Sediment resuspension mechanisms and their contributions to high-turbidity events in a large lake

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Abstract

High-resolution field data, collected during April to October of 2008–2009, were analyzed to investigate the quantitative contribution of sediment resuspension to high-turbidity events in central Lake Erie. Resuspension events were distinguished within high-turbidity events according to turbidity, fluorescence and acoustic backscatter timeseries, as well as satellite images. We observed 16 high-turbidity events, causing a total duration of $\approx 20$ d (out of 344 d) with elevated nearbed turbidity ($> 10$ NTU). Of these events, 64% were correlated with algal biomass, with the remaining 18%, 5%, and 4% being attributed to sediment resuspension by surface waves, storm-generated currents and enhanced nearbed turbulence induced by high-frequency internal waves, respectively. This is the first time that resuspension by enhanced nearbed turbulence from high-frequency linear internal wave degeneration has been observed in a large lake. Resuspension was parameterized as a function of the instantaneous critical bottom velocity, bottom shear stress and the Shields parameter. From the in situ measurements, we suggest an extended Shields diagram for silty bed material that can be used to predict resuspension in other aquatic systems with similar sediment composition ($\approx 20\%$ cohesive sediment).

Sediments in aquatic systems are resuspended into the watercolumn when the bottom shear stress becomes greater than the critical shear stress for the initiation of suspension (Van Rijn 1993). During resuspension, turbulent eddies overcome the settling velocity and lift sediments above the bottom (Bagnold 1966). Resuspended sediments influence aquatic biogeochemistry by increasing the turbidity and thereby changing the rate of photosynthesis, as well as the vertical distributions of biomass, nutrients and contaminants (Fréchet et al. 1989; Gloor et al. 1994; Lou et al. 2000). In lakes experiencing hypoxia, such as Lake Erie (e.g., Rao et al. 2008), resuspension of organic biomass can increase the sediment oxygen demand by enhancing the surface area of decaying organic matter (Ackerman et al. 2001; Lorke and MacIntyre 2009). Resuspension by physical processes may originate from storm-driven currents (Lick et al. 1994; Beletsky et al. 2003; Churchill et al. 2004; Hawley and Eadie 2007; Marti and Imberger 2008), surface waves (Lou et al. 2000; Hawley et al. 2004), and/or shoaling and breaking of packets of high-frequency internal waves (HFIWs; Hawley 2004). HFIW-induced resuspension results from progressive nonlinear internal wave (e.g., solitary wave) shoaling in lab experiments and the coastal ocean (e.g., Quresma et al. 2007; Stastna and Lamb 2008; Boegman and Ivey 2009); although there is also the potential for resuspension from shoaling of both obliquely propagating and convectively unstable linear modes (e.g., Kelvin–Helmholtz billows; e.g., Bouffard et al. 2012).

In the Laurentian Great Lakes, storm-induced mean currents, from winds with speeds of 15–20 m s$^{-1}$, frequently resuspend bottom sediments (Lick et al. 1994; Beletsky et al. 2003; Churchill et al. 2004; Hawley and Eadie 2007). In shallow systems, surface waves’ orbital velocities have the potential to impinge on the bed leading to resuspension (Hawley 2000; Hawley et al. 2004). The simultaneous action of storm-induced currents and surface waves lead to an increase in the
bottom shear stress, which may exceed the critical shear stress and cause resuspension (Hawley and Lesht 1992; Hawley 2000; Lou et al. 2000; Churchill et al. 2004; Hawley et al. 2004). In the nearshore zone of the Laurentian Great Lakes, and shallow lakes and ponds, it is well-established that surface waves have a more pronounced contribution to resuspend sediments than mean currents (Luettich et al. 1990; Hawley 2004; Mian and Yanful 2004; Chung et al. 2009; Reardon et al. 2014). Shoaling of HFIWs has been observed in lakes (MacIntyre et al. 1999; Boegman et al. 2003; Lorke et al. 2006; Dorostkar et al. in press); although their influence on resuspension remains speculative (Hawley 2004). Linear HFIW modes can generate nearbed patches of turbulence with the ability to resuspend sediment either in the form of a turbulent patch propagating toward the bed (Boegman 2009), or critical breaking of obliquely propagating HFIWs that are reflected off the bottom (Imberger 1998; Ivey et al. 2000).

The objective of the present study is to examine high-resolution field data from central Lake Erie, to identify and quantify the occurrence of resuspension by storm-driven currents, surface waves, and HFIWs. We describe the mechanisms driving resuspension with particular emphasis on the enhanced nearbed turbulence by HFIWs. We parameterize resuspension events in terms of the observed instantaneous critical velocity (1 m above the bed), bottom stress and Shields parameter. Using in situ observations, an extended Shields diagram for silty bed materials is proposed, which can be generalized to model resuspension throughout Lake Erie and other aquatic systems with similar sediment composition.

**Measurements and methods**

**Study area**

Lake Erie (Fig. 1; 388 km long and 92 km wide) is the shallowest of the Laurentian Great Lakes and consists of distinct western, central and eastern basins, which have maximum depths of −11 m, −25 m, and −64 m, respectively (herein depth is positive upward). The inertial period in the lake is 18 h (42° latitude), giving an inertial frequency \( f = 0.97 \times 10^{-4} \text{ rad s}^{-1} \). During seasonal stratified periods, there is a basin-scale near-inertial (Poincaré) waves cell in the central basin where the present moorings are located (Valipour et al. 2015b).

