<thead>
<tr>
<th>Study area</th>
<th>$d_{50}$ (µm)</th>
<th>$\tau_{cr}$ (Pa)</th>
<th>Resuspension rate $\alpha$ (g m$^{-2}$ d$^{-1}$)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Lake Erie$^a$</td>
<td>1, 10</td>
<td>0.01; 0.025</td>
<td>450</td>
</tr>
<tr>
<td>Lake Erie$^b$</td>
<td>10</td>
<td>0.045-0.28*</td>
<td>N/A</td>
</tr>
<tr>
<td>Lake Erie$^c$</td>
<td>10</td>
<td>0.03-0.3</td>
<td>N/A</td>
</tr>
<tr>
<td>Lake Erie$^d$</td>
<td>N/A</td>
<td>0.01-0.1</td>
<td>N/A</td>
</tr>
<tr>
<td>Lake Erie$^e$</td>
<td>N/A</td>
<td>0.1</td>
<td>N/A</td>
</tr>
<tr>
<td>Lake Erie$^f$</td>
<td>10</td>
<td>0.1-0.2</td>
<td>32.8-864</td>
</tr>
<tr>
<td>Lake Erie$^g$</td>
<td>N/A</td>
<td>0.05</td>
<td>31.1</td>
</tr>
<tr>
<td>Lake Erie$^h$</td>
<td>4</td>
<td>0.05-5</td>
<td>0.864-311</td>
</tr>
<tr>
<td>Lake Michigan$^i$</td>
<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 Superior$^k$</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>


* $\tau_{cr}$ was modified based on a correlation between instantaneous burst currents and burst-average currents (see text for further explanation).
Table 3.2  Simulated parameter combinations, with RMSE and percent bias between TSS concentration from the model for the parameter combinations and from water samples collected during days 203-220 at Sta. 1227, 1228, 341, and 1231.

<table>
<thead>
<tr>
<th>Combination</th>
<th>$d_{50}$ (µm)</th>
<th>$\tau_{cr}$ (Pa)</th>
<th>$\alpha$ (g m$^{-2}$ d$^{-1}$)</th>
<th>RMSE in overall water column (mg L$^{-1}$)</th>
<th>RMSE in bottom layer (mg L$^{-1}$)</th>
<th>Percent bias in water column (%)</th>
<th>Percent bias in bottom layer (%)</th>
</tr>
</thead>
<tbody>
<tr>
<td>(A)</td>
<td>1 (SS$<em>{B1}$); 0.045 (SS$</em>{B1}$); 450</td>
<td>0.63</td>
<td>0.74</td>
<td>13</td>
<td>10</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>10(SS$_{B2}$)</td>
<td>0.045 (SS$_{B2}$)</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>(B)</td>
<td>1 (SS$<em>{B1}$); 0.03 (SS$</em>{B1}$); 450</td>
<td>0.64</td>
<td>0.74</td>
<td>15</td>
<td>11</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>10(SS$_{B2}$)</td>
<td>0.045 (SS$_{B2}$)</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>(C)</td>
<td>1 (SS$<em>{B1}$); 0.01 (SS$</em>{B1}$); 450</td>
<td>0.81</td>
<td>0.74</td>
<td>33</td>
<td>36</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>10(SS$_{B2}$)</td>
<td>0.025 (SS$_{B2}$)</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>(D)</td>
<td>1 (SS$<em>{B1}$); 0.01 (SS$</em>{B1}$); 300</td>
<td>0.73</td>
<td>0.74</td>
<td>27</td>
<td>25</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>10(SS$_{B2}$)</td>
<td>0.025 (SS$_{B2}$)</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>(E)</td>
<td>1 (SS$<em>{B1}$); 0.03 (SS$</em>{B1}$); 300</td>
<td>0.63</td>
<td>0.74</td>
<td>11</td>
<td>13</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>10(SS$_{B2}$)</td>
<td>0.045 (SS$_{B2}$)</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

Five different parameter combinations (Tables 3.1 and 3.2) were considered in a sensitivity analysis. (A) was the baseline, with $\tau_{cr} = 0.045$ Pa for both sediment classes and $\alpha = 450$ g m$^{-2}$ d$^{-1}$ for the median of the observations (Table 3.1). The results (Figure 3.2A) showed that combination (A) only captured resuspension events on days 250 and 259 but missed the resuspension events on days 224 and 239-242. The TSS was simulated to settle too fast from the water column, qualitatively inconsistent with ADV-amp and turbidity data. Combination (B) had a lower $\tau_{cr}$ for SS$_{B1}$, but also missed events on day 224 and showed little resuspension on days 239 - 242. We further decreased $\tau_{cr}$ based on the 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 recorded by ADV-amp and turbidity data (i.e. day 224, 239-242, 250, 259). By adjusting $\alpha$, within the observed range, we were able to control the
amount of bed sediment entrained into the water column, resulting in a better fit of the model to
the ADV-amp and turbidity observations (Figure 3.2C, D). The Root-mean-square-error (RMSE)
between the TSS concentration in the bottom layer, from these five combinations, and from
water samples was 0.6 – 0.8 mg L\(^{-1}\), around one third of the background ambient TSS
(\(~ 2\) mg L\(^{-1}\)) (Table 3.2), and did not vary over the parametric study.

Our visual best-fit analysis above did not agree with percent bias (Table 3.2), which was
lower for combination (A), (B) and (E) than (C) and (D). However, this does not indicate
superior model performance with the former combinations, as their TSS time-series showed little
variation, remaining near the mean of the TSS observations, resulting in low percent bias but
also poor model performance. On the other hand (C) and (D), with higher percent bias, captured
minor oscillations in observed TSS concentration. Unfortunately, the TSS observations were
infrequent and in fair weather, when these was no storm events to drive strong sediment
resuspension, limiting their utilities for model validation beyond the background concentration.
Priority, in calibration, was therefore given to the ability to reproduce both background and
timing of large resuspension events, resulting in combination (C) giving the overall best fit as the
calibrated parameter set for the model.
Figure 3.2 Time-series at Sta. 341 of model output TSS for parameter combinations (A–E) in Table 3.2, and averaged backscatter signal amplitude (ADV-amp) 1 m above bottom. Orange circles are the observed TSS concentration from water samples. Among them, combination (C) gave the visual best-fit result. Panels (F) and (G) are time-series at Sta. 341 of turbidity and Chl-a observations recording from XR-420, 620 with sampling interval 2 and 3 min, respectively.

3.3.2 Effects of algae and particle aggregation

We applied the calibrated model to further test its performance against observed data. Model comparisons against turbidity data at Sta. 1228 (around 10 km away from Sta. 341) showed good agreements in simulating resuspension events on days 240, 245-250 and 260, except continuous high turbidity events after day 260 (Figure 3.3). Settling of algae from summer blooms or sediment advected from shallower sites could be the source of these high turbidity values (Valipour et al., 2017). At Sta. 341, turbidity was continuously high during days 265-280, which was not simulated by model. Moreover, during days 255-280, the turbidity 5 m above bottom was unusually high compared to that in bottom layer (Figure 3.2F). During this
period, Chl-a concentration spiked to > 2.5 μg L⁻¹, indicating the existence of algae (Figure 3.2G). Wynne et al. (2010) reported that the algae bloom in 2008 started on day 243 and persisted for over 2 months in the western basin, with areas of bloom extending to west central basin (Sta. 341 and 1228), possibly causing high turbidity after day 260.

Fine particles, like clay, which occupy 20% of sediment on the lake bed, tend to form aggregates that accelerate their settling velocity (Hawley, 1982). But the CAEDYM model is unable to account for sediment cohesiveness and aggregation. Therefore, the computed settling velocity, based on Stokes’ Law, may significantly underestimate the sedimentation of cohesive particles, possibly causing spurious high TSS after resuspension events (e.g., days 242-245, 250-255 at Sta. 341 (Figure 3.2C)).

Although there were some discrepancies between model results and observations, the model simulated both the background TSS concentration (Table 3.2) and qualitative (Figure 3.2, 3.3) nature of the sediment resuspension. Coupled with the hydrodynamic information output by model, we further investigated the basin-scale sediment resuspension and mechanisms behind it.
**Figure 3.3** Time-series at Sta. 1228 of (a) model output TSS 1 m above bottom and (b) turbidity data from YSI 6600 sampled at 1 m above bottom with 1 h sampling interval.

**Figure 3.4** Time-series of (a) observed wind speed at station 45005 (NDBC), and (b) observed wave height at buoy 45005 (NDBC), modelled wave height and modelled wave period at Sta. 341. Observation data was 1-h averaged. The correlation coefficient ($R^2$) between modelled and observed wave height is 0.70, and root-mean-square-error (RMSE) is 0.28 m.
3.4 Result and Discussion

3.4.1 Temporal change of thermal structure and TSS concentration

The modelled temperature fields have been validated for the present 2008 setup (Liu et al., 2014; Valipour et al., 2015b) and other years (e.g., León et al., 2005; Oveisy et al., 2014) in Lake Erie. We examined the relationships between modelled temperature and TSS profiles (Figure 3.5) in the western, west-central, central and eastern basins (represented by Sta. 357, 341, 84 and 452, respectively). In general, the seasonal pattern showed high TSS during fall; suggesting the more frequent storm events drove resuspension, which was strongest in the shallow western basin (Figure 3.5a, b). In the central basin, TSS concentrations were low during summer stratification, with wind energy potentially energizing internal Poincare waves (Valipour et al., 2015b). The exception was when the thermocline intersected the bed and resuspension was modelled (Figure 3.5c-f). In the eastern basin, the model results suggested that resuspension occurred in nearshore regions and suspended solids were advected offshore to Sta. 452 (Figure 3.5g, h; Lou et al., 2000).

In the western basin, (Sta. 357; Figure 3.5a, b), the water column was well mixed and the modelled 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 agreement with observations at Sta. 341 (Valipour et al., 2017).

In the west central basin, with a stratified water column (Boegman et al., 2008), the TSS concentration remained low in the stratified period (Sta. 341; Figure 3.5c, d). The significant wave height did not exceed 1.2 m until day 235, and the associated surface wave energy could not extend down to or below the thermocline. As the thermocline deepened to the lake bed in September (around day 250), there was sediment resuspension at the bottom (around day 259),
which resulted from bottom currents likely induced by the internal Poincare waves (see 3.4.4). These resuspended sediments remained in the hypolimnion when the water column re-stratified through up-welling, showing the effect of stratification on the vertical distribution of TSS (Figure 3.5c, d). After day 270, during weak stratification, high TSS concentrations from the bottom were transported into the surface layer, during two resuspension events (days 291, 300-303).

In the central and eastern basins, the TSS concentrations were much lower compared to the shallower sites. At the deeper sites (Sta. 84 and 452), TSS concentrations were high near the surface (Figure 3.5f, h), indicating that suspended solids with low setting velocity ($SS_{B1}$) were likely advected from shallower parts of the lake, including the west central basin, the Grand river or the internal swash zone at the depth of the thermocline in the north central basin (see 3.4.6). This advection mechanism was described in detail in Appendix B (Text B1). TSS profiles at Sta. 84 and 452 did not show obvious bottom resuspension events during the simulation.
Figure 3.5 Time-series of modelled temperature and TSS concentration profiles in the western (Sta. 357; panels (a) and (b)), west-central (Sta. 341; panels (c) and (d)), central (Sta. 84; panels (e) and (f)) and eastern (Sta. 452; panels (g) and (h)) basins. Note the change in scale of color bar between (b), (d) and (f), (h). The data missing in the surface layer of (a – f) is due to the change of water level.

3.4.2 Near-bed TSS concentration

To visualize the lake-wide near-bed TSS concentration, we extracted maps of the bottom layer at selected times (Figure 3.6). In general, the TSS concentration in the lake was predicted to increase from summer through fall as the thermocline deepened and the intensity of the daily winds increased (Figure 3.4). The TSS at the bottom of the western basin was higher than the other two basins, and bottom sediment in the west central basin began to resuspend when the thermocline moved eastward through seasonal deepening. This suggests that the absence of the thermocline, in this region, allowed for a more efficient transfer of wind energy from the surface to the bottom (Hawley and Lee, 1999). The model also predicted high TSS concentrations near the northern shoreline of the central basin (Figure 3.6b-f) where the thermocline intersected the
lake bed, indicating sediment resuspension in the internal swash zone related to baroclinic motion, that likely generated the modelled nepheloid layers (See 3.4.6.2). There are no tributaries in this region and shoreline erosion was not modelled.

The strong storm on 29 October (day 303; Figure 3.6f) caused sediment resuspension in both the western and central basins and thus, dramatically increased the near-bed TSS. High TSS was also modelled to occur along the western flank of Long Point, suggesting bottom resuspension may be reworking bed material and supplying sediment (i.e. resuspension-transportation-resettling processes) to the 40-km long sand spit (Hawley and Eadie, 2007). Further details of the modelled transport, in this region, are given in Appendix B (Figure B4).

Our results differ from Niu et al. (2018), who argued western basin high turbidity events occurred during fall, and not during summer (Figure 3.6a-c). Their model was calibrated with satellite images and near-surface TSS observations near the Detroit and Maumee River mouths, and was not able to validate the occurrence of near-bed resuspension. Relying on published resuspension parameters, their resuspension threshold was chosen to be ~2-5 times that in the present study, which was based on our sensitivity tests. This may lead to their model underestimating TSS near the bed.

Figure 3.6 also indicates the regions with active resuspension processes: western, west-central and north-central basins. These regions are likely to have enhanced sediment-water-interface nutrient exchange and sediment oxygen demand, thus affecting local biogeochemistry (Moriarty et al., 2017; Paytan et al., 2017; Valipour et al., 2019).
Figure 3.6 Model output (24-h average) showing TSS in the bottom layer on: (a) 12 Aug (day 224); (b) 27 Aug (day 240); (c) 6 Sep (day 250); (d) 15 Sep (day 259); (e) 19 Oct (day 293); and (f) 29 Oct (day 303). The dates were chosen based on resuspension in Figure 2b. The black lines are isotherm contours increasing from 16 °C at 2 °C intervals toward the lake perimeter in (a-d). These indicate where the thermocline intersected the lake bed; (e) only shows the 16 °C and contour, and the lake was colder than 16 °C in (f).

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

By simulating $SS_R$ (Appendix B, Figure B3), we were able to compare the relative contribution of resuspension (internal loading; $SS_B$) and river inputs (external loading; $SS_R$) to TSS. The $SS_R$ concentrations were consistent, relative to $SS_B$ from storm events (Figure B3f). High $SS_R$ ($\sim 5$ mg L$^{-1}$) was modelled at the Maumee (Figure B3d, e) and Detroit (Figure B3a-f) river mouths, indicating that tributary loading was a significant TSS source in the western basin (Bolsenga and Herdendorf, 1993; Binding et al, 2012). Niu et al. (2018) showed that the duration and area of influence associated with riverine turbidity was negligible during fall, but during summer ~90% of turbidity events were from river loads. Our modelled riverine TSS
indicated that SS$_R$ contributed ~50% (Figure 3.6a-c, Figure B3a-c) and ~25% (Figure 3.6d-f, Figure B3d-f) of the TSS concentration near the bottom during summer and fall, respectively. The differences between these two models may result from different periods of analysis, different model parameters and/or lower inputs from the 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 input of sediments in the entire lake. The Maumee River has 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 (Figure B1), but because of the low discharge, only led to a high concentration near the river mouth (Figure B3d). The Detroit River discharge does not have 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 loads, during late summer and fall, the Maumee River provides most of the nutrients to the western basin (e.g., Schwab et al., 2009; Bridgeman et al., 2012). Images showing the fate of river inputs can be found in Appendix B (Movie B1).

3.4.4 Bottom shear stress

The model-computed $\tau_b$, $\tau_c$ and $\tau_w$ (Eq. (3.3), (3.4) and (3.5)) were investigated to understand the relative contributions of surface waves and mean currents to resuspension. In the western basin (Sta. 357; Figure 3.7a), $\tau_w > \tau_{cr}$ during most of resuspension events and thus, was likely the main driving force (e.g., days 240, 260, and ~300). Because of the shallowness of the western basin, a large amount of energy input from wind-induced surface waves penetrated into
the bottom layer, triggering resuspension events. After day 270, $\tau_w$ exceeded the critical threshold more frequently, as storm events increase during fall (Figure 3.4a, Figure 3.7a), and $\tau_w$ was approximately one order of magnitude higher than $\tau_c$. In the western basin, periodic currents were sometimes related to storm events. For example, on days 259 and 298 (Figure 3.7a), $\tau_c > \tau_{cr}$ with significantly high $\tau_w (> 0.1 \text{ Pa})$, indicating storm events not only induced surface waves, but mean currents that triggered resuspension.

Conversely $\tau_c$ and $\tau_w$ in the west-central basin (Sta. 341, Figure 3.7b) were comparable (Table 3.3), but the magnitude of $\tau_b$ was not as large as that in western basin, due to the greater water depth. The resuspension events on days 240, 250, 259 and 300 were caused mainly by increased $\tau_w$. On these days, wave heights $\geq 1.1 \text{ m}$ and wave periods $\geq 4.5 \text{ s}$ (Figure 3.4). After day 290, $\tau_b$ increased dramatically due to higher $\tau_w$ from wave heights exceeding 1.5 m, explaining the high turbidity event (Figure 3.2b). These simulation results were in agreement with observations (Valipour et al., 2017), which concluded that surface waves in Lake Erie with significant periods ($T_s \geq 5 \text{ s}$) and significant heights ($H_s \geq 1.5 \text{ m}$) were able to resuspend bed material at Sta. 341. Near the end of the simulation (day 300), when the strongest storm event during the simulation occurred (Figures 3.4 and 3.8), resuspension events resulted from not only surface waves, but also near-bed currents, revealing a complex resuspension mechanism, different from the western basin.

The intensity of $\tau_w$ was negligible in the deep region of central basin (Sta. 84), and $\tau_b$ did not exceed $\tau_{cr}$ most of the time, except during the day 300 storm event. This showed that resuspension events were uncommon in the deep regions of the central basin, and high TSS in the hypolimnion of the central basin (Figure 3.5f) was not from local resuspension but transport.
from shallower sites or due to settling in the water column.

Valipour et al. (2017) identified various factors which contribute to high-turbidity events at Sta. 341 during the summers of 2008-2009 and, among 12 recorded high turbidity events, the contribution from surface waves was over three times that from the mean currents. Conversely, our modelled bottom shear stress showed that the contributions of resuspension from surfaces waves and mean currents were comparable at Sta. 341 (Figure 3.7b). This discrepancy may result from more storm-induced mean currents during our late summer through fall simulations, compared to the spring through fall observations from Valipour et al. (2017) (Figure 3.4).

The spatial patterns of modelled $\tau_c$ and $\tau_w$ at Sta. 357, 341 and 84 (Table 3.3) indicated that the frequency of resuspension events and the contribution from $\tau_w$ both decreased with water depth. Conversely, storm-induced mean currents were of increased importance at deeper stations (Sta. 341 and 84) in the central basin. In the west central basin (Sta. 341), the $\tau_b$ contribution from mean currents was greatest (48%), compared to the other two stations (12% at Sta. 357; 32% at Sta. 84; Table 3.3). The reason for the nonlinear change in contribution from $\tau_c$ with water depth was because at intermediate water depths (~15 m), baroclinic and storm-induced bottom currents are relatively stronger than at deeper sites (>20 m), during both the stratified and mixing periods. Resuspension only occurred in the middle of the central basin (Sta. 84) when there was a strong storm event (day 300).
Figure 3.7 Time-series of bottom shear stress in Lake Erie indicating \( \tau_b \) (total bottom shear stress, Eq. 3.3), \( \tau_{cr} \) (critical shear stress), \( \tau_c \) (current-induced bottom shear stress, Eq. 3.4), and \( \tau_w \) (surface wave-induced bottom shear stress, Eq. 3.5) at (a) Sta. 357 (~12 m), (b) Sta. 341 (~17.5 m), (c) Sta. 84 (~25 m).

Table 3.3 Contribution of mean currents and surface waves to sediment resuspension at western (Sta. 357), west-central (Sta. 341), and central (Sta. 84) basin stations during the simulation (days 203-303). The averaged ratio of \( \tau_w \) and \( \tau_c \) over \( \tau_b \) when \( \tau_b > \tau_{cr} = 0.25 \text{ Pa} \) were used to calculate \( \tau_w : \tau_b \) and \( \tau_c : \tau_b \).

<table>
<thead>
<tr>
<th></th>
<th>Sta. 357</th>
<th>Sta. 341</th>
<th>Sta. 84</th>
</tr>
</thead>
<tbody>
<tr>
<td>( \tau_c : \tau_b )</td>
<td>12%</td>
<td>48%</td>
<td>32%</td>
</tr>
<tr>
<td>( \tau_w : \tau_b )</td>
<td>88%</td>
<td>52%</td>
<td>68%</td>
</tr>
<tr>
<td>( T_b : T_t )</td>
<td>8.2 d (8.2%)</td>
<td>5.6 d (5.6%)</td>
<td>0.95 d (0.95%)</td>
</tr>
<tr>
<td>Contribution from ( \tau_c )</td>
<td>0.98%</td>
<td>2.7%</td>
<td>0.3%</td>
</tr>
<tr>
<td>Contribution from ( \tau_w )</td>
<td>7.2%</td>
<td>2.9%</td>
<td>0.65%</td>
</tr>
</tbody>
</table>

65
3.4.5 Surface wave-induced resuspension

Storm events transfer wind momentum to the surface layer, which is the main energy source for resuspension (Hawley, 2000; Hawley et al. 2004; Chung et al. 2009). Valipour et al. (2017) reported that in Lake Erie, surface waves were the major abiotic driving forces for high turbidity events at Sta. 341 during the summers of 2008-09. The bottom shear stress analysis presented above for the modelled stations (Figure 3.7) also demonstrated that surface waves were the dominant mechanisms driving resuspension in the western basin and central basin. To extend this analysis to the entire lake, we showed the modelled sediment response in the water column of the western and west-central basins, as well as the bottom layer of the entire lake during three fall storm events (Figure 3.8). During fall 2008, the water column in the western basin was well mixed (Figure 3.5a, b). The storm event resuspended sediment through the water column to the surface layer (Figure 3.8a, c, e). In the deeper west-central basin, there was no obvious increase in TSS during the event on day 241 (Figure 3.8a); whereas, resuspension occurred here during the event on day 259, and the intensity of resuspension was relatively weak compared to the western basin (Figure 3.8c, e). The most intense response, in both the western and west-central basins, occurred during the storm event on day 300; when surface wave energy penetrated to 20 m depth, and sediments were lifted from the bottom of the west-central basin to the surface.

In addition to the western and west central basins, the northern shoreline of the central basin and Long Point also had resuspension responses to these three storms, especially on day 300 (Figure 3.8d, f). Regions with water depths > 20 m (e.g., the central and southern central basin and most of the eastern basin), had little resuspension, even during the strongest storm (day 300; Figure 3.8f). Bottom sediments near the northern shoreline of the central basin and Long Point consist of glacial till, and sand/gravel, respectively (Haltuch et al., 2000). Glacial till and sand
can be partially resuspended, but gravel is unable to resuspend; hence, our modelled sediment diameters may be inappropriate in regions with gravel. Or these regions offshore of Long Point may have remnant gravel substrates as a result of scour. Most of the mid and southern central basin has mud substrate, which can be resuspended, so the water depth and hydrodynamic forces are the main factors that limit resuspension. The model did not produce resuspension in this region, suggesting the local bottom boundary layer is not dynamically active.

**Figure 3.8** Daily-averaged TSS concentration in the western and west-central basin for a western basin curtain (panels a, c and e) and a bottom layer sheet (panels b, d and f) during storm events on: (a, b) day 241, wind speed =9.8 m s\(^{-1}\), wave height =1.4 m; (c, d) day 259, wind speed =12.77 m s\(^{-1}\), wave height =1.2 m; and (e, f) day 300, wind speed =15.9 m s\(^{-1}\), wave height = 1.77 m.

### 3.4.6 Thermocline motion and current-induced resuspension

Bottom currents associated with vertical thermocline motions including up-welling and down-\(\downarrow\)welling events, and near-inertial internal waves had been found to be correlated with near-bed high turbidity events in Lake Michigan (Hawley and Muzzi, 2003; Hawley, 2004). Gloor et al.
(1994) showed similar observations from Lake Alpnach, but high-frequency burst-like currents resuspended bottom particles more efficiently than mean currents from internal seiche motions. Observations from Valipour et al. (2017) showed that the contribution to resuspension from near-bed turbulence, induced by high-frequency internal waves (HFIW), on high turbidity events in Lake Erie was similar to that from mean currents. While our RANS model does not parameterize sub-grid-scale turbulent bursts, we also found motions of the thermocline, in Lake Erie, to have an effect on the distribution of TSS and resuspension events. From the model output at varying locations and water depths (Figure 3.9), we modelled three mechanisms of thermocline-related resuspension.

3.4.6.1 Resuspension from down-welling

At locations with different depths (Figure 3.9a-c), the thermocline downwelled to the lake bed, leading to a sharp increase of $\tau_c$ and modelled spikes in TSS (Figure 3.9d-f, D1-6). With the downwelling motions, $\tau_c > \tau_{cr}$ (Eq. (3.4)) indicated that mean current-induced bottom shear stress was enough to drive resuspension. The spatial dynamics of resuspension from downwelling (Figure 3.10c), is shown for the north shore of the central basin where the thermocline intersects the bottom slope.

3.4.6.2 Resuspension from up-welling

Thermocline swashing was modelled near the northern shoreline of the central basin (Sta. 2586; Figure 3.9e, f; U1, U2) during storm events (Figure 3.5, days 250, 260) to compress the bottom layer. Cold hypolimnion water, squeezed from this location (Figure 3.9c), upwelled when the wind stopped under the effect of buoyancy, inducing strong $\tau_c$ (Figure 3.9f) and triggering resuspension. On day 259 (Figure 3.10a), upwelling along the bottom and northern flank of the central basin was modelled, corresponding with resuspension events at the bottom of the central
basin, which was in agreement with our assumption that resuspension was caused by upwelling of cold bottom water after a storm. It also agreed 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 induced resuspension only occurred along the northern shore, with upwelling events near the surface and downwelling events at depth (Figure 3.10). These two mechanisms generated internal swash zones, lifting bottom sediments and forming a turbid intrusion layer in the metalimnion (Figure 3.10a-c). Similar results were observed in Lake Kinneret (Marti and Imberger, 2008). Conversely, the southern flank was relatively quiescent during these events. We suspected this pattern resulted from predominant southwest winds, in summer, having a longer fetch to the northern shore, which pushed the thermocline down at the windward shore, relative to rise at the leeward shore (Monismith, 1986), resulting in more intense up- and down-welling events, creating stronger bottom currents and near-bed turbulence.

3.4.6.3 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., 2015b). Wavelet analysis on modelled temperature at Sta. 1228, showed energy peaks at ~17 h during days 205-240 (Appendix B, Figure B5), indicating the existence of Poincaré waves. At Sta. 1228 (Figure 3.9b, e, PW1), during the Poincaré wave signals, the baroclinic currents below the thermocline became stronger by conservation of volume, leading to strong \( \tau_c \), exceeding \( \tau_{cr} \) and causing resuspension of bottom sediments.
Hawley (2004) showed the correlation between near-inertial waves and the vertical distribution of sediment in Lake Michigan, indicating 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 Sta. 341 in the summer of 2009. Even though our model was unable to simulate sub-grid scale HFIWs, we suspected that mean currents generated by Poincaré waves, which can be captured by the model, contribute to the resuspension events.

Resuspension induced by Poincaré waves was less intense, compared to upwelling and downwelling mechanisms, and resuspension did not occur all of the time. 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 motions of the thermocline could cause more intensive resuspension (Hawley, 2004; Puig et al., 2001).

To investigate spatial distributions of bottom stress between stations, the lake-wide distribution of $\tau_c$ was computed from Eq. (3.4) and averaged over the simulation, showing ‘hot spots’ of enhanced mean current-induced resuspension near the Detroit River mouth, west central basin, northern shoreline of central basin, and northwest corner of the eastern basin (Figure B6). The model was unable to output $\tau_w$ and $\tau_b$ at more than one location per model run, and so plotting distributions of $\tau_w$ and $\tau_b$ is not currently feasible.
Figure 3.9 Time-series of (a-c) modelled temperature profiles at Sta. 1227, 1228, and 2586; (b-f) corresponding modelled 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).
Figure 3.10 Model output central basin curtain (see Figure 3.1) showing (a) up-welling (day 259, U1), (b) day 266 and (c) down-welling (day 271, D6) along the northern shore of the central basin. The color bar shows TSS concentration and the black lines are isotherms contours through the metalimnion from 14°C (bottom) and increasing at 2°C intervals. (d) shows $\tau_c$ (current-induced bottom shear stress) in these three days along the bottom of this curtain.