**Measurements**

In April-October of 2008–2009, extensive field measurements (Fig. 1; Table 1) were carried out in central Lake Erie to address the hypoxia problem (see also, Bouffard et al. 2012, 2013, 2014; Bouffard and Boegman 2013; Valipour et al. 2015a,b). These included high temporal resolution temperature, velocity, turbidity, fluorescence chlorophyll \( a \) (Chl-\( a \)), and pressure measurements (Table 1). Water temperatures were recorded at station (Sta.) 341 using temperature loggers (TR-1060, RBR Ltd., accuracy ±0.002°C). A tripod was also deployed on the lakebed at a depth of −17.5 m at Sta. 341 (Valipour et al. 2015a; Fig. 1 therein). The tripod was equipped with upward and downward looking acoustic Doppler current profilers (ADCP; Nortek Aquadopp, accuracy ±1% of measured values ≥0.005 m s\(^{-1}\)) at 1.8 m
above the bottom (mab). The upward looking 600 kHz ADCP burst sampled velocity every 15 min at 1 Hz to obtain 180 s ensemble averages in 1 m bins to the surface and the downward looking pulse-coherent 2 MHz (hereafter, HR-ADCP) burst sampled velocity every 15 min at 1 Hz over 256 s in 3 cm bins to the bed. A Nortek Vector acoustic Doppler velocimeter (ADV), was moored on the tripod at 1 mab and burst recorded velocity at 16 Hz for 5 min every 20 min. Multi-parameter water quality sondes (XR-620 and XR-420; RBR Ltd.) were equipped with autoranging Seapoint turbidity and fluorescence sensors (accuracy ±2% measured values) and deployed at 1.5 mab on the tripod (XR-620) and on a nearby (≈30 m) mooring line at 5 mab; this mooring also supported the RBR TR-1060 loggers (Table 1). To prevent biofouling, the sensor housings were covered with copper tape. Meteorological data were obtained from an Environment and Climate Change Canada MET3 buoy at Sta. 341 that recorded average wind speed and direction. Significant wave height ($H_s$), period ($T_s$) and mean wave direction data for both years were obtained from station NDBC-45005 located 15 km to the south-west of Sta. 341 (Fig. 1). In 2009, the tripod was also equipped with a bottom mounted wave logger (TWR-2050, RBR Ltd, ±0.05% measured values); enabling limited surface wave measurements (only surface waves whose orbitals impinge on the bed). Additional YSI 6600 sondes with wiped turbidity sensors (Yellow Springs Instruments, accuracy ±2% measured values) were deployed at Sta. 1227 (1.5 mab), 1229 (1 mab), 1231 (1 mab), and 1233 (1 mab), which were located ~15 km to ~27 km from Sta. 341 (Fig. 1), thus enabling investigation of the spatial variability in bottom turbidity observations.

### Sediment size, density, and type

Two superficial sediment samples were collected at Sta. 341 on 26 August 2009 from Ponar grabs. Using laser diffraction (AWWA Standard No. 2560D; J. D. Ackerman pers. comm.), particle diameters were measured as $d_{90} = 38 \mu m$ and $d_{50} = 10 \mu m$. Previous work reported $d_{50} = 10 \mu m$ in central Lake Erie (Fukuda and Lick 1980; Hawley and Eadie 2007) and categorized sediment type in the region as gray silt-clay material (Thomas et al. 1976) with composition of mostly (>90%) inorganic and ~3% organic carbon (Hawley and Eadie 2007). From the particle diameters analysis, the bed sediment has 0.98% fine sand, 3.55% very fine sand, 75.21% silt, and 20.26% silt.

### Table 1. Details of instrument deployments at study sites: Sta. 341, −17.5 m deep; Sta. 1227, Sta. 1229, Sta. 1231, and Sta. 1233, −11.6 m, −16.6 m, −20.3 m, −19.4 m deep, respectively.

| Station | Instrument | Year  | doy   | Sampling frequency | Sampling interval | No. of samples in each Interval | Depth (m) × −1 |
|---------|------------|-------|-------|-------------------|------------------|--------------------------------|----------------|
| Sta. 341 | TR-1060 temperature loggers | 2008 | 121–212 | 0.1 Hz | N/A | N/A | [1,2,3,4,5,6,7,8,9,10,11,12,12.5,13,13.5,14,14.5,15.25] |
|         |            | 2008 | 212–288 | 0.1 Hz | N/A | N/A | [1,2,3,4,5,6,7,8,9,10,11,12,12.5,13,13.5,14,14.5,15.25] |
|         |            | 2009 | 121–288 | 0.1 Hz | N/A | N/A | [1,2,3,4,5,6,7,8,9,10,11,12,12.5,13,13.5,14,14.5,15.25] |
| 600 kHz ADCP(upward looking) | 2008 | 121–288 | 1 Hz | 15 min | 1* | 1 to 15 (1 m bins) |
|         | 2009 | 121–288 | 1 Hz | 15 min | 1* | 1 to 15 (1 m bins) |
| 2 MHz HR-ADCP (downward looking) | 2008 | 212–233 | 1 Hz | 15 min | 256 | 1.78 mab (3 cm bins) |
|         | 2009 | 121–196 | 1 Hz | 15 min | 256 | 1.69 mab (3 cm bins) |
| ADV | 2008 | 212–288 | 16 Hz | 20 min | 4800 | 16.5 |
|         | 2009 | 121–288 | 16 Hz | 20 min | 4800 | 16.5 |
| XR-620 (Turbidity, temperature, pressure) | 2008 | 121–212 | 1 Hz | 3 min | 1 | 16 |
|         | 212–288 | 2 min | 16 |
|         | 2009 | 121–288 | 3 min | 1 | 16 |
| XR-420 (Turbidity, fluorescence) | 2009 | 121–196 | 1 Hz | 3 min | 1 | 12.5 |
| TWR-2050 (surface waves) water elevation, temperature | 2009 | 121–288 | 2 Hz | 1 h | 1 | 16.5 |
| Sta. 1227 | YSI-6600 (Turbidity) | 2009 | 121–288 | 1 Hz | 1 h | 1 | 10.1 (1.5 mab) |
| Sta. 1229 |            | 196–288 | 1 Hz | 1 h | 1 | 15.6 (1 mab) |
| Sta. 1231 |            | 196–288 | 1 Hz | 1 h | 1 | 19.3 (1 mab) |
| Sta. 1233 |            | 196–288 | 1 Hz | 1 h | 1 | 18.4 (1 mab) |