3.5 Conclusions

We found that to simulate sediment resuspension with a RANS model, resuspension parameters should be set according to *in situ* measurements, and also based on sensitivity tests against resuspension observations. We believed this was because resuspension was a highly localized and sub-grid-scale process. Differences will exist between the instantaneous and RANS modelled flow fields, as well as actual versus parameterized sediment characteristics and
associated resuspension algorithms; therefore, it is not expected that calibrated model parameters match those observed in situ.

Using turbidity, ADV-amp and TSS observations, we validated ELCD simulations of sediment resuspension in Lake Erie. The model was able to simulate the background near-bed TSS concentration with reasonable error (RMSE = 0.8mg L$^{-1}$) and predict resuspension events indicated by turbidity and ADV-amp. Several factors may contribute to inaccuracy in the model result, which include: particle cohesiveness, spatial variations in sediment composition, the effect of local sub-grid-scale bathymetry and turbulence, etc.

The model results were analyzed to identify mechanisms leading to resuspension and we found resuspension events in the western and central basins were both mainly triggered by surface waves, but the contribution of $\tau_w$ decreased with water depth. In the west central basin, the effect of mean currents on resuspension was comparable to that of surface waves. The existence of a stable thermocline during summer prevented resuspended sediment from being transported into the upper water column. However, vertical motions of thermocline (i.e. up- and down-welling events) and Poincaré wave-induced bottom currents, caused resuspension in the central basin, potentially forming internal swash zones along the northern shoreline with subsequent turbid intrusion layers in the metalimnion.

Future work should improve model performance by including better spatial distribution of sediment types, and shoreline erosion. Also, the ability to output bottom shear stress-related parameters as global variables would allow for improved analysis of model results.
Chapter 4
Measurement and Reynolds-averaged modelling of bottom shear stress and sediment resuspension parameterizations

4.1 Introduction

Sediment resuspension, in shallow lakes and nearshore regions, can contribute significantly to the concentration of total suspended solids (TSS), which is an important biogeochemical component of aquatic systems that regulates light extinction, nutrient flux and predator/prey relationships (e.g., Bruton, 1985; Donohue and Molinos, 2009; Valipour et al., 2017). Bottom shear stress ($\tau_b$) drives resuspension and is, therefore, a link between hydrodynamic forcing and water quality (e.g., Kim et al., 2000; Biron et al., 2004; Salim et al., 2018). Resuspension in the bottom boundary layer (BBL) occurs when $\tau_b$ is sufficient at the sediment water interface, so that sediment motion is initiated (bedload transport) and turbulent eddies induce vertical velocity components that exceed the particle fall velocity (Bagnold, 1966; van Rijn, 1993). $\tau_b$ is defined as a combination of viscous stress ($\tau_v$) and Reynolds stress ($\tau_R$),

$$\tau_b = \tau_v + \tau_R = (\rho \nu \frac{\partial U}{\partial z} - \rho u'w')_{z=0} \tag{4.1}$$

Here, the instantaneous horizontal velocity $u = U + u'$ is Reynolds decomposed into mean and turbulent components, $w'$ is the turbulent vertical velocity, $\nu$ is the kinematic fluid viscosity and $\rho$ is the fluid density.

Theoretically, $\tau_b$ is assumed to be constant throughout the boundary layer (constant stress layer), and a turbulent velocity scale was introduced to represent the shear strength, i.e. friction velocity $u_*$. 


\[ \tau_b = (\rho u^2)_{z=0} \]  \hspace{1cm} (4.2)

To obtain \( u_* \), the measured \( U(z) \) profiles were fitted to the logarithmic law-of-the-wall (Eq. 1.6).

In Reynolds-averaged Navier-Stokes (RANS) models, applied at field-scale, these processes are often parameterized using the Quadratic Stress Law, which casts \( u_* \) in terms of the mean current velocity at a certain height above the sediment and a drag coefficient \( C_D \) (e.g., Boudreau and Jorgensen, 2001; Lorke, 2007). The resulting \( u_*^2 \) can be expressed as

\[ u_*^2 = C_D U^2 \]  \hspace{1cm} (4.3)

, where the value of \( C_D \) depends on the height where the current velocity was measured, with 1 m usually applied as a reference value (Soulby et al., 1994; Lorke, 2007; Valipour et al., 2015a).

In natural aquatic systems, \( \tau_b \) is not only generated from mean currents (Lick et al., 1994; Churchill et al., 2004), as surface wave orbital velocities may also impinge on the bottom (Lou et al., 2000; Hawley et al., 2004; Valipour et al., 2017). As a result, commonly used parameterizations of \( \tau_b \) in the field (e.g., Hawley et al., 1996; Hawley and Eadie, 2007) or in models (e.g., Lick et al., 1994) are a summation of quadratic stress and surface wave-induced stress. The concept that initiation of sediment resuspension depends on whether \( \tau_b \) exceeds the theoretical time-averaged critical value (\( \tau_{cr} \)) has long played a central role in sediment transport theory (Shields, 1936; van Rijn, 1993; Soulsby and Whitehouse, 1997), and been applied in sediment transport models (van Rijn, 1987; Warner et al., 2008). With the development of three-dimensional RANS models, this parameterization concept, and its modified versions, have also been used in field-scale numerical simulation of sediment resuspension and transport (e.g., Hu et al., 2009 [Delft3D]; Morales-Marin et al., 2018 [FVCOM-SED]; Niu et al., 2018 [FVCOM-SED]; Chapter 3 [ELCOM-CAEDYM]). However, the algorithms applied in these RANS
models are not identical, with model-specific parameters requiring adjustment through calibration and validation against observed resuspension events (See 3.3).

While computationally suitable for inclusions in RANS equations models, the applicability of the Quadratic Stress Law to predict the occurrence of various types of resuspension events has been recently questioned (Boegman and Stastna, 2019). For example, in laboratory experiments (e.g., Boegman and Ivey, 2009; Aghsae and Boegman, 2015;) and field observations (e.g., Bourgault et al., 2014; Salim et al., 2018) sediment resuspension was associated with turbulent bursts, at times with sub-maximal $\tau_b$, and also when current velocities were below the critical value (e.g., Soulby et al., 1994; Yang et al., 2016; Salim et al., 2018). Thus, parameterization of $\tau_b$ based on turbulent velocity fluctuations has been proposed

$$\tau_{TKE} = \rho C_t \overline{w'w'}$$

(4.4), where $C_t$ is a proportionality constant, and compared with the quadratic stress law (Soulsby, 1983; Kim et al., 2000; Biron, et al., 2004; Aghsae and Boegman, 2015). Using single-point acoustic Doppler velocimeter (ADV) measurements of turbulent velocity fluctuations, Bluteau et al. (2016) found $\tau_{TKE}$ to better predict sediment resuspension over quadratic stress on the continental shelf, where internal waves shoaled. However, Zulberti et al. (2018) showed quadratic stress to be as accurate as that from near-bed turbulence-based parameterizations in a similar flow when measurements were close enough to the bottom.

It is evident that further research is required to enable better determination of $\tau_b$ from observational data and to better parameterize sediment resuspension in RANS models. The present enquiry-based study compares different methodologies to compute bottom stress from observations in central Lake Erie. The ability of RANS models to reproduce sediment resuspension events, resulting from different hydrodynamic forcing conditions, is also assessed.
4.2 Methods

4.2.1 Study area

Lake Erie (Figure 4.1a) is a large (388 km long and 92 km wide) and shallow lake (19 m average and 64 m maximum depth) that can be divided into western, central, and eastern basins. The shallowness of the western and central basins makes them susceptible to sediment resuspension by wind-induced surface waves (Sheng and Lick, 1979; Hawley and Eadie, 2007; Valipour et al., 2017; Chapter 3). In the central and eastern basins of Lake Erie, a seasonal thermocline forms with near-inertial (~17 h) Poincaré waves being the dominant wind-induced motions during the stratified period, in addition to the prominent ~14 h surface seiche (e.g., Boegman et al., 2001; Rao et al., 2008; Valipour et al., 2017). Although the topographic features of Lake Erie are complex and the sediment types and grain sizes vary among the basins, the most prevalent substrates in the lake include resuspendible silt, mud, and partially resuspendible glacial tills with grain size less than 63 μm (Haltuch et al., 2000).

Figure 4.1 (a) Map of Lake Erie showing the location of field observation (Sta. 341) and NDBC wave buoy (Sta. 45005). Negative numbers show the depth contours in meters. (b) The tripod equipped with ADCPs, an ADV and RBR TR-1060s before deployment on the lake bed at Sta. 341 in 2008.
4.2.2 Field observations and critical shear stress

Field observations were conducted in west central Lake Erie (Sta. 341; Figure 4.1a) during April-October of 2008-09, measuring water temperature, turbidity, TSS, and both mean and turbulent current velocities near the lake bed (Table 4.1). Water temperature was recorded at Sta. 341 using temperature loggers (RBR TR-1060) on a mooring line. A 1.8 m tripod was deployed nearby the mooring line (~30 m) on the lake bed, equipped with upward and downward looking Nortek Aquadopp acoustic Doppler current profiles (ADCPs; Figure 4.1b). A Nortek Vector acoustic Doppler velocimeter (ADV) was on the tripod at 1 m above bottom (1 mab). Meteorological data and wave information was obtained from National Data Buoy Center (NDBC) Sta. 45005 located 15 km to the south-west of Sta. 341, from which surface wave orbital velocities ($U_{orb}$) and surface wave-induced bottom stress ($\tau_w$) were calculated (see 4.2.4.1). Autoranging Seapoint turbidity and chlorophyll a (Chl-a) sensors logged to multi-parameter water quality sondes (RBR XR-620 and XR-420) located at 1.5 mab and 5 mab, respectively.
Table 4.1 Details of instrument deployments at Sta. 341 (water depth -17.5 m).

<table>
<thead>
<tr>
<th>Instrument</th>
<th>DOY (Year)</th>
<th>Sampling frequency (Hz)</th>
<th>Sampling interval (min)</th>
<th>No. of samples in each interval</th>
<th>Depth (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>TR-1060 temperature loggers</td>
<td>122-212</td>
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<td>N/A</td>
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<td>N/A</td>
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<tr>
<td></td>
<td>118-274</td>
<td>0.1</td>
<td>N/A</td>
<td>N/A</td>
<td>[1, 2, 3, 4, 5, 6, 8, 10, 12, 13, 14.5, 15.5, 16.5]</td>
</tr>
<tr>
<td>600 kHz upward looking ADCP</td>
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<td>15</td>
<td>1</td>
<td>0.6 to 14.6 (1 m bins)</td>
</tr>
<tr>
<td></td>
<td>118-196</td>
<td>1</td>
<td>15</td>
<td>1</td>
<td>1.3 to 15.3 (1 m bins)</td>
</tr>
<tr>
<td>2 MHz downward looking HR-ADCP</td>
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<td>15</td>
<td>256</td>
<td>1.83 mab (3 cm bins)</td>
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<tr>
<td>ADV</td>
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<td>20</td>
<td>4800</td>
<td>16.5</td>
</tr>
<tr>
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<td>119-274</td>
<td>16</td>
<td>20</td>
<td>4800</td>
<td>16.5</td>
</tr>
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<tr>
<td></td>
<td>212-304</td>
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<td>2</td>
<td>1</td>
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<td>119-275</td>
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<td>3</td>
<td>1</td>
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</tr>
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</table>

From two superficial sediment samples collected at Sta. 341 on 26 August 2009, sediment particle diameters were measured $d_{50} = 10 \mu m$ (J. D. Ackerman personal communication), and the bulk and granular densities were $\rho_b = 1093 \text{ kg m}^{-3}$ and $\rho_s = 2150 \text{ kg m}^{-3}$. The existing Shields diagram does not give critical value for sediment finer than $40 \mu m$. Valipour et al. (2017) concluded that in west-central Lake Erie, the critical value for time-averaged current velocity, $U_{cr}$, to trigger resuspension was $0.1 \text{ m s}^{-1}$, and the bottom drag coefficient $C_D = 0.0045$, obtained by least-square fitting the burst averaged HR-ADCP velocity profiles to the law-of-the-
Thus, we determined the time-averaged critical bottom shear stress
\[ \tau_{cr} = \rho C_D U_{cr}^2 = 0.045 \text{ Pa}. \]

4.2.3 Identification of sediment resuspension events

Sediment resuspension events were qualitatively identified by an increase of turbidity and acoustic backscatter signals, including the ADV backscatter amplitude (hereafter, ADV-amp, unit [Count]) and in 2009, the HR-ADCP backscatter, corrected following Lohrmann (2001) for attenuation (hereafter, ADCP echo level, unit [dB]). The cross-correlation of these three indicators can be found in Valipour et al. (2017).

The turbidity sensor measurements include sediment and algal biomass, whereas the ADV and ADCP backscatter occur off sediment but not algae (Lohrmann, 2001). The Chl-a concentration recorded by the XR-420 in the spring of 2009 was used to exclude algal biomass settling events from the turbidity data. The ADCP echo profiles enabled identification of the particulate source, as originating from advection or local resuspension.
4.2.4 Bottom shear stress parameterization based on observed data

Four parameterization methods were used to estimate $\tau_b$: 1) the surface wave-induced stress ($\tau_w$); 2) the quadratic stress law ($\tau_c$); 3) the log-law ($\tau_L$); and 4) the turbulent kinetic energy ($\tau_{TKE}$). The total bottom stress, $\tau_b$ is fully represented by $\tau_L$, $\tau_{TKE}$ and the summation $\tau_b = \tau_w + \tau_c$.

4.2.4.1 Surface wave-induced stress

From wave theory, $\tau_w$ is (Jonsson, 1966; van Rijn, 1990)

$$\tau_w = 0.5 \rho f_w U_{orb}^2$$

(4.5)

, where $f_w$ is the wave friction coefficient. In the laminar range ($\frac{a \times U_{orb}}{\nu} < 10^4$), $f_w = 2 \left( \frac{a \times U_{orb}}{\nu} \right)^{-0.5}$; in the hydraulic smooth range ($10^5 > \frac{a \times U_{orb}}{\nu} > 10^4$), $f_w = 0.09 \left( \frac{a \times U_{orb}}{\nu} \right)^{-0.2}$; and in the hydraulic rough range ($\frac{a \times U_{orb}}{\nu} > 10^5$),\( \exp \left[-6 + 5.2 \left( \frac{a}{k_s} \right)^{-0.19} \right] \)

$U_{orb}$ and $a$ are the maximum orbital velocity (m s$^{-1}$) and the maximum bottom amplitude (m), respectively, given by linear wave theory

$$U_{orb} = \frac{\pi H_s}{T_s \sinh \left( \frac{2\pi h}{L} \right)}$$

(4.6)

$$a = \frac{H_s}{2 \sinh \left( \frac{2\pi h}{L} \right)}$$

(4.7)

, where $h$ and $H_s$ are the water depth (m) and wave height (m), $T_s$ is the wave period (s), and $L$ is the wave length (m). These parameters were estimated from wind speed, fetch and water depth (Barua, 2005).

4.2.4.2 Quadratic stress
Multiplying by density, an equation relating stress \( \tau_c \) to the mean current velocity is obtained

\[
\tau_c = \rho u^2 = \rho C_D U^2 \quad (4.8)
\]

where \( U = \sqrt{U_x^2 + U_y^2} \) is the burst-averaged mean horizontal current velocity 1 mab from the ADV. Here, \( U_x \) and \( U_y \) are the 5-min averaging current velocities in east-west and north-south directions, which filters surface wave information. The drag coefficient \( C_D = 0.0045 \) from Valipour et al. (2015a).

4.2.4.3 Log-law method

In the log-law, \( u_\ast \) may be derived from the rate of dissipation of turbulent kinetic energy (TKE) \( \varepsilon \)

\[
\tau_L = \rho u^2 = \rho (\varepsilon \kappa z)^{2/3} \quad (4.9)
\]

where \( \kappa = 0.4 \) is the von Karman constant and \( \varepsilon \) was obtained by inertial fitting to the ADV burst data at a height \( z \) (1 mab), denoted \( \varepsilon_{ID} \).

Both \( \tau_c \) and \( \tau_L \) assume the mean current velocity profile is logarithmic and the flow is steady and unidirectional, but \( \tau_c \) filters turbulent fluctuations and \( \tau_L \) retains turbulence information. To compute \( \varepsilon_{ID} \), we fit the energy spectrum of the vertical turbulent velocity component (\( w' = w - \sqrt{\bar{w}^2}; w \) is the instantaneous vertical velocity, and the overbars denote 5-min averaging) from ADV burst data to the theoretical form within the inertial sub-range (see Appendix C; Text C1, Figure C1).

4.2.4.4 TKE method
We applied a modified TKE method (Eq. (4.4)) following Kim et al. (2000); Biron et al. (2004) and Bluteau et al. (2016). The average ratio of $\tau_b$ to TKE is constant in the atmosphere (0.19; Stapleton and Huntley, 1995; Kim et al., 2000; Biron et al., 2004); by assuming linear relationships between TKE and the vertical variance: $\overline{w'w'} = 0.59TKE$ (Pope, 2000), the constant $C_t$ was set to 0.32. The modification was suggested (Kim et al., 2000), not only because vertical velocity fluctuations have smaller instrument noise than horizontal velocity fluctuations (Voulgaris and Trowbridge, 1998), but bursts of vertical velocity lift the bed sediment more efficiently (Yuan et al., 2009; Aghsaee and Boegman, 2015). The TKE method is expected to be better representation of $\tau_b$ in complex flow fields or when the measurements are outside the logarithmic layer. At Sta. 341, the logarithmic layer can extend to more than 10 m above the bed, but it also becomes limited by baroclinic currents when stratification strengthens (Kim et al., 2000; Valipour et al., 2015a). Thus, the independence of the TKE method from the logarithmic layer is supposed to make it more accurate in parameterization than other methods in our observation site.

4.2.5 Flow interference

The turbulence measurements were evaluated to identify if $\epsilon_{ID}$ or $w'$ were contaminated by vortex shedding from the mooring. The orientation of the ADV and locations of the battery canister were different in 2008 than in 2009, with varying directions of flow interference. In 2008 (Figure 4.1b), the main interference came from the acoustic Doppler velocimeter battery canister; whereas, in 2009, the tripod frame was the source of interference (Valipour et al., 2015a). To identify interference, we followed McGinnis et al. (2014), and correlated $\epsilon_{ID}$ with the third-power of the mean flow velocity 1 m above bottom ($U_{1m}^3$) (Figure 4.2a-c). The ratio between predicted $\epsilon_{p,1m}$ and observed $\epsilon_{ID}$ gave the flow directions contaminated with
interference (Figure 4.2d-f). The largest deviations in 2008 came from a broad angle consistent with the location of the battery canister (Figure 4.2d). In 2009, the largest deviation came from three narrow angles, indicating the tripod arms (Figure 4.2e, f). The data contaminated with interference were removed, leading to gaps in \( \tau_L \) and \( \tau_{TKE} \) time-series.

![Graphs showing predicted relationships between dissipation and current velocities](image)

**Figure 4.2** (a-c) Predicted relationship \((\varepsilon_{p,1m}, \text{solid lines})\) between the dissipation and the current velocities at 1 m above the bed with 95% confidence (dashed line) for three deployment periods (days 214-275 in 2008, days 119-196 in 2009 and days 196-274 in 2009). (d-f) Ratio of the calculated dissipation from inertial fitting \((\varepsilon_{ID})\) to the predictions \((\varepsilon_{p,1m})\) versus current direction at 1 m above the bed for corresponding three deployment periods.

### 4.2.6 Bottom shear stress parameterization in RANS models

We applied two coupled hydrodynamic and water quality RANS models ELCOM-CAEDYM (hereafter, ELCD) and AEM3D-iWaterQuality (hereafter AEM3D). Both models are distributed by Hydronumerics (www.hydronumerics.com.au) and differ primarily in AEM3D being a new version of ELCD. The models solve the unsteady RANS equations for
incompressible flow using Boussinesq and hydrostatic approximations. The water quality modules both predict resuspension when \( \tau_b > \tau_{cr} \), where \( \tau_b \) is the summation of surface wave-induced \( (\tau_w) \) and current-induced stresses; however, the algorithms for predicting these stresses differ between the two models.

### 4.2.6.1 Surface wave-induced stress

In ELCD, \( \tau_w \) is from Eq. (4.5), where \( f_w \) is assumed to be for hydraulically rough flow, with \( k_s = 2.5 \, d_{50} \) and \( d_{50} \) is the median sediment grain size.

In AEM3D, \( \tau_w \) is related to a user-defined bottom drag coefficient \( C_D \),

\[
\tau_{w,m} = \rho_w C_D U_{orb}^2
\]  

(4.10)

### 4.2.6.2 Quadratic stress

Both AEM3D and ELCD predict \( \tau_c \) according to quadratic stress law. In ELCD,

\[
\tau_{c,m} = \rho_w \frac{f_c U_{bot}^2}{8}
\]  

(4.4)

, where \( f_c \) is the friction coefficient for hydraulically rough flow (van Rijn, 1993; Eq (3.7)). In AEM3D,

\[
\tau_{c,m} = \rho_w C_D U_{bot}^2
\]  

(4.5)

, where \( U_{bot} \) is the RANS modeled current speed in the bottom layer. Rather than relying on \( d_{50} \) for resuspension, through \( f_c \) (Eq. (3.7)), which also impacts particle settling, AEM3D allows users to apply an \emph{in situ} measured \( C_D = 0.0045 \) (Valipour et al., 2015a) or canonical \( C_D = 0.0024 \) (Soulby et al., 1994; for mud/sand/gravel). In the present application, we ran AEM3D with both values. The consistency of \( C_D \) in algorithms of \( \tau_{w,m} \) and \( \tau_{c,m} \) means the threshold of resuspension in the model can be set identically for surface waves and currents.
In the Discussion (section 4.4.1), the results from these algorithms were compared to predictions from $\tau_c$ algorithms in other commonly-used sediment transport models (i.e., FVCOM-SED and Delft3D).

4.2.6.3 Log-law method

The mixed layer closure scheme in AEM3D and ELCD computes a TKE balance (Hodges et al., 2000; Spigel et al., 1986) giving TKE dissipation as

$$\varepsilon_m = \frac{1}{2} C_\varepsilon \Delta t \left( \frac{TKE}{\Delta z} \right)^{3/2}$$

(4.6)

where the dissipation coefficient $C_\varepsilon = 1.15$, $\Delta t$ is the timestep, $\Delta z$ is vertical layer size, and TKE is the available mixing energy, which is the summation of wind stirring energy production, shear production between layers, and buoyancy production. While Eq. (4.13) is also filtered in a RANS scheme, it would be informative to compare Eq. (4.9) using grid-cell averaged dissipation ($\varepsilon_m$) to that from direct inertial fitting dissipation ($\varepsilon_{ID}$) to observed data.

4.2.6.4 TKE method

Reynolds-averaging filters sub-grid-scale turbulent fluctuations, providing only mean flows; making it unrealistic to resolve turbulent vertical velocities and apply the TKE method (Eq. (4.4)) to parameterize $\tau_b$ within a RANS model.

4.2.7 Model setup

ELCD and AEM3D were configured as in the validated ELCOM model applied by Liu et al. (2014), including meteorological forcing, inflows, outflows, and a $2 \times 2$ km horizontal grid with 45 vertical layers. A finer 0.5 m grid was set near the surface, through the thermocline and thin central basin hypolimnion, and coarser 5 m grid was set in the deep (~65 m) eastern basin (Figure 3.1c). At Sta. 341, this gave a 0.75 m bottom layer, to capture the thin bottom boundary
layer, and the validation of bottom mean currents can be found in Appendix (Figure B2) and following content (Figure 4.10). Spatial variability of meteorological conditions across the lake was considered using 6 sub-sections: western, central (further subdivided into 4 quadrants), and eastern basins. The sources of meteorological data included (1) Environment and Climate Change Canada (ECCC) lake buoy data (central basin, Port Stanley 45132; eastern basin, Port Colborne 45142), (2) US National Data Buoy Center (NDBC) buoys (western basin, station 45005), (3) ECCC land stations, (4) US NDBC land stations, (5) climate model data (CRCM5 and GEMS). There were eleven inflows, including the Detroit, Maumee, Grand (Ontario), Sandusky, Cuyahoga, Raisin, Buffalo, Cattaraugus, Grand (Ohio), Rocky, and Vermilion Rivers, and only one outflow, the Niagara River (León et al., 2011). To satisfy the Courant-Friedrichs-Levy condition, the timestep was 5 min. We initialized the model using observed water temperature profiles throughout the lake from a spring survey (U.S. EPA); 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 (Valipour et al., 2015b). In the 2008 model run, we had the observed TSS concentration from river loading and water samples collected at multiple stations as input data and initial conditions (León et al., 2011; Chapter 3). The models were run for days 203-303 (2008). Sediments in models were separated into three classes: river loads (SSR, $d_{50} = 3 \mu m$; Fukuda and Lick, 1980) and lake bed (SSB1, $d_{50} = 1 \mu m$; SSB2, $d_{50} = 10 \mu m$; Lick et al., 1994). SSB1 represented clay-like superficial (nepheloid) sediments (Lick and Lick, 1988; Lick et al., 1994), whereas SSB2 represented the silt-like sediments (Hawley and Eadie, 2007; Valipour et al., 2017). The lake bed sediment classes were proportioned at 20% (clay) and 75% (silt), according to observations from the PONAR grabs (J. D. Ackerman, personal communication). For the 2009 model run (days 118-275), river loading suspended solids and observed TSS
concentration were not available for input and validation, and so we only simulated lake hydrodynamics to obtain $\tau_b$.

Results from ELCD and AEM3D were quantitively compared to observation-based parameterization using the percent bias ($P_{bias}$), Pearson correlation coefficient ($R$), and the root-mean-square error ($RMSE$).

4.3 Results

4.3.1 Prediction of resuspension from observed $\tau_b$

We calculated the 7-day moving average and standard deviation of ADV-amp and turbidity data; identifying sediment resuspension events when observed data exceeded one standard deviation from the mean. In the spring and summer, settling of algae (Paerl et al., 2011; Modis, NOAA Coastwatch-Great Lakes) contributed to some turbidity peaks consistent with high fluorescence (Chl-a) during the first deployment period (days 119-195) of 2009. Valipour et al. (2017) used the MEdium Resolution Imaging Spectrometer (MERIS) to exclude turbidity peaks from high algal biomass (Sta. 341, days 226, 236, 245; Figure 4.5b). We then identified twenty-three sediment resuspension events (Figure 4.3-5; R1-23) from turbidity, ADV-amp and ADCP echo data. All three indicators showed resuspension during several especially intense events (R1, 3, 4, 7, 9, 12, 13, 21-23).
Figure 4.3 Time-series at Sta. 341 of 2008 (a) wind speed (blue line; left y-axis) and direction (red dash line; right y-axis) at 10 m above water surface, (b) turbidity at 1.5 mab from XR-620 (black stars) and its 7-day moving mean (grey line), (c) ADV-amp at 1 mab (blue dots) and its 7-day moving mean (blue line), (d) $\tau_L$ (purple dash-dot line, Eq. 4.9), $\tau_{\text{TKE}}$ (green dash-dot line, Eq. 4.8) and critical value for resuspension ($\tau_{cr} = 0.045$ Pa; black dash line), (e) $\tau_w$ (blue line, Eq. 4.5), $\tau_c$ (red line, Eq. 4.3) and $\tau_{cr}$, (f-h) Modeled $\tau_{c,m}$ and $\tau_{w,m}$ based on varying $C_D$ and algorithms.
Figure 4.4 Time-series at Sta. 341 of 2009 first deployment period (a) wind speed (blue line; left y-axis) and direction (red dash line; right y-axis) at 10 m above water surface, (b) turbidity at 1.5 mab from XR-620 (black stars) and its 7-day moving mean (grey line) (left y-axis), and Chl-a concentration at 5 mab from XR-420 (right y-axis), (c) ADV-amp at 1 mab (blue dots) and its 7-day moving mean (blue line), and colorbar shows the ADCP echo level, (d) $\tau_L$ (purple dash-dot line, Eq. 4.9), $\tau_{TKE}$ (green dash-dot line, Eq. 4.8) and $\tau_{cr}$ (black dash line), (e) $\tau_w$ (blue line, Eq. 4.5), $\tau_c$ (red line, Eq. 4.3) and $\tau_{cr}$, (f-h) Modeled $\tau_{c,m}$ and $\tau_{w,m}$ based on varying $C_D$ and algorithms.
Figure 4.5 Time-series at Sta. 341 of 2009 second deployment period (a) wind speed (blue line; left y-axis) and direction (red dash line; right y-axis) at 10 m above water surface, (b) turbidity at 1.5 mab from XR-620 (black stars) and its 7-day moving mean (grey line), and green shading indicates the high turbidity from algae (Paerl et al., 2011), (c) ADV-amp at 1 mab (blue dots) and its 7-day moving mean (blue line), (d) $\tau_L$ (purple dash-dot line, Eq. 4.9), $\tau_{TKE}$ (green dash-dot line, Eq. 4.8) and $\tau_{cr}$ (black dash line), (e) $\tau_w$ (blue line, Eq. 4.5), $\tau_c$ (red line, Eq. 4.3) and $\tau_{cr}$, (f-h) Modeled $\tau_{c,m}$ and $\tau_{w,m}$ based on varying $C_D$ and algorithms.