* Record was averaged over the sampling interval.
clay (Wentworth grain size chart) and the sediment type is silty (Shepherd and Folk’s classification systems). We also measured the bulk and granular densities as \(\rho_b = 1093 \, \text{kg m}^{-3}\) and \(\rho_s = 2150 \, \text{kg m}^{-3}\), respectively (Das 2008).

**Identification of resuspension events**

Increases in the beam amplitude of the ADV backscattering signal (at 1 mab) is used as an indicator of resuspension (hereafter, ADV-amp, unit Count). This is qualitatively and statistically compared, and cross-correlated with turbidity, velocity time series and HR-ADCP acoustic backscatter. Fluorescence time series and satellite images (MERIS, ESA) are used to distinguish inorganic sediment resuspension events from algal bloom-type events in the turbidity time series.

We also relate elevated HR-ADCP acoustic backscatter to resuspension events (Quaresma et al. 2007). We correct HR-ADCP backscatter (ADCP-amp) for attenuation following Lohrmann (2001). The ADV-amp point measurements did not require a correction for attenuation.

**Processing of current velocities**

We calculate the burst-average current speed (from the ADV and HR-ADCP) at 1 mab, \(u_{\text{mean}} = (u_{\text{EW}}^2 + u_{\text{NS}}^2)^{0.5}\) where \(u_{\text{EW}}\) and \(u_{\text{NS}}\) are the mean velocity in the east-west and north-south directions, respectively. Raw ADV and HR-ADCP data were de-spiked (RC Filter method, Goring and Nikora 2002) but no difference was found between de-spiked and raw signals, except for two bursts on day 256.8 and 258.7 in 2008, and two bursts on day 146.2 and 273.8 in 2009, which were removed from the analysis.

In each burst, we define the 1 mab maximum velocity as \(u_{\text{max}}\) (from the 4800 samples in each ADV burst; Table 1), and the 1% of velocities that exceed the 99th percentile as \(u_{99.9}\). We assume a normal distribution where \(u_{99.9} = u_{\text{mean}} + 2.7 \sigma\) (Goring and Nikora 2002; Thomson and Emery 2014), where \(\sigma\) is the standard deviation of the velocity per sampling interval. The calculation of \(u_{99.9}\) enables us to obtain the maximum statistical estimate with 99% confidence in each ADV burst interval and compare it to the actual maximum value (Fig. 5).

**Bottom stress from currents and waves**

The bottom stress can be decomposed into a mean current bottom stress and surface wave-induced bottom stress, \(\tau_{\text{cw}} = \tau_{\text{current}} + \tau_{\text{wave}}\). The mean current bottom stress is \(\tau_{\text{current}} = \rho_o \, u^2\), where \(\rho_o\) is the shear velocity and \(\rho_o = 1000 \, \text{kg m}^{-3}\) is the characteristic water density. In previous studies, we computed \(\tau_{\text{w}}\) and bottom drag coefficient \(C_d\) by least-square fitting the burst averaged HR-ADCP velocity profiles to the law-of-the-wall and found \(C_d = 4.5 \times 10^{-3}\) (method was explained in Valipour 2012; Valipour 2015a); which has been also used in recent Lake Erie field investigations and numerical modellings (Valipour et al. 2015b; 2016). We follow Churchill et al. (2004), and band-pass filter the ADV velocity signals around 3–15 s to obtain the surface-wave induced orbital velocity (the median observed significant period of the surface waves is \(T_s \sim 5 \, \text{s}\)).

**Critical shear stress**

Resuspension of non-cohesive bed material is parameterized according to the Shields parameter \((\theta)\), which is the ratio of the destabilizing forces \((\tau_b)\) to the submerged particle weight (Van Rijn 1993):

\[
\theta = \frac{\tau_b}{(\rho_s - \rho)g}d_s
\]

where \(\rho_s\) is the grain density, with diameter of \(d_s\) (typically \(d_{50}\)) and \(\rho = \rho_o\) is the fluid density. For a Shields parameter greater than its critical value \(\theta_{cr}\), initiation of suspension, is modelled to occur. \(\theta_{cr}\) is a function of \(d_{50}\) and can be obtained from well-known diagrams (e.g., Van Rijn 1993; his Fig. 4.1.4), which can be used to solve Eq. 1 for \(\theta_{cr}\) or \(\tau_{cr}\). For \(~150 \mu m > d_{50} > 40 \mu m\), laboratory-based diagrams (i.e., Bagnold curve, Van Rijn curve or Shields diagram as in Van Rijn 1993, Fig. 4.1.4; hereafter, Shields diagram) give similar values for \(\theta_{cr}\) because the initiation of motion and suspension occur simultaneously. Whereas there are significant differences between initiation of motion and suspension for coarser sediments with \(d_{50} > 150 \mu m\). The existing Shields diagram has not been extended to give \(\theta_{cr}\) for sediments finer than \(d_{50} < 40 \mu m\).