The four parameterizations (Eq. (4.5), (4.8), (4.9), and (4.4)) were applied to compute $\tau_b$ from the observed data and compared with $\tau_{cr}$ to predict the occurrence of resuspension events.

During intensive resuspension events, $\tau_b = \tau_w + \tau_c$, $\tau_L$, $\tau_{TKE}$ were qualitatively consistent, with spikes of different magnitudes (R1, 3, 4, 7, 9, 12, 14, 18, and 21-23).
Strong wind events created significantly increased $\tau_w$, leading to surface wave-dominated resuspension events (e.g., R21 and R22; Figure 4.6). During these events, high wind speeds (> 10 m s$^{-1}$) were observed (see also Hawley and Eadie 2007). Given that wave orbitals penetrating to the lake bed can form turbulent eddies, $\tau_L$ and $\tau_{TKE}$ often showed remarkable increases (> 0.2 Pa) during the surface wave-dominated resuspension events.

Bottom current-dominated resuspension events were also observed (e.g., R14-17; Figure 4.7). With increased $\tau_c$ exceeding $\tau_{cr}$, $\tau_L$ and $\tau_{TKE}$ were elevated (>0.045 Pa), indicating the turbulent eddies formed due to bottom friction. Compared to surface wave-dominated resuspension, R14–17 had less sharp increases in $\tau_L$ and $\tau_{TKE}$ (< 0.1 Pa), indicating bottom currents were less efficient in triggering turbulent bursts compared to wave orbitals.

Most resuspension events were not induced by a single mechanism, but resulted from combined effects of surface waves and mean currents. Storm-induced mean currents were observed after strong wind events in Lake Erie (Lick et al., 1994; Beletsky et al., 1999; Hawley and Eadie, 2007), leading to increased $\tau_c$ and generating resuspension (e.g., R1-5, 7, 9, 10, 12, 18, 19, 20, 23). However, independent $\tau_w$ or $\tau_c$ did not always capture the exact timing of strong resuspension, rather $\tau_L$ and $\tau_{TKE}$ corresponded with peaks of ADV-amp and turbidity more accurately (R7, R10, and R12; Figure 4.8). This was in agreement with oceanic (Bluteau et al., 2016) and laboratory (Aghsaee and Boegman, 2015) data, showing drag to result in bedload transport, with turbulent bursts required to lift sediment into the water column. For example, during day 146 in 2009 (R12) wind-driven surface wave orbitals impinged on the lake bed, generating turbulence ($\tau_w$ ~ 0.1 Pa, $\tau_L$ and $\tau_{TKE}$ ~ 0.2 Pa; Figure 4.8) and triggering significant peaks in ADV-amp ADCP echo level, and turbidity. This was followed by barotropic currents from basin-scale seiche events that formed as the wind subsided (Beletsky et al., 1999; Valipour
et al., 2015b) and generated \( \tau_c > \tau_{cr} \) on day 147, leading to another peak in three resuspension indicators.

**Figure 4.6** Details of resuspension events R21-23 in Figure 4.5. Time-series at Sta. 341 of (a-c) wind speed (blue lines; left y-axis) and direction (red dash lines; right y-axis) at 10 m above water surface, (b-f) turbidity at 1.5 mab from XR-620 (black stars) and its 7-day moving mean (grey line), (g-i) ADV-amp at 1 mab (right y-axis), and its 7-day moving mean (blue line), (j-l) show \( \tau_{TKE} \) (purple dash-dot line) and \( \tau_L \) (green dash-dot line) based on turbulent velocity from ADV, and *in situ* critical value for resuspension \( \tau_{cr} = 0.045 \) Pa, (m-o) \( \tau_w \) (blue line) based on buoy from NDBC station 45005, \( \tau_c \) (red line) based on mean current velocity from ADV, and \( \tau_{cr} \) (black dash line), (p-r) AEM3D output \( \tau_{w,m} \) (Eq. 4.10; blue dot line) and \( \tau_{c,m} \) (Eq. 4.12 red dot line) used \( C_D = 0.0024 \), (s-u) AEM3D output \( \tau_{w,m} \) (Eq. 4.10; blue dashed line) and \( \tau_{c,m} \) (Eq. 4.12; red dashed line) used \( C_D = 0.0045 \); and (v-x) are ELCD output \( \tau_{w,m} \) (Eq. 4.5; blue dash line) and \( \tau_{c,m} \) (Eq. 4.11; red dash line).
Figure 4.7 Details of resuspension events R14, 15, 17 in Figure 4.4. Time-series at Sta. 341 of (a-c) wind speed (blue lines; left y-axis) and direction (red dash lines; right y-axis) at 10 m above water surface, (b-f) turbidity at 1.5 mab from XR-620 (black stars) and its 7-day moving mean (grey line), green shading indicates the high turbidity from algae (Paerl et al., 2011), (g-i) ADV-amp at 1 mab (right y-axis), and its 7-day moving mean (blue line), (j-l) show $\tau_{TKE}$ (purple dash-dot line) and $\tau_c$ (green dash-dot line) based on turbulent velocity from ADV, and in situ critical value for resuspension $\tau_{cr} = 0.045$ Pa, (m-o) $\tau_w$ (blue line) based on buoy from NDBC station 45005, $\tau_c$ (red line) based on mean current velocity from ADV, and $\tau_{cr}$ (black dash line), (p-r) AEM3D output $\tau_{w,m}$ (Eq 4.10; blue dot line) and $\tau_{c,m}$ (Eq 4.12 red dot line) used $C_D = 0.0024$; (s-u) AEM3D output $\tau_{w,m}$ (Eq 4.10; blue dash line) and $\tau_{c,m}$ (Eq 4.12; red dash line) used $C_D = 0.0045$, and (v-x) are ELCD output $\tau_{w,m}$ (Eq 4.5; blue dash line) and $\tau_{c,m}$ (Eq 4.11; red dash line).
Figure 4.8 Details of resuspension events R7, 10, 12 in Figure 4.3, 4.4. Time-series at Sta. 341 of (a-c) wind speed (blue lines; left y-axis) and direction (red dash lines; right y-axis) at 10 m above water surface, (b-f) turbidity at 1.5 mab from XR-620 (black stars) and its 7-day moving mean (grey line), and Chl-a concentration at 5 mab from XR-420 (right y-axis), (g-i) ADV-amp at 1 mab (right y-axis), its 7-day moving mean (blue line), and colorbar shows the ADCP echo level; (j-l) show \( \tau_{TKE} \) (purple dash-dot line) and \( \tau_L \) (green dash-dot line) based on turbulent velocity from ADV, and in situ critical value for resuspension (\( \tau_{cr} = 0.045 \) Pa; black dash line); (m-o) \( \tau_w \) (blue line) based on buoy from NDBC station 45005, \( \tau_c \) (red line) based on mean current velocity from ADV, and \( \tau_{cr} \) (black dash line); (p-r) AEM3D output \( \tau_{w,m} \) (Eq 4.10; blue dot line) and \( \tau_{c,m} \) (Eq 4.12; red dot line) used \( C_D = 0.0024 \); (s-u) AEM3D output \( \tau_{w,m} \) (Eq 4.10; blue dash line) and \( \tau_{c,m} \) (Eq 4.12; red dash line) used \( C_D = 0.0045 \); and (v-x) are ELCD output \( \tau_{w,m} \) (Eq 4.5; blue dash line) and \( \tau_{c,m} \) (Eq 4.11; red dash line).

We have investigated the ability to parameterize resuspension by wave-orbital and seiche-induced mean currents; however, \( \tau_b = \tau_c + \tau_w \) is not expected to be able to parameterize resuspension resulting from near-bed turbulent events forced by other processes (e.g.,
convection, Kelvin-Helmholtz billows). For example, Valipour et al. (2017) suggested that degeneration of the Kelvin-Helmholtz billows could resuspend bottom material, when the induced turbulence penetrated to the bed (see also Hawley, 2004; Austin, 2013). These events could only be captured by $\tau_L$ and $\tau_{TKE}$, because $\tau_c$ utilizes time-averaged mean currents that filter turbulence. Here, we test the observational and RANS parameterizations for this type of event. After a wind event on day 249 (Figure 4.9), the thermocline, acting as a waveguide for high-frequency internal waves (HFIW), impinged upon the lake bed (days 250-254). During this 10-day event, $\tau_b = \tau_c + \tau_w > \tau_{cr}$ on days 248-9, corresponding to an increase of ADV-amp and turbidity (R6); rather $\tau_{ID}$ and $\tau_{TKE}$ spiked above $\tau_{cr}$ on days 252 and 253.5, corresponding to peaks in ADV-amp and high turbidity. At these times, when $\tau_w$ and $\tau_c$ were close to zero and HFIWs were carried on the near-bed thermocline (Figure 4.9b), the peaks in $\tau_L$ and $\tau_{TKE}$ matched peaks in ADV-amp and turbidity, suggesting the mechanism triggering near-bed high turbidity could be the turbulent eddies generated by HFIW degeneration (Figure 4.9c, d). Compared to resuspension induced by $\tau_c$ or $\tau_w$, the intensity of resuspension generated by degeneration of HFIW was lower. Both turbidity and ADV-amp were elevated for several days, showing that turbulent eddies, generated when the thermocline impinged on the lake bed, created an oscillatory nepheloid-type layer in the hypolimnion. In this example, $\tau_L$ and $\tau_{TKE}$ provided better estimations of sediment resuspension.

Overall, the observations spanning the summer of 2008 and spring-fall of 2009 revealed that $\tau_L$ and $\tau_{TKE}$ showed peaks during resuspension triggered by wave orbitals, bottom mean currents, and HFIWs, with relatively higher stress magnitude in resuspension involving a contribution from surface waves (e.g., R7, 12, 21-23; Figure 4.6, 4.8) while relatively lower stress magnitude in resuspension was only contributed from the bottom mean currents or HFIWs.
(e.g., R10, 14-17; Figure 4.7, 4.8), rather $\tau_b = \tau_w + \tau_c > \tau_{cr}$ was able to predict all resuspension events induced by wave orbitals (R8, 21, 22), increased bottom currents (R13, 14-17), and combination of these two mechanisms (R1-5, 7, 9, 10, 12, 18, 19, 20, 23).

Figure 4.9 Details of resuspension events R6 in Figure 4.3. Time-series at Sta. 341 of 2008 (a) wind speed (left y-axis) and direction (right y-axis) at 10 m above water surface, (b) temperature contours from TR-1060 temperature loggers, red arrows show locations of temperature loggers as in Table 1, (c) turbidity at 1.5 mab from XR-620 (left y-axis), and ADV-amp at 1 mab (right y-axis); (d) and (e) show $\tau_L$ (green dash-dot line), $\tau_{TKE}$ (purple dash-dot line), $\tau_w$ (blue line), $\tau_c$ (red line) based on observed data and $\tau_{cr}$ (black dash line); (f) and (g) are ELCD output of temperature and TSS concentration at Sta. 341, respectively.

### 4.3.2 RANS modeled $\tau_b$
Chapter 3 showed that ELCD qualitatively captures the occurrence of strong resuspension events induced by both bottom currents and surface waves in 2008. However, the magnitude of $\tau_{c,m}$ by ELCD model, parameterized based on Reynold-averaged current speed (Eq. (4.11)), was much less than the observed $\tau_c$ (Figure 4.3-5h). Thus, calibrated against turbidity data, the threshold of sediment resuspension in the ELCD model (0.01-0.025 Pa) was lower than the observed in situ threshold (0.045 Pa), which was appropriate for surface wave-induced resuspension but not bottom mean current-induced resuspension in the model. Therefore, unrealistic setting of $\tau_{cr}$, to capture current-induced resuspension, will cause these events to respond excessively to surface wave forcing, and overestimate the contribution of surface waves to resuspension (e.g. R14 - R17; Figure 4.3-4.5).

AEM3D uses different algorithms, directly applying identical $C_D$ to parameterize $\tau_{w,m}$ and $\tau_{c,m}$ (Eq. (4.10), (4.12)). Figure 4.3-5f, g show $\tau_{c,m}$ and $\tau_{w,m}$ output by AEM3D, with the canonical $C_D = 0.0024$ and in situ $C_D = 0.0045$. The model reproduced spikes of current speeds, but the magnitude was overestimated, likely due to Reynolds-averaging (Figure 4.10b). The AEM3D-modeled bottom current velocities were not sensitive to small variations of $C_D$ (C. Dallimore, personal communication). Thus, when applying the $C_D = 0.0045$, the magnitude of $\tau_{c,m}$ was larger than $\tau_c$, with the highest percentage bias amongst the three algorithms (Table 4.2). Taking $C_D = 0.0024$ and Eq. (4.8) was chosen as the optimal algorithm for $\tau_{c,m}$ because of the lowest RMSE and low $P_{bias}$. Figure 4.3-5e, f show that this algorithm correctly reproduced most spikes in $\tau_{c,m}$ with the appropriate magnitude.

Similarly, ELCD overestimated wave orbital velocities impinging on the bottom (Eq. (4.5); Figure 4.10a), with RMSE = 0.049 (Figure 4.3-5e, h). Using the algorithm in AEM3D (Eq. (4.10)), RMSE could be reduced to 0.031 and $R^2$ increased (Figure 4.3-5e, f).
Consistent with the observations, ELCD model output temperature and TSS concentration, during the HFIW-induced resuspension event, showed that the strong resuspension event on day 249 of 2008 with maximal TSS ~ 6 mg L\(^{-1}\) near the lake bed (Figure 4.9g), corresponded to the rapid increase in hypolimnion thickness (Figure 4.9f). Discrepancies, between model and observation, arose when the TSS concentration decreased between days 249-55, when the TSS concentration did not exceed 4 mg L\(^{-1}\), indicating the oscillating turbid layer under thermocline, generated by HFIW induced near-bed turbulence, was not obvious in model results (days 253-255). HFIWs have wavelengths ~ 10 m in Lake Erie (Bouffard et al., 2012), which will not be resolved with the 2 km horizontal grid and model timestep (5 min).