Compared to purely non-cohesive sediments, establishing a \(\tau_{cr}\) threshold for the initiation of suspension of mixed cohesive/non-cohesive material is more challenging because of the additional cohesive forces (Van Rijn 1993; van Ledden et al. 2004; Dalyander et al. 2012). As an alternative for the Shields diagram, \(\tau_{cr}\) for mixed sedimentary materials can be obtained from an empirical relationship with \(\rho_{wcr}\), \(\tau_{cr} = 0.015 \times (\rho_{wcr} - 1000)^{0.73}\) (Mitchener and Torfs 1996).

To obtain \(\tau_{cr} = \rho_b C_d u_{cr}^2\) from our field data, we estimate the observed instantaneous critical velocity \(u_{cr}\) from timeseries plots of \(u_{max}\) and turbidity. This observed \(u_{cr}\) is validated against timeseries of ADV-amp, and contour plots of ADCP-amp. Our thus determined \(\tau_{cr}\) are then compared with those predicted according to the Shields diagram for non-cohesive sediments (e.g., following Dusini et al. 2009; Mier and Garcia 2011), and according to the Mitchener and Torfs (1996) empirical relationship for the mixture of sediments (e.g., following O’Callaghan et al. 2010).

**Watercolumn turbulence parameters**

The turbulent intensity is quantified from the turbulent kinetic energy (TKE) at 1 mab:

\[
TKE = \frac{1}{2T_{int}} \int_{0}^{T_{int}} (u_\text{c}'^2 + u_\text{w}'^2) \, dt
\]

where \(T_{int}\) is the duration of each ADV burst (Table 1), \(u_\text{c}'\) and \(u_\text{w}'\) are the Reynolds decomposed turbulent velocities in the vertical and horizontal directions, respectively. To remove surface wave related fluctuations in calculation of TKE,
we high-pass filter velocity data at the observed significant wave period $T_s$. The dissipation of turbulent kinetic energy $e$ is obtained from the ADV burst data by fitting to the theoretical wavenumber spectrum through the inertial subrange (Lorke and Wüst 2005; Bluteau et al. 2011). Here, we follow standard practice (e.g., Lorke and Wüst 2005) and use the mean velocity in each burst to convert the Reynolds decomposed velocity timeseries to wavenumber space (i.e., Taylor’s hypothesis).

We use ADV burst data to compute Reynolds stresses ($\tau_{\text{Reynolds}}$) at 1 mab as the covariance of $u'_v$ and $u'_h$ during each sampling interval:

$$\tau_{\text{Reynolds}} = \rho_0 u'_h u'_v$$  

We also evaluate the potential for shear instability from the gradient Richardson number:

$$Ri(z) = \frac{N(z)^2}{\left(\frac{\partial u_{\text{mean}}(z)}{\partial z}\right)^2}$$  

where $N(z)^2$ is the square of the Brunt–Väisälä frequency profile:

$$N^2(z) = -\frac{g}{\rho_0} \frac{\partial \rho(z)}{\partial z}$$  

Here, $\partial \rho/\partial z$ is the vertical density gradient calculated from the resampled 15-min temperature data using the UNESCO equation of state (Fofonoff and Millard 1983). Because velocity and temperature field data were recorded at different depths, we linearly interpolate the temperature data onto the 1-m upward looking ADCP bins in the above analysis.

**Results**

**Seasonal stratification**

At Sta. 341, the seasonal temperature stratification formed in early June and the thermocline remained at $\sim -9$ m depth, until deepening in late August intersect the lakebed.
In 2008 and 2009, we also observed upwelling events depending on the wind and stratification conditions, such as doy 205 in 2008 (Fig. 2a,b) or doy 236 in 2009 (Fig. 3a,b). In 2009, the thermocline was substantially deeper than in 2008. Persistent isotherm oscillations indicated the presence of near-inertial (Poincaré) waves with period of 16.8 h (Valipour et al. 2015b).

Nearbed high-turbidity events

As the stratification evolved in the watercolumn, there were changes in nearbed turbidity (Figs. 2a–c, 3a–c). We characterized high-turbidity events to be when the nearbed turbidity at Sta. 341 is >10 NTU; consequently, there were four high-turbidity events in 2008 (T8-1 through T8-4, Fig. 2c) and twelve high-turbidity events in 2009 (T9-1 through T9-12, Fig. 3c). We have assumed a value of 10 NTU as a threshold to identify a high-turbidity event which is 3 times more than the average ambient near-bed turbidity (∼3.3 NTU (e.g., doy 120–210 in 2008), and is a reasonable value for nearbed aquatic ecosystem purposes (McCabe and O’Brien 1983; Newcombe and Macdonald 1991; Newcombe 1994; Bilotta and Brazier 2008).

Nearbed high-turbidity events were observed simultaneously at all stations (e.g., T9-10 and T9-11; Figs. 3a, 4); however, there were also events with correlation between two sites (Sta. 341 and Sta. 1231) and not at others (e.g., T9-7; Fig. 4), and events where high-turbidity was only observed at Sta. 341 (e.g., T9-3 to T9-4; Fig. 4).

Correlation between turbidity, velocity, and acoustic backscatter

To quantitatively compare timeseries from the turbidity sensors and backscatter amplitudes (ADV-amp and ADCP-
amp), correlations between data from these three instruments were examined. Timeseries of ADV-amp and ADCP-amp were highly correlated when data from both instruments were available (doy 214–238 in 2008 and doy 135–196 in 2009; Fig. 5a,b). Timeseries of ADV-amp and turbidity were also highly correlated in 2008 (Fig. 5c), and weakly correlated in 2009. This difference is likely due to presence of algal biomass at Sta. 341 suggesting that the turbidity sensor recorded both sediment resuspension and algae, while elevated ADV-amps were only due to sediment resuspension. The turbidity sensor records particulates of varying diameter, including sediment and algal biomass, whereas ADV (and ADCP) backscatter signals scatter off of sediment but not algae (Lohrmann 2001). The correlation between ADV-amp and $u_{\text{mean}}$ is weaker than with $u_{\text{max}}$ (Figs. 5e,f, 6); suggesting that instantaneous burst-type currents are a better indicator of resuspension. This is consistent with recent studies showing that instantaneous burst currents, as opposed to mean currents, are better predictors of resuspensions (Gloor et al. 1994; Mathis et al. 2014; Aghsae and Boegman 2015). In the Great Lakes, surface waves induce orbital velocities with 3–15 s periods (Churchill et al. 2004; Hawley and Eadie 2007) that will be best captured with $u_{\text{max}}$ as the averaging filters out the wave-induced component of velocity. Thus, we conclude that $u_{\text{max}}$ is a better parameter to characterize resuspension.

In many field deployments instantaneous velocity data are not available (e.g., 15 min to 1 h time-averaged ADCP data is more common than instantaneous ADV data). Therefore, it is useful to relate $u_{\text{mean}}$ to $u_{\text{max}}$ using quantile-quantile analysis (Fig. 7a,c; also see Supporting Information 1). This shows that $u_{\text{mean}}$ and $u_{\text{max}}$ are normally correlated with a ratio of $u_{\text{max}} / u_{\text{mean}} \sim 2.5$ for $u_{\text{mean}} < 0.1$ m s$^{-1}$, above which the ratio increased to $u_{\text{max}} / u_{\text{mean}} \sim 10$ (i.e., for $u_{\text{mean}} > 0.1$ m s$^{-1}$, we expect $u_{\text{max}} \sim 10 u_{\text{mean}}$).

**Critical velocity and shear stress**

Individual resuspension events were examined to determine $u_{\text{cr}}$ from the $u_{\text{max}}$ observed to resuspend sediment. For example, in Fig. 6 there was stronger backscatter and elevated turbidity when $u_{\text{max}} > 0.25$ m s$^{-1}$. The duration of suspension in the water column ranged from 0.03 d to 5 d at 1 mab (Table 2; defined as the time until turbidity/backscatter returned to background levels). There were also instances when the HR-ADCP or ADV recorded short duration velocity impulses without an increase in the backscatter signal (e.g., doy 151.7, 152.2 in Fig. 6a–c). These are suggested to occur when the turbulence was not energetic enough to resuspended material to 1 mab. For example, on doy 151.7, 152.2 (Fig. 6a–c), the velocity impulses are consistent with the elevated HR-ADCP backscatter signals below 1 mab.

The probability distribution of $u_{\text{max}}$ against turbidity using normal quantile-quantile plots (Fig. 7b,d; also see Fig. S1.1, S1.2 Supporting Information 1) is used to determine a velocity threshold for sediment resuspension. The strong normal relationship for turbidity changed when $u_{\text{max}} > u_{\text{cr}} = 0.25$ m s$^{-1}$ with an error range of 0.18–0.27 m s$^{-1}$ (see Fig. S1.3 Supporting Information 1). Higher

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**Fig. 4.** Timeseries of near-bed turbidity in 2009 recorded at five different stations: (a) Sta. 341 at 1.5 mab, (b) Sta. 1227 at 1.5 mab, (c) Sta. 1229 at 1 mab, (d) Sta. 1231 at 1 mab, and (e) Sta. 1233 at 1 mab.
velocities will result in random scattering of sediments into the overlying water column and a breakdown of the normal relationship. For velocities stronger than $u_{\text{max}} > u_{\text{cr}} = 0.25 \text{ m s}^{-1}$ (Fig. 7b,d), the turbidity is $> 10$ NTU; which is also in agreement with the only reported in situ observations of 0.08–0.26 m s$^{-1}$ (i.e., $\tau_{\text{cr}} = 0.03–0.3 \text{ N m}^{-2}$) for a nearby site (Hawley and Eadie 2007). We argue that the error range results from differences in the amount of local consolidation and bioturbation between resuspension events (Sanford 1992; Hawley and Eadie 2007).
To calculate the critical bottom stress and Shields parameter, we followed Gloor et al. (1994) and Mathis et al. (2014) and used the instantaneous observed $u_{cr} = 0.25 \text{ m s}^{-1}$ to obtain $\tau_{cr} = \rho_d \zeta_d \cdot \frac{1}{d} \cdot \frac{1}{u_{cr}^2} = 0.28 \text{ N m}^{-2}$ and (from Eq. 1) $\Theta_{cr} = 2.48$. In order to compare these observed values with theoretical ones, we followed Mitchener and Torfs (1996) and obtained $\tau_{cr} = 0.41 \text{ N m}^{-2}$ for $\rho_d = 1093 \text{ kg m}^{-3}$. Note that $d_{so} = 10 \mu\text{m}$, which is the key parameter to calculate $\Theta$ (Eq. 1), is below the minimum particle diameter of the existing Shields diagram (Van Rijn 1993; his Fig. 4.1.4).

**Observations of algae blooms and high-turbidity events**

Our fluorescence measurements showed the occurrence of spring and summer algae blooms, which may be causing high-turbidity events, not caused by sediment resuspension. For example, at Sta. 341 in 2009 (Figs. 3c, 8b) on day 140–147 the elevated turbidity (T9-01) was not associated with a change in the acoustic backscatter signal, while fluorescence timeseries spiked (Fig. 3d). Similar trends (i.e., elevated turbidity but not acoustic backscatter) were observed on day 234–240 (T9-07) in the absence of fluorescence data (Fig. 8b). During these periods, satellite images indicated patches of elevated Chl-$a$ (Fig. 9b,c). Likewise for T9-06, T9-08 and T9-09 (Table 2). Of the high turbidity events, there was no evidence of algae on day 226 in 2009 (T9-05; Fig. 9a) or in 2008 (Fig. 2). Table 2 and Fig. 8b summarize algal peaks associated with the high-turbidity events, showing the phytoplankton contribute to turbidity 12.57 d out of ~ 20 d (63.7%).