Unable to resolve the sub-grid turbulence, the model may only predict a subset of resuspension events, and neglect of HFIW-induced resuspension can be expected. Thus, we applied Eq. (4.9), based on modeled \(\varepsilon_m\) (Eq. (4.13)), and assessed the possibility of applying the log-law method parameterization in a RANS model. The comparisons between \(\varepsilon_{ID}\) and \(\varepsilon_m\) at selected periods are shown in Appendix C (Figure C2). Figure 4.11 shows modeled \(\tau_{L,m}\) was higher than observed \(\tau_L\) in most of the time and the model was unable to capture peaks in observed \(\tau_L\), \(\tau_{L,m}\) and \(\tau_L\) showed a low correlation coefficient, and the RMSE and \(P_{bias}\) between them were quite high (Table 4.2). Thus, the parameterization of \(\varepsilon\) should be further calibrated and validated before being applied to the parameterization of \(\tau_L\). The limitation of \(\tau_{L,m}\) existing in the present RANS models is discussed in the following section (Discussion 4.4.2).
Figure 4.10 Time-series at Sta. 341 of (a) surface wave orbital velocity at the bed calculated from wave height and period recorded by NDBC-45005 (black line) and model output (red line); (b) current velocities 1-m above the bottom from the ADV (black line) and model output (red line). The stability of buoy in 2009 was not good, leading to many missing measurements.

Table 4.2 Assessment of bottom stress parameterization in model

<table>
<thead>
<tr>
<th>Modeled $\tau_{w,m1}$</th>
<th>Modeled $\tau_{w,m2}$</th>
<th>Modeled $\tau_{w,m3}$</th>
<th>Modeled $\tau_{c,m1}$</th>
<th>Modeled $\tau_{c,m2}$</th>
<th>Modeled $\tau_{c,m3}$</th>
<th>Modeled $\tau_{L,m}$</th>
</tr>
</thead>
<tbody>
<tr>
<td>RMSE</td>
<td>0.049</td>
<td>0.043</td>
<td>0.031</td>
<td>0.018</td>
<td>0.041</td>
<td>0.025</td>
</tr>
<tr>
<td>$R^2$</td>
<td>0.41</td>
<td>0.45</td>
<td>0.45</td>
<td>0.12</td>
<td>0.17</td>
<td>0.12</td>
</tr>
<tr>
<td>$P_{bias}$</td>
<td>0.31</td>
<td>0.16</td>
<td>0.55</td>
<td>0.45</td>
<td>1.29</td>
<td>0.35</td>
</tr>
</tbody>
</table>

$\tau_{w,m1} = 0.5 \rho f_w U_{orb}^2; \tau_{w,m2} = 0.0045 \rho U_{orb}^2; \tau_{w,m3} = 0.0024 \rho U_{orb}^2$

$\tau_{c,m1} = 0.125 \rho f_c U^2; \tau_{c,m2} = 0.0045 \rho U^2; \tau_{c,m3} = 0.0024 \rho U^2$

Figure 4.11 Time-series at Sta. 341 of $\tau_L$ (Eq. (4.9)) based on $\epsilon_m$ (Eq. (4.13) orange line) and $\epsilon_{ID}$ (green dash-dot line).
4.3.3 Hot-spots of sediment resuspension

After stratification broke down in autumn of both 2008 and 2009, intense resuspension events were more frequent (Figure 4.3b, 5b, R4, 7, 21-23), in agreement with ELCD results in Chapter 3.

**Figure 4.12** Mean value of AEM3D modeled \( \tau_{w,m}, \tau_{c,m}, \) and \( \tau_{b} \) (a-c) over days 203 – 245 of 2008 and days 119-250 of 2009, that is, before fall turnover, and (d-f) over days 245-303 of 2008 and days 250 – 300 of 2009, that is, after fall turnover in Lake Erie.

Table 4.2 shows that the parameterization based on Eq. (4.12) with \( C_D=0.0024 \) is optimal in the AEM3D model. According to this model parameterization, we reproduced the mean value of \( \tau_{w,m}, \tau_{c,m}, \) and \( \tau_{b} = \tau_{w,m} + \tau_{c,m} \) before and after the fall turnover in 2008 and 2009 across the whole lake, and defined the sites where the mean value exceeded \( \tau_{cr} = 0.045 \) Pa as potential hot-spots for sediment resuspension. In general, both \( \tau_{w,m} \) and \( \tau_{c,m} \) showed general increases after fall turnover. The effect of surface wave orbitals decreases with increasing water depth, and the western basin and littoral zones have the highest \( \tau_{w,m} \). The area of modelled surface wave-
induced resuspension hot-spots increased from 80 km² before fall turn over to 2592 km² after fall turnover (Figure 4.12a, d) due to frequent storms in fall. The current-induced resuspension hot-spots were sporadically distributed in the western basin and north shore of the central and eastern basin, with the area of hot-spots increasing from 84 km² before fall turnover to 168 km² after fall turnover (Figure 4.12b, e). The current-induced resuspension were related to wind-driven seiche events and baroclinic currents (e.g., Hawley, 2004; Rao et al., 2008; Valipour et al., 2017). The presence/absence of a thermocline affects the latter one, eliminating one of the factors leading to strong bottom currents. After fall turn over, frequent storms also strengthen the wind-driven seiche currents. The relative magnitudes of these two factor resulted in the increase of current-induced resuspension hot-spots to be less than those induced by surface waves.

Combining these two processes, the total area of resuspension hot-spots was 1920 km² before fall turnover and 5196 km² after fall turnover, concentrating in the western basin, and the north shoreline of the central and eastern basins (Figure 4.12c, f).

4.4 Discussion

4.4.1 Comparison of algorithms in commonly applied hydrodynamic models

Table 4.3 $\tau_{c,m}$ algorithms in varying sediment models.

<table>
<thead>
<tr>
<th>Method</th>
<th>$\tau_c$ equation</th>
<th>$d_{50}$ (m)</th>
<th>$\tau_c$</th>
</tr>
</thead>
<tbody>
<tr>
<td>In situ observation</td>
<td>$\rho C_D U^2$</td>
<td>10⁻⁵</td>
<td>$4.5 U^2 (C_D = 4.5 \times 10^{-3})$</td>
</tr>
<tr>
<td>FVCOM-SED</td>
<td>$\rho \max \left[ \dfrac{k^2}{\ln \left( \dfrac{\Delta z_{bot}}{z_p} \right)^2}, 0.0025 \right] U_{bot}^2$</td>
<td>10⁻⁵</td>
<td>$\max[0.00082, 0.0025] \rho U^2 = 2.5 U_{bot}^2$</td>
</tr>
<tr>
<td>Delft3D</td>
<td>2D flow $\rho \left( \dfrac{g}{18 \log_{10} \left( \dfrac{12h^3}{30z_c} \right)} \right)^{2} U_{bot}^2$ (White Colebrook)</td>
<td>10⁻⁵</td>
<td>$0.64 U_{bot}^2$</td>
</tr>
<tr>
<td></td>
<td>$\rho \left( \dfrac{g}{n^2 \left( \dfrac{A}{n} \right)^2} \right)^{2} U_{bot}^2$ (Manning)</td>
<td>10⁻⁵</td>
<td>$3849 n^2 U_{bot}^2 = 0.23 U_{bot}^2$</td>
</tr>
</tbody>
</table>

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\[
\mathbf{3D \ flow} \quad \frac{\kappa^2}{\ln (1 + \frac{\Delta z_{bot}^2}{2 z_o^2})} u_{bot}^2 \quad 10^{-5} \quad 0.9 u_{bot}^2
\]

<table>
<thead>
<tr>
<th>ELCD</th>
<th>[\rho \frac{0.03}{\log \left(\frac{12 \Delta z_{bot}}{k_s^*}\right)} u_{bot}^2]</th>
<th>(10^{-5})</th>
<th>(0.97 u_{bot}^2)</th>
</tr>
</thead>
</table>

| AEM3D | \[\rho C_D u_{bot}^2\] | \(10^{-5}\) | \(4.5 u_{bot}^2 \ (C_D = 4.5 \times 10^{-3})\) \(2.4 u_{bot}^2 \ (C_D = 2.4 \times 10^{-3})\) |

* \(\Delta z_{bot} = 1\) m  
# \(h = 16.5\) m  
† \(n\)  
‡ \(z_o\)  
⋆ \(k_s\)

The thickness of bottom layer in models 
The position of model output current velocities 
User-defin manning coefficient. Theoretically, \(n = 0.045(2.5 d_{50})^{1/6}\) (van Rijn, 1993). 
Roughness height of the lake bed (i.e. zero velocity level) [m] 
bed roughness, \(k_s = 30 z_o [m]\). Several relations between \(k_s\) and bottom sediment grain size have been proposed, with one of the most widely used being: \(k_s = 2.5 d_{50}\).

By comparing \(\tau_c\) computed from observations with that from the parameterizations in ELDC and AEM3D, we have found the overall underestimation of bottom stress based on the algorithms used in ELDC. These results are now discussed with reference to the parameterizations in other commonly-applied hydrodynamic models, specifically FVCOM-SED and Delft3D (Table 4.3). ELCD, FVCOM-SED and Delft3D all use logarithmic fitting for \(C_D\) based on \(k_s\) or \(z_o\) and both are associated with the bedform of the lake bed. For simulating basin-scale sediment resuspension, the setting of bathymetry is not fine enough to resolve the bedform of lake bed, thus, the grain sizes of bottom sediments are used in the calculation of \(C_D\) (Table 4.3). Ignoring the effect of bedform, \(C_D\) in these models are smaller than in situ observations (0.0045). To bring \(\tau_{c,m}\) equivalent to the in situ \(\tau_c\), \(d_{50}\) should be set to 0.03 m, which is not realistic. FVCOM-SED sets the minimum \(C_D = 0.0025\) which is close to the canonical value (0.0024), and it makes \(\tau_{c,m}\) in FVCOM-SED closest to the the practical values among these three models (Table 4.3; Morales-Marin et al., 2018; Niu et al., 2018). In the field-scale application of Delft3D, Hu et al. (2009) applied the \(\tau_{c,m}\) equation of 2D flow, and set the
Manning coefficient independent of \(d_{50}\) (Table 4.3), so that the model can reproduce sediment resuspension and transport correctly using literature-based \(\tau_{cr}\).

The ELCD results in 2008 showed that in west central basin of Lake Erie, \(\tau_{w,m}\) played a dominant role during intense storm events (Chapter 3). But Morales-Marin et al. (2018), who applied FVCOM-SED, modelled a much larger proportion of the lake bed to be potentially resuspended by currents during the extreme wind events in an upland shallow lake in the UK. One explanation of this discrepancy is that Lake Erie has a longer fetch and so can develop stronger surface waves and has a shallow water depth making lake bed vulnerable to surface waves. Another possible reason is the underestimation of the contribution from current because of the inappropriate algorithms in ELCD. Thus, applying AEM3D with Eq. (4.12), the magnitude of modeled stress was close to observed value, and we proved that bottom currents contributed to bottom stress comparable to surface wave during storms (e.g. R1, 13, 18; Figure 4.3-4.5) in the west central basin of Lake Erie (Sta. 341). But due to the special morphology of Lake Erie, strong surface waves during intense storm events (e.g. R23; Figure 4.5) still dominate the resuspension.

Thus, we provided a suggestive flow chart for the design of a sediment resuspension and transport module in a RANS model (Figure 4.13). With the in situ \(C_D\) or the canonical \(C_D\) of certain bottom type (Soulby et al., 1994; Zulberti et al., 2018), and accurate modeled bottom current velocities, models should apply Eq. (4.12) to calculate \(\tau_{c,m}\). Note that the value of \(C_D\) may require adjustment according to the thickness of bottom layer (\(\Delta z_{bot}\)) in model settings. Another option is the algorithm similar with FVCOM-SED, which picks the maximum value between the logarithmic derived \(C_D\) and the user-defined minimum \(C_D\) (\(C_{D,min}\)). This option is recommended when bed roughness \(z_o\) can be specified in the model.
Figure 4.13 Suggestive flow chart for parameterization of $C_D$ in sediment module of RANS model. $z_{ref}$ is the height above bottom where \textit{in situ} $C_D$ was collected, and for canonical $C_{D,ref}, z_{ref} = 1$ m.

4.4.2 Parameterization based on near-bed turbulence

The $\tau_{cr}$ defined by existing threshold models is mostly determined by flume experiments using mean current velocity profiles (Shields, 1936; Soulsby et al., 1997). However, on larger scales and in more complex systems (e.g. shallow marine environments and large lakes), the threshold could be reduced because of the enhanced intensity of intermittent turbulence events (Salim et al., 2018; Yang et al., 2016). Therefore, parameterizing $\tau_b$ from time-averaged current speeds has been shown to be inappropriate for modelling the bottom nepheloid layer generated by nonlinear internal wave-induced vertical velocities (Quaresma et al., 2007), resuspension from boundary layer instability and vortex shedding beneath waves propagating over a flat bottom (Bourgault et al., 2014; Aghsae and Boegman, 2015), as well as the turbid hypolimnion generated by HFIWs in this study. In these cases, $\tau_L$ and $\tau_{TKE}$ are more appropriate because both methods parameterize near-bed turbulence (Eq. (4.5), (4.9)).

Existing RANS models are unable to resolve $w'$, so parameterization based on the TKE method is unrealistic. Present algorithms for $\varepsilon_m$ in RANS models (e.g., AEM3D) do not consider the energy flux path associated with surface wave generation and breaking (e.g., Spigel
et al., 1986; Hodges et al., 2000), leading to overestimation of the energy flux entering the lake interior most of the time (Figure C2). Figure 4.11 shows that \( \tau_{L,m} \) was smaller than the observed \( \tau_L \), only when bottom stresses were mainly from surface wave orbital velocities (R4, 7, 9, 12, 21-23). Therefore, complete replacement of the present parameterization (Eq. (4.9), (4.13)) with \( \tau_{L,m} \) is not suitable for shallow water systems with resuspension frequently triggered by surface waves. However, the log-law parameterization should be further investigated in water systems with intensive convective turbulence (Anderson et al., 1979), where shear-driven models are not appropriate.