Zooplankton have particle sizes comparable to sediment sizes that may appear as spikes in the acoustic backscatter and/or a turbidity signal (Lohrmann 2001; Walsh et al. 2012). Therefore, it is necessary to eliminate zooplankton...
from being spuriously interpreted as resuspension. During the resuspension events (Table 2), spikes in Chl-a (as required for zooplankton grazing; Blukacz et al. 2010) and/or diurnal migration patterns were not observed. High backscatter measurements during resuspension events are not due to zooplankton, because the turbulence generated velocities are stronger than the typical zooplankton swimming speed, preventing the development of coherent patches of zooplankton (Stich and Lampert 1981; Blukacz et al. 2010; Walsh et al. 2012). Thus we conclude the resuspension events listed in Table 2 are all due to bottom sediment resuspension (see also Churchill et al. 2004; Hawley and Eadie 2007).

Observations of resuspension by surface waves and storms

Surface waves were observed to intensify $u_{\text{wave}}$ and the $u_{\text{wave}}$ to values greater than $u_{cr}$ (Fig. 8a,c). Specifically, waves with $H_s \geq 1.5 \text{ m}$ and $T_s \sim 5 \text{ s}$ increased $u_{\text{wave}} > u_{cr} = 0.25 \text{ m s}^{-1}$ and resuspended sediment (Fig. 8, see doy 240 in 2008 as an example).

Despite the strong correlation between wind forcing and storm-driven currents at this location in Lake Erie (Valipour et al. 2015a,b), we did not observe many direct correlations between wind speed and resuspension from storms (Figs. 2c,d, 3c,d, 8c). The sole exception is T9-04 with $u_{\text{mean}} > 0.1 \text{ m s}^{-1}$, which had a significant contribution of mean currents to the bottom stress (Table 2). Overall, surface waves and wind-induced currents cause resuspension during 3.49 d (17.7%) and 0.97 d (4.9%), respectively, out of 20 d (Table 2).

Observations of HFIW breaking and resuspension

Shoaling of progressive nonlinear internal solitary waves has long been associated with sediment resuspension in the coastal ocean (e.g., Klymak and Moum 2003; Quaresma et al. 2007; Boegman and Ivey 2009; Aghsae and Boegman 2015). These waves are commonly observed in lakes (e.g., Finger lakes Farmer 1978; Dorostkar et al. in press), but have not been recorded in the Great Lakes (indeed the present set of moorings was specifically designed to measure resuspension from internal solitary waves). Despite this lack of observation Hawley (2004) provides indirect evidence that near-inertial wave-induced and/or HFIW-induced resuspension may be occurring in large lakes. Similar speculations on near-inertial waves induced resuspension in large lakes were also reported by Austin (2013). Given the lack of observations of internal solitary waves, here we examine the ability of the linear HFIWs regularly observed on the thermocline of Lake Erie (Bouffard et al. 2012), to resuspend sediments. To our knowledge, this process has not heretofore been observed nor investigated in detail in a large lake. An example of these HFIWs can be seen at the trough of the Poincaré waves

Fig. 7. Normal quantile-quantile plots at Sta. 341. (a, c) mean ($u_{\text{mean}}$) and maximum ($u_{\text{max}}$) currents, (b, d) turbidity and $u_{\text{max}}$. The dashed red line indicates the theoretical quantile from a normal distribution. The periods of quantile-quantile analysis are shown in each panel.
between day 226.5 and 229 (Fig. 10), when Chl-a was low (Figs. 8, 9a), and winds, surface waves (Fig. 10a) and mean near-bed currents (Fig. 10g) were very weak.

The flow field associated with these waves shows that after the first wave packet (near doy 227; Fig. 10b,e), the metalimnion expanded and deepened towards the bed. At the same time, we observed periods when the hypolimnion re-stratified and isotherm detachment from the metalimnion region allowed HFIWs to propagate toward the bed. At 3.5 mab, \( Ri < 0.25 \) below the thermocline (Fig. 10c) suggesting the possibility of shear instability and associated billowing (Boegman 2009). During this time, TKE and rate of dissipation were elevated (doy 227.2–227.4 and 228.2–228.6 in 2009, Fig. 10f). High-pass filtered temperature timeseries data (2nd order Butterworth filter, where the HFIWs are 5–45 min as in Bouffard et al. 2012; Fig. 10e) at 1.05 mab showed the elevated dissipation was preceded by packets of HFIWs. Similar temperature fluctuations were recorded at 1 mab, but at lower frequency (Fig. 10f, Table 1), which confirmed again the presence of high-frequency temperature oscillations and revealed weak near-bed stratification. In the presence of HFIWs, the ADV frequently energized at the predominant period of 17 min (Figs. 10f, 11a–e,g). This 17 min are the predominant peak of HFIWs as previously was expected (Bouffard et al. 2012; Valipour et al. 2015).

During and after the passage of the HFIWs packets, the ADV timeseries recorded \( u_{max} > u_{cr} \) (doy 227.3–227.4 and 228.2–228.5 in 2009, Fig. 10g), along with an increase in TKE (Fig. 10f) and resuspension (elevated turbidity and ADV-amp on doy 227.6 and 228.5 in 2009, Fig. 10h); with the HFIWs being strongest during the weakly stratified periods (i.e., waveguide present) and instantaneous velocities/dissipation strongest between wave events, when the non-stratified watercolumn was more amenable to overturns and turbulence production. Resuspension occurred during these turbulence events, after the passage of the wave packets, for a total of 0.87 d out of the 24 d with high near-bed turbidity (4.4% of the total; Table 2).

Other processes observed to cause nearbed turbidity

Of the 16 high-turbidity events (Figs. 2c, 3c, 8), the processes leading to high-turbidity during T8-03 and T9-03 (Table 2) remained unknown because no significant wind, waves, elevated bottom velocities or Chl-a were observed. Because our hydrodynamic data were limited to one field station and local spatial and temporal variability of near-
turbidity has been observed (Fig. 4), we cannot rule out the possibility that the turbidity was advected to Sta. 341. However, it is unlikely that this is movement of resuspended sediment, which we showed to primarily result from spatially and temporally coherent surface wave-induced resuspension patterns (“Nearbed high-turbidity events” section). Satellite images suggest the variability in turbidity results from advection of algae from the south or south-east (Fig. 9c), potentially from the nutrient rich Maumee or Sandusky River plumes. Lateral advection of suspended sediment has been observed in the Great Lakes (e.g., in Lake Erie by Hawley and Eadie 2007; or in Lake Michigan by Cardenas et al. 2005). These two unknown events collectively contributed 9.3% of the 20 d of high-turbidity events (Table 2).

**Discussion**

We have presented high-resolution field observations of surface waves, storm-induced mean currents and enhanced nearbed turbulence induced by HFIWs leading to resuspension. We have also quantified the contribution of these processes in generating near-bed high turbidity events. To assess the generality of these observations, we compare our results to published observations from Lake Erie and elsewhere and develop mechanistic resuspension models.

**Resuspension criteria**

By observation (Figs. 2, 3, 8), we found an instantaneous critical velocity \( u_{cr} = 0.25 \text{ m s}^{-1} \), shear stress \( \tau_{cr} = 0.28 \text{ N m}^{-2} \) and Shields parameter \( \Theta_{cr} = 2.48 \) (in 8–25°C water) for sediment resuspension in central Lake Erie. This is comparable to \( \Theta_{cr} = 2.3 \) proposed by Van Rijn (1993) for similar sediments in 15°C water. Our observed \( \tau_{cr} = 0.28 \text{ N m}^{-2} \) is also within the observed range of 0.03–0.3 N m\(^{-2}\) in central Lake Erie (Hawley and Eadie 2007), and greater than 0.1 N m\(^{-2}\), 0.2 N m\(^{-2}\), and 0.18 N m\(^{-2}\) by Lick et al. (1994), Fukuda and Lick (1980) and Dusini et al. (2009), respectively, for Lake Erie. It is also greater than \( \tau_{cr} = 0.25 \text{ N m}^{-2} \) in Lake St. Clair (Tsai and Lick 1986), and \( \tau_{cr} = 0.13 \text{ N m}^{-2} \) in Lake Michigan (Lou et al. 2000). Our observed \( \tau_{cr} \) is based on in situ observations of resuspension at 1 mab, and includes silty bed material with significant cohesive forces. The main composition of the bed material is silt (~75%) which is non-cohesive. However, van Ledden et al. (2004) and Dalyander et al. (2012) reported that for a percentage of clay > 7.5%, cohesive forces can influence resuspension, which is relevant for our study site with ~20% clay (see “Sediment size,
density and type" section). Cohesiveness increases the required shear stress for resuspension with the same $d_{50}$. The experimental $\tau_{cr} = 0.41$ N m$^{-2}$ from Mitchener and Torfs (1996) is greater than our observed $\tau_{cr} = 0.28$ N m$^{-2}$, which is in agreement with the statement by Mitchener and Torfs (1996) that their laboratory experiments overestimate the critical shear stress for field purposes.

Our observations show instantaneous velocities, and in particular $u_{\text{max}}$, are better predictors (Figs. 6, 8, 10) for resuspension events. Our nearbed mean currents of 0.15–0.2 m s$^{-1}$ give much lower $\tau_{cr} = 0.11$–0.20 N m$^{-2}$ and $\theta_{cr} = 0.99$–1.77; which are comparable to $\tau_{cr} = 0.1$ N m$^{-2}$ obtained following Parker (2004). Our observations are also consistent with observations in Lake Alpnach (Gloor et al. 1994), which show instantaneous currents of 0.07 m s$^{-1}$ resuspended bed material where the mean currents were 0.02 m s$^{-1}$; coastal ocean data (Mathis et al. 2014), which demonstrated the role of instantaneous shear stress in resuspension; and laboratory experiments (Boegman and Ivey 2009; Aghsae and Boegman 2015), which parameterize resuspension beneath internal waves as a function of the maximum vertical velocity.

Further comparisons between our observed $u_{cr}$, $\tau_{cr}$ and $\theta_{cr}$ in Lake Erie with other locations are listed in Table 3, including rivers (e.g., St. Clair River), estuaries (e.g., Westerschelde Estuary), Coastal Oceans (e.g., Portuguese mid-shelf), and laboratory experiments.