**4.5 Conclusions**

Multiple parameterization methods for bottom stress (\( \tau_b \)), including from summation of surface waves stress (\( \tau_w \)) and mean currents (quadratic) stress (\( \tau_c \)), log-law (\( \tau_L \)), and turbulent kinematic energy (\( \tau_{TKE} \)), have been assessed in this study, based on observed data and model output. For large and shallow natural water systems, like Lake Erie, bottom currents and surface wave orbitals were the two major processes driving bottom resuspension. Sub-grid-scale HFIWs also induce low-intensity resuspension events when the seasonal thermocline became close to the lake bed, forming a turbid hypolimnion. The bottom stress, estimated as \( \tau_b = \tau_w + \tau_c \) was sufficient to predict the resuspension induced by surface waves and bottom currents, but only \( \tau_L \) and \( \tau_{TKE} \) were able to capture the turbid bottom layer generated by the sub-grid-scale motions.

Overall, the combination of \( \tau_{w,m} \) and \( \tau_{c,m} \) was still the most practical parameterization for sediment resuspension simulations in field-scale RANS models, but uncertainties about \( C_D \) remain. This study assessed different algorithms for \( \tau_{c,m} \) and proposed an optimal solution based on validation against observed \( \tau_c \). By following the suggestive flow chart for \( \tau_{c,m} \) calculation,
future modelers can model sediment resuspension according *in situ* or a literature-based resuspension threshold.

Although the sub-grid-scale turbulent fluctuations are not reproduced in RANS models, a possibility of using model output $\varepsilon_m$ to parameterize $\tau_L$ and simulate sediment resuspension was also investigated, and may be more capable of simulating bottom nepheloid layers generated by HFIWs. However, many models do not account for TKE dissipation associated with localized processes, like surface wave generation and breaking, which could lead to inaccurate $\varepsilon_m$. Therefore, the log-law parameterization method should be further tested and improved by better parameterization of $\varepsilon$. 
Chapter 5

Conclusions and Recommendations for future research

Chapter 2 revealed the regional characteristic intensity of turbulence and the level of vertical mixing in Lake Erie during the springs and summers 1997 and 2008-2009. Chapter 3 applied a field-scale three-dimensional RANS model to qualitatively simulate the basin-scale sediment resuspension in Lake Erie. Chapter 4 improved prediction of sediment resuspension in observations and RANS models, by comparing multiple parameterizations of bottom stress. The main conclusions of this study are summarized below:

1. The western basin of Lake Erie has the highest average $\varepsilon$ and least variation with depth due to its shallowness. The presence of seasonal stratification affects the level of $\varepsilon$ and $K_Z$, and makes averaged $\varepsilon$ and $K_Z$ profiles more quantifiable. During late summer, $K_Z$ decreases to molecular diffusivity levels in the bottom boundary layer at the transition to the central basin when the seasonal thermocline intersects the lakebed. $\varepsilon$ increases in the metalimnion of the central and eastern basins because of shear instability generated by HFIWs. Also, thermocline entrainment, wind-induced internal seiche currents under the thermocline, and hypolimnetic return flows increase $\varepsilon$ and $K_Z$ in the hypolimnion of the central and eastern basins.

2. The TKE budget and vertical mixing in Lake Erie are controlled by its morphometry, seasonal stratification and wind forcing. Combined with existing literature from other lakes, this study found that Lake Erie dissipated more wind energy beneath the surface mixed layer than median- and small-sized lakes, especially during strong wind forcing. In general, mixing
beneath the surface layer increased with lake surface area, but also decreased with increasing Wedderburn number (weaker winds and stronger stratification).

3. When calibrated against a three-month observational dataset, application of a field-scale RANS model enabled the qualitative simulation of basin-scale sediment resuspension and quantitative simulation of background TSS concentration in Lake Erie.

4. Model results showed surface waves predominated ($\tau_w = 88 \% \tau_b$) sediment resuspension in the shallow western basin. During a storm event, $\tau_w$ increased significantly, up to $\sim 0.25$ Pa. The relative importance of bottom mean-current increased with depth, reaching $\tau_c = 48 \% \tau_b$ in the western part of the central basin, where up- and down-welling events and internal Poincaré waves contributed to $\tau_c$. Conversely, resuspension in the deeper regions (> 25 m) of the central and eastern basins was not modelled. In general, the model identified the regions with active resuspension (western, west-central and north-central basins) where biogeochemical fluxes may be enhanced.

5. Using field observations, four parameterizations of $\tau_b$ were assessed. These had different abilities in predicting various sediment resuspension mechanisms. Parameterizations based on near-bed turbulent parameters ($\tau_L$ and $\tau_{TKE}$) were able to predict resuspension induced by increased bottom current velocities, increased surface wave orbital velocities, as well as HFIWs carried by a near-bed thermocline. Parameterizations based on the quadratic stress law and surface waves could only predict events triggered by the first two mechanisms.

6. Summation of current-induced bottom stress $\tau_c$ and surface wave-induced bottom stress $\tau_w$ was the most practical parameterization of $\tau_b$ in existing RANS equation models. When compared with observed values based on the same parameterizations, we assessed the performance of different algorithms of $\tau_w$ and $\tau_c$ in commonly applied RANS models. Even
though turbulence-based parameterizations ($\tau_L$ and $\tau_{TKE}$) were associated with sub-grid-scale processes, making them unrealistic in field-scale RANS models, we investigated the possibility of applying $\tau_L$ through parameterizing $\varepsilon$ in RANS models. Given that $\varepsilon$ reproduced by the mixed-layer parameterization in the model neglects localized processes, such as surface wave breaking, the parameterization of $\varepsilon$ should be further calibrated and validated before being applied to calculate $\tau_L$.

This thesis not only benefits the understanding of small-scale hydrodynamic processes within Lake Erie, but helps to understand what drives biogeochemical fluxes in the lake. Our observations were in spring and summer, focusing on the features related to seasonal thermocline. Future observations should span longer times, so that the inter-annual changes and processes such as during the ice-cover season can be investigated.

The application of a three-dimensional RANS model allowed us to visualize basin-scale sediment resuspension and transport, especially at the locations where there were no observations. Model results identified the interactions between hydrodynamic processes and sediment layers. However, quantitative simulation of sediment resuspension events requires water samples collected during resuspension events; a challenging task! Considering the effect of complex topography and the spatial variation in the bottom material should be added into future models.
References


Boegman, L., 2006. A model of the stratification and hypoxia in central Lake Erie. s.l., IAHR, Univ. of Western Australia, pp. 608-613.


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Appendix A

Text A1 Batchelor fitting

The Batchelor spectrum of temperature gradient fluctuations can be written as

$$S(k) = \left(\frac{q}{2}\right)^2 \chi_\theta f(\alpha) \frac{k_B D}{k_B D}$$

where $\chi_\theta$ is the temperature variance dissipation rate given by

$$\chi_\theta = 6D \int_0^\infty S(k)dk = 6D \left(\frac{\partial T'}{\partial z}\right)^2$$

Here, $k$ is the wavenumber, $q = 2\sqrt{3}$ is a universal constant, $D$ is the molecular diffusivity, $T'$ is the temperature fluctuation, $z$ is the depth, $k_B$ is the Batchelor cutoff wave number $(\epsilon \nu^{-1}D^{-2})^{1/4}$, $\epsilon$ is the dissipation rate of turbulent kinetic energy, $\nu$ is the kinematic viscosity, and $\alpha$ is a nondimensional wavenumber. The function $f(\alpha)$ is defined as

$$f(\alpha) = \alpha \left( e^{-\frac{\alpha^2}{2}} - \alpha \int_{\alpha}^{\infty} e^{-\frac{x^2}{2}} dx \right)$$

By assuming that the cutoff wavenumber, at which the spectrum falls to one-tenth of its peak value, corresponds to the theoretical Batchelor wavenumber, we calculate a dissipation rate as (Dillon and Caldwell, 1980),

$$\epsilon_B = k_B^4 \nu D^2$$

Table A1 Bounds of $Re_b$ and formulas for $K_\varepsilon$ and $\Gamma$ (Bouffard and Boegman 2013)

<table>
<thead>
<tr>
<th>Regime</th>
<th>Molecular</th>
<th>Buoyancy-controlled</th>
<th>Transitional</th>
<th>Energetic</th>
</tr>
</thead>
<tbody>
<tr>
<td>$Re_b$ range ($Pr = 7$)</td>
<td>$Re_b &lt; 1.7$</td>
<td>$1.7 &lt; Re_b &lt; 8.5$</td>
<td>$8.5 &lt; Re_b &lt; 100$</td>
<td>$Re_b &gt; 100$</td>
</tr>
<tr>
<td>$K_\varepsilon$ estimate</td>
<td>$D$</td>
<td>$0.0615\nu Re_b^{3/2}$</td>
<td>$0.2\nu Re_b$</td>
<td>$2\nu Re_b^{1/2}$</td>
</tr>
<tr>
<td>$\Gamma$ estimate</td>
<td>$0$</td>
<td>$0.0615 \frac{\varepsilon^{1/2}}{\nu^{1/2}N}$</td>
<td>$0.2$</td>
<td>$2\nu^{1/2} \frac{N}{\epsilon^{1/2}}$</td>
</tr>
</tbody>
</table>

* $D = 1.4 \times 10^{-7}$ is molecular diffusivity
Text A2 Occurrence frequency of $Re_b$, $\Gamma$, and $K_z$

In Group W, the distribution of $Re_b$ in the Bouyancy-controlled, Transitional, and Energetic regimes (Fig. A1a) were 13%, 30%, 52%, respectively, indicating the turbulent intensity was high. Histograms of $Re_b$, $\Gamma$ and $K_z$, in the metalimnion (Fig. A1m-r) of Group C, show that during seasonal stratification the distribution of $Re_b$ is shifted toward laminar flow (15%, 30%, 33% and 21% of segments in the Molecular, Buoyancy-controlled, Transitional, and Energetic Regimes); whereas, during the neutrally stratified period, $Re_b$ is shifted toward energetic turbulence (3%, 7%, 17% and 73% of segments in the Molecular, Buoyancy-controlled, Transitional, Energetic Regimes). During seasonal stratification, 1/3 of the segments were in the Transitional Regime, with $\Gamma = 0.2$ (the commonly adopted oceanic value), but most segments during neutral stratification were in the Energetic Regime where $\Gamma$ decreases with increasing $Re_b$ (Bouffard and Boegman 2013; Monismith 2018). $Re_b$ in Group E was generally high (Fig. A1j-l), with over 80% of segments in the Energetic regime and high $K_z$, which unexpectedly differs from Group C in the seasonally stratified period.
Figure A1 Frequency of occurrence (%) of $Re_b$, $\Gamma$ and $K_z$ in Group W (a-c), neutrally stratified (d-f) and seasonally stratified (g-i) periods of Group C, Group E (j-l), and metalimnion of neutrally stratified (m-o) and seasonally stratified (p-r) Group C respectively. Dash lines in (a, d, g, j, m, p) are bounds of 4 regimes in Table 2. Notice that the right bound of Energetic regime extends outside the x-axis.
Table A2 Typical $\varepsilon$ and $K_z$ values, with corresponding Kolmogorov length, time, and velocity scales calculated in the present study and former researches in different water layers of various lakes.

<table>
<thead>
<tr>
<th></th>
<th>Lake</th>
<th>Baikal</th>
<th>Michigan</th>
<th>Erie</th>
<th>Ontario</th>
<th>Constance</th>
<th>Kinneret</th>
<th>Mono</th>
<th>Opeongo</th>
<th>Alpnach</th>
</tr>
</thead>
<tbody>
<tr>
<td>Surface layer</td>
<td>$\varepsilon \times 10^{-8}$ (m$^2$ s$^{-3}$)</td>
<td>0.1-50</td>
<td>/</td>
<td>100</td>
<td>/</td>
<td>/</td>
<td>&gt;100</td>
<td>/</td>
<td>10</td>
<td>/</td>
</tr>
<tr>
<td></td>
<td>$K_z \times 10^{-5}$ (m$^2$ s$^{-1}$)</td>
<td>&gt;50</td>
<td>/</td>
<td>10</td>
<td>/</td>
<td>/</td>
<td>/</td>
<td>/</td>
<td>10-100</td>
<td>/</td>
</tr>
<tr>
<td></td>
<td>$L_\eta = (\frac{\varepsilon}{\varepsilon})^{\frac{1}{2}}$ (mm)</td>
<td>1.2-5.6</td>
<td>/</td>
<td>1</td>
<td>/</td>
<td>/</td>
<td>&lt;1</td>
<td>/</td>
<td>1.8</td>
<td>/</td>
</tr>
<tr>
<td></td>
<td>$T_\eta = (\frac{\nu}{\varepsilon})^{\frac{1}{2}}$ (s)</td>
<td>1.4-3.1</td>
<td>/</td>
<td>1</td>
<td>/</td>
<td>/</td>
<td>&lt;1</td>
<td>/</td>
<td>3.2</td>
<td>/</td>
</tr>
<tr>
<td></td>
<td>$U_\eta = (\nu\varepsilon)^{1/4}$ (mm s$^{-1}$)</td>
<td>0.18-0.84</td>
<td>/</td>
<td>1</td>
<td>/</td>
<td>/</td>
<td>&gt;1</td>
<td>/</td>
<td>0.56</td>
<td>/</td>
</tr>
<tr>
<td>Metalimnion</td>
<td>$\varepsilon \times 10^{-8}$ (m$^2$ s$^{-3}$)</td>
<td>0.1-100</td>
<td>/</td>
<td>1-100</td>
<td>/</td>
<td>/</td>
<td>/</td>
<td>0.1-1000</td>
<td>&lt;1 (offshore) 100 (inshore)</td>
<td>0.05-0.5</td>
</tr>
<tr>
<td></td>
<td>$K_z \times 10^{-5}$ (m$^2$ s$^{-1}$)</td>
<td>200</td>
<td>/</td>
<td>1-10</td>
<td>/</td>
<td>/</td>
<td>0.5</td>
<td>0.001 (offshore) 10 (inshore)</td>
<td>0.01</td>
<td>/</td>
</tr>
<tr>
<td></td>
<td>$L_\eta = (\frac{\varepsilon}{\varepsilon})^{\frac{1}{2}}$ (mm)</td>
<td>1.5-6</td>
<td>/</td>
<td>1-3.2</td>
<td>/</td>
<td>/</td>
<td>0.56-5.6</td>
<td>&gt;3.2 (offshore) 1 (inshore)</td>
<td>3.8-6.9</td>
<td>/</td>
</tr>
<tr>
<td></td>
<td>$T_\eta = (\frac{\nu}{\varepsilon})^{\frac{1}{2}}$ (s)</td>
<td>1-3.1</td>
<td>/</td>
<td>1-10</td>
<td>/</td>
<td>/</td>
<td>0.32-32</td>
<td>&gt;10 (offshore) 1 (inshore)</td>
<td>14-45</td>
<td>/</td>
</tr>
<tr>
<td></td>
<td>$U_\eta = (\nu\varepsilon)^{1/4}$ (mm s$^{-1}$)</td>
<td>0.18-1</td>
<td>/</td>
<td>0.32-1</td>
<td>/</td>
<td>/</td>
<td>0.18-1.8</td>
<td>&lt;0.32 (offshore) 1 (inshore)</td>
<td>0.15-0.27</td>
<td>/</td>
</tr>
<tr>
<td>Bottom layer</td>
<td>$\varepsilon \times 10^{-8}$ (m$^2$ s$^{-3}$)</td>
<td>/</td>
<td>10</td>
<td>10</td>
<td>1-10</td>
<td>10</td>
<td>10</td>
<td>/</td>
<td>/</td>
<td>4</td>
</tr>
<tr>
<td></td>
<td>$K_z \times 10^{-5}$ (m$^2$ s$^{-1}$)</td>
<td>/</td>
<td>/</td>
<td>10</td>
<td>0.1-50</td>
<td>75</td>
<td>10-100</td>
<td>/</td>
<td>/</td>
<td>33</td>
</tr>
<tr>
<td></td>
<td>$L_\eta = (\frac{\varepsilon}{\varepsilon})^{\frac{1}{2}}$ (mm)</td>
<td>/</td>
<td>1.8</td>
<td>1.8</td>
<td>1.8-3.2</td>
<td>1.8</td>
<td>1.8</td>
<td>/</td>
<td>/</td>
<td>2.2</td>
</tr>
<tr>
<td></td>
<td>$T_\eta = (\frac{\nu}{\varepsilon})^{\frac{1}{2}}$ (s)</td>
<td>/</td>
<td>3.2</td>
<td>3.2</td>
<td>3.2-10</td>
<td>3.2</td>
<td>3.2</td>
<td>/</td>
<td>/</td>
<td>5</td>
</tr>
<tr>
<td></td>
<td>$U_\eta = (\nu\varepsilon)^{1/4}$ (mm s$^{-1}$)</td>
<td>/</td>
<td>0.56</td>
<td>0.56</td>
<td>0.32-0.56</td>
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<td>0.56</td>
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<td>0.44</td>
</tr>
</tbody>
</table>
## Appendix B