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**Fig. 9.** MEdium Resolution Imaging Spectrometer (MERIS) images of Lake Erie showing the spectral shape at 681 nm in 2009 for day of year (a) 226, (b) 236, and (c) 245. The colored pixels in the panels indicate the likelihood of the last known position of the *Microcystis* spp. The chl index is the cyanobacterial algal biomass index with color from red to blue indicating the highest to the lowest concentration (as shown in the colorbar), respectively. Brown, purple, and white represent the land, chl index $\leq 0$, and clouds, respectively. The MERIS imagery was distributed by the NOAA Coast Watch Program and provided by the European Space Agency (see also Experimental Lake Erie Harmful Algal Bloom Bulletin, 2009-007, 03 September 2009, National Ocean Service, Great Lakes Environmental Research Laboratory). Magenta dashed rectangle shows the location of Sta. 341. For details about the chl index and its annual changes in Lake Erie, see Stumpf et al. (2012).
Mechanisms for storm and surface wave induced resuspension

Our observations (Fig. 8) show that surface waves in Lake Erie with significant period $T_s \geq 5$ s and significant height $H_s \geq 1.5$ m are able to resuspend bed material. These waves are transitional/shallow water waves in a depth of $h = 17.5$ m ($h > 0.5L$ is the deep water limit, USACE 2002) and using the dispersion relationship (e.g., Dean and Dalrymple 1984) with wavelength, $L = 35$ m ($h/0.5$) we can estimate the theoretical $T_s = 4.8 \text{ s} < 5$ s, above which surface
Fig. 11. Spectral plots of observed ADV currents and TR-1060 temperatures during doy 226.5 to 229 in 2009 (period shown in Fig. 10). (a) Spectral plots of zero-padded currents in the east-west direction for the timeseries shown in panel (e). Panels (b) and (c) are the same but in the north-south and the vertical directions for the timeseries shown in panels (f) and (g), respectively. (d) Spectral plot of observed temperature at 1.05 mab for the timeseries shown in Fig. 10d. In (a–d), the gray dashed lines at bottom show 90% confidence levels obtained by multiplying the background spectrum by the 90 percentile value of Chi-distribution. Note that High-Frequency Internal Waves cause the peak of 17-min.
wave induced velocities theoretically reach the bed (see Supporting Information 2). The orbital velocities for waves $T_s \sim 5\text{s}$, therefore, always can reach the bottom, increasing bottom stress and TKE, and may resuspend bottom material. In central Lake Erie, waves with $T_s \geq 5\text{s}$ usually have $H_s \geq 1.5\text{ m}$ (NDBC data) and we conclude that the significant wave period for Lake Erie is a critical parameter for resuspension.

Our results are consistent with a numerical investigation of surface waves in Lake Erie (Dusini et al. 2009), which found waves with periods of $T_s \sim 5.9\text{s}$ cause sediment mobilization to a depth of $-17.5\text{ m}$. Our results are also in agreement with Hawley and Eadie (2007), Hick et al. (1994), Churchill et al. (2004), and Beletsky et al. (2003) who find wind speeds of $\sim 10\text{ m s}^{-1}$, $\sim 10-20\text{ m s}^{-1}$ and $\sim 20\text{ m s}^{-1}$ have the potential to resuspend sediment in Lake Erie at $-24\text{ m}$ depth, Lake Superior at $-90\text{ m}$ depth, and Lake Michigan at $-60\text{ m}$ depth, respectively. These suggest that resuspension can also be expected throughout Lake Erie (see Supporting Information 2).

**Mechanisms for HFIWs breaking**

HFIWs, and their associated resuspension during shoaling, are expected but have not been previously observed in a large lake (Hawley 2004). The presently observed HFIWs form during linear instability (i.e., convective-shear instability implied by a subcritical Richardson number) of the thermocline during the passage of Poincaré waves (Figs. 10, 11). The resuspension, by HFIWs, results either directly from turbulence created during the gravitational degeneration of the Kelvin–Helmholtz billows near the thermocline, or by breaking of obliquely propagating high-frequency internal waves on the lakebed. The latter mechanism is similar to the speculated resuspension by direct shoaling of shorter periods waves in Lake Michigan (Hawley 2004). These two mechanisms are investigated below.

The first mechanism is that the instabilities grow and eventually billow; causing overturns and nearby turbulence that resuspend sediments. This occurs at the billowing timescale $T_b \sim 20\Delta U g^{-1}$ (Turner 1973) where $\Delta U$ is the velocity difference over the layer of interest (above and below the $12^\circ\text{C}$ isotherm, which is between $-11\text{ m}$ and $-15\text{ m}$ depth; Fig. 10), $g' = (\rho_w - \rho)g\rho_w^{-1}$ is the reduced gravity, $g = 9.81\text{ m s}^{-2}$ and $\rho_w = 1000\text{ kg m}^{-3}$ is the water density. The wavelength of instabilities will be $\lambda \sim 7\delta$, where the shear layer thickness $\delta \sim 0.3(\Delta U)g^{-1}$. Thus, we calculate $T_b \sim 2\text{ h}$, $\delta = 2.4\text{ m}$ and $\lambda = 16.8\text{ m}$ for $\Delta U = 0.05\text{ m s}^{-1}$ and $g' = 10^{-4}\text{ m s}^{-2}$, which is consistent with observed $T_b \sim 3\text{ h}$

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**Table 3.** Comparison of observations of $u_{cr}$ (critical velocity), $\tau_{cr}$ (critical shear stress) and $\Theta_{cr}$ (critical Shields parameter) in Lake Erie and other locations ($\rho_s$ is the grain density, and $d_{50}$ is the median grain size of sediment).

| Reference | Site of study | $u_{cr}$ (m s$^{-1}$) | $\tau_{cr}$ (N m$^{-2}$) | $\Theta_{cr}$ | Temperature ($^\circ\text{C}$) | $d_{50} \times 10^{-6}$ (m) | $\rho_s$ (kg m$^{-3}$) |
|-----------|--------------|----------------------|------------------------|--------------|----------------------|---------------------------|-----------------------|
| Present study | Lake Erie | 0.25 | 0.28 | 2.48 | 8–25 | 10 | 2150 |
| Hawley and Eadie (2007); 13 observations, central Lake Erie | | 0.08–0.26 | 0.03–0.3 | 0.27–2.66 | 4–20 | 10 | 2150$^f$ |
| Dusini et al. (2009) | | 0.16* | 0.12 | 