Table B1 Details of stations and instrument deployments in 2008. See Figure 1 for a map of the stations.

<table>
<thead>
<tr>
<th>Station</th>
<th>Latitude</th>
<th>Longitude</th>
<th>Water depth (m)</th>
<th>Instrument</th>
<th>Day of Year 2008</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, temperature, pressure)</td>
<td>212-288</td>
<td>16.5</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>XR-420 (Turbidity, temperature, pressure)</td>
<td>212-288</td>
<td>16</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>Pumped water sample</td>
<td>204, 205, 213, 219, 220, 221</td>
<td>12.15</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>Superficial bed sediment sample (PONAR)</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>Nortek Vector ADV</td>
<td>214-265</td>
<td>16.5</td>
</tr>
<tr>
<td>1227</td>
<td>41°48’37’’</td>
<td>82°30’11’’</td>
<td>11.4</td>
<td>Pumped water sample</td>
<td>203, 205, 214, 219, 221</td>
<td>0.5, 3.7, 6.7, 9.7, 10.7, 11.2</td>
</tr>
<tr>
<td>1228</td>
<td>41°47’53’’</td>
<td>82°20’54’’</td>
<td>14</td>
<td>YSI-6600 Turbidity</td>
<td>155-275</td>
<td>13.5</td>
</tr>
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<td></td>
<td></td>
<td></td>
<td></td>
<td>Pumped water sample</td>
<td>204, 205, 213, 219, 220, 221</td>
<td>1, 3, 7, 10, 13, 14, 14.5</td>
</tr>
<tr>
<td>1231</td>
<td>41°47’26’’</td>
<td>82°11’26’’</td>
<td>19.8</td>
<td>Pumped water sample</td>
<td>204, 212, 219, 220, 221</td>
<td>1, 5, 8, 12, 15, 18, 19</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>Superficial bed sediment sample (PONAR)</td>
<td>204</td>
<td></td>
</tr>
<tr>
<td>357</td>
<td>41°49’57’’</td>
<td>82°58’05’’</td>
<td>10.4</td>
<td>Pumped water sample</td>
<td>214</td>
<td>1, 2, 5, 8, 9, 9.6</td>
</tr>
<tr>
<td>84</td>
<td>41°56’12’’</td>
<td>81°38’70’’</td>
<td>24.8</td>
<td>Pumped water sample</td>
<td>212</td>
<td>1, 8, 12, 16, 19, 22</td>
</tr>
<tr>
<td>452</td>
<td>42°34’44’’</td>
<td>79°55’25’’</td>
<td>54</td>
<td>Pumped water sample</td>
<td>211</td>
<td>1, 10, 18, 20, 25, 38, 49.5, 52.5</td>
</tr>
</tbody>
</table>
### Table B2 Governing hydrodynamic equations in ELCOM

| Transport of momentum | \[
\frac{\partial U_{\alpha}}{\partial t} + U_j \frac{\partial U_{\alpha}}{\partial x_j} = -g \left( \frac{\partial \eta}{\partial x_{\alpha}} + \frac{1}{\rho_0} \frac{\partial}{\partial x_{\alpha}} \int_{z'} \rho' dz' \right) + \frac{\partial}{\partial x_1} \left( v_1 \frac{\partial U_{\alpha}}{\partial x_1} \right) + \frac{\partial}{\partial x_2} \left( v_2 \frac{\partial U_{\alpha}}{\partial x_2} \right) \\
+ \frac{\partial}{\partial x_3} \left( v_3 \frac{\partial U_{\alpha}}{\partial x_3} \right) - \epsilon_{\alpha\beta} f U_{\beta}
\] |
| Continuity | \[
\frac{\partial U_j}{\partial x_j} = 0
\] |
| Transport of scalars | \[
\frac{\partial C}{\partial t} + \frac{\partial}{\partial x_j} (CU_j) = \frac{\partial}{\partial x_1} \left( \kappa_1 \frac{\partial C}{\partial x_1} \right) + \frac{\partial}{\partial x_2} \left( \kappa_2 \frac{\partial C}{\partial x_2} \right) + \frac{\partial}{\partial x_3} \left( \kappa_3 \frac{\partial C}{\partial x_3} \right) + S_c
\] |
| Free-surface evolution | \[
\frac{\partial \eta}{\partial t} = -\frac{\partial}{\partial x_{\alpha}} \int_b^n U_\alpha dz
\] |
| Free-surface wind shear | \[
(u_*)_\alpha^2 = C_{10} \frac{\rho_a}{\rho_w} (W_\beta W_\beta)^{\frac{1}{2}} W_\alpha
\] |
| Momentum input by wind | \[
\frac{dU_{\alpha}}{dt} = \frac{(u_*)_\alpha^2}{h}
\] |

- $i, j, k, m$: Three component space (e.g. $i = 1, 2, 3$)
- $\alpha, \beta$: Horizontal two component space (e.g. $\alpha = 1, 2$)
- $U, \eta$: Reynolds-averaged velocity and free surface height
- $\rho_0$: Reference density
- $\rho'$: Density anomaly
- $f$: Coriolis constant
- $h$: Height of wind-mixed layer
- $\epsilon_{\alpha\beta}$: Two-component permutation tensor
- $\nu$: Molecular viscosity
- $S_c$: Turbulent Schmidt number (or Prandtl number for temperature)
- $C$: Scalar concentration
- $W_\beta$: Vector wind speed in $\beta$ direction
- $C_{10}$: Bulk wind stress coefficient for wind values at 10 meters
- $(u_*)_\alpha$: Wind shear velocity in $\alpha$ direction
### Table B3 Calculation of wave properties in ELCOM-CAEDYM

<table>
<thead>
<tr>
<th>Property</th>
<th>Formula</th>
</tr>
</thead>
<tbody>
<tr>
<td>Wave period (T_s)</td>
<td>(T_s = 2.4\pi \frac{W}{g} \tanh \left( \frac{0.0379 \left( \frac{gF}{W^2} \right)^{0.333}}{\tanh(0.833 \left( \frac{gh}{W^2} \right)^{0.375})} \right) )</td>
</tr>
<tr>
<td>Wave height (H_s)</td>
<td>(H_s = 0.283 \frac{W^2}{g} \tanh(0.53(\frac{gh}{W^2})^{0.75}) \tanh(\frac{0.00565 \left( \frac{gh}{W^2} \right)^{0.5}}{\tanh(0.53 \left( \frac{gh}{W^2} \right)^{0.75})}) )</td>
</tr>
<tr>
<td>Wave length (L)</td>
<td>(L \approx L_0 \sqrt{\tanh(\frac{2\pi h}{L_0})}; \quad L_0 = \frac{gT^2}{2\pi} )</td>
</tr>
<tr>
<td>Maximum bottom orbital velocity (U_{orb})</td>
<td>(U_{orb} = \frac{\pi H_s}{T_s \sinh(\frac{2\pi h}{L})} )</td>
</tr>
</tbody>
</table>

\(F\) Wind fetch  
\(W\) Wind speed  
\(h\) Water depth  
\(g\) Gravitational acceleration

---

**Text B1 Sediment transport in the north-central basin**

Hydrodynamic forcing (currents and turbulence) are stronger in nearshore areas than offshore (Bouffard et al., 2012); particularly the northern and eastern shorelines of the central and eastern basins (Thomas et al., 1976; Rao and Schwab, 2007). Strong hydrodynamic forcing causes active sediment dynamics, including resuspension, shoreline erosion and sediment transport. Our model captures these dynamics near the northern shoreline of the central basin and Long Point. In this region, bluff erosion is a major sediment source, accounting for 40% of material (Kemp et al., 1977); however, bluff erosion is not considered in our model. Here, we investigated the potential for bottom sediment to be transported toward Long Point from storm induced resuspension.

Model output of the TSS concentration and currents in the bottom layer and surface layer, during the storm on days 302–304, are shown in Figure B3a, b. High TSS concentration (15-20mg L\(^{-1}\)) is modelled in both the bottom and surface layers of the northern region of the central basin,
where observed sediments consisted of sand, glacial till and mud (Haltuch et al., 2000). Strong offshore currents (> 0.2 m s\(^{-1}\)) are simulated in the bottom layer (Figure B4b), which were expected as the result of southwest wind and Coriolis force, and an overall eastward current transport in surface layer (Figure B3a). Supporting our results, Valipour et al. (2017) reported elevated turbidity events when observed near-bed currents >0.25 m s\(^{-1}\). We suspected that resuspension is frequent in this region due to the shallowness along the shoreline area and long fetch. The sediments are simulated to be lifted into the upper water column and transported eastward by currents in surface layer toward Long Point and deeper regions of the eastern basin (Figure B3b). Lick et al. (1994) and Hawley and Eadie (2007) found that deposition in the eastern basin was almost entirely due to very large but infrequent storms that transport suspended material, eroded from the near-shore lake bed. Our simulation does not span winter; therefore, we do not model the significant large storms, but the steady southwest winds of ~10 m s\(^{-1}\) on days 302-304 are within the range of 10-20 m s\(^{-1}\) indicated by Hawley and Eadie (2007) as typical storms which rework sediments. So the events reported in Figure B3 are more likely to repeat resuspension-settlement processes and not contribute to significant transport. The Long Point sediment budget shows a net gain of 500 m\(^3\) during May 1989 to September 1990 (Davison-Arnott and Van Heyningen, 2003) and the central eastern basin is a principal deposition zone in Lake Erie (Thomas et al., 1976). If the largest storms are responsible for sediment transport to the eastern basin, these lesser storms may resuspend bottom sediment and supply TSS in the epilimnion that can be transported by larger storms. Given the lack of recent observational data along the northern central basin, the fate of sediments resuspended here should be further studied. But it is noted that the shape of high TSS area in north shoreline is similar with shoreline erosion inputs regions described in (Kemp, 1977; Figure 4)
Figure B1 Time-series of observed sediment loading in the Maumee River during 2008 (data is from USGS; https://cida.usgs.gov/sediment/).

Figure B2 Time-series of modelled and observed (5 min-averaged) current velocities 1 m above bottom (16.5 m depth at Sta. 341, Table B1). Positive velocities in the figure present northward and eastward velocities.
**Figure B3** Model output of SSR (suspended solid from rivers) in the bottom layer (a. day 224; b. day 240; c. day 250; d. day 260; e. day 290; f. day 303). The dates correspond to Figure 3.6.
Figure B4 Model output of TSS concentration at the (a) surface and (b) bottom on day 303, red arrows were shown every four grids indicating the magnitude and direction of 48h averaged currents.
Figure B5 (a) Time-series of potential energy and (b) continuous wavelet transforms showing the existence of Poincaré waves at Sta. 1228 during days 205 to 240. Typical period of Poincaré wave is ~17 h. The diurnal period ~24 h is shown, with strong peaks during seasonal stratification that are weaker in fall.

Figure B6 Averaged lake-wide $\tau_c$ during the simulation.
Movie B1 Model output animation of $SS_R$ in bottom layer from day 255 to 265.

Movie B2 Model output animation of $SS_B$ in bottom layer from day 299 to 304.

Movie B3 Central basin curtain (see Figure 3.1) showing vertical motion of thermocline and corresponding sediment dynamics from day 258 to 272. The color bar shows TSS concentration and black lines are isotherm contours.
Appendix C

Figure C1 Spectral density of vertical velocity measurements from ADV at 1 m above the bed on day 120.76 (blue lines); the black dash lines show the inertial subrange $-5/3$ slope.

**Text C1 inertial fitting**

Spectral densities of $w'$ are obtained using Welch’s method. For each burst, data points were partitioned into equal length segments (each 512 data), and each segment was windowed with a Hanning window. The frequencies are them converted to wavenumber by dividing $U$ in accord with Taylor’s hypothesis of ‘frozen turbulence’, where $U$ is the mean horizontal velocity.
Figure C2 Time-series at Sta. 341 of modeled $\epsilon_m$ and observed $\epsilon_{ID}$ during selected period in 2008 and 2009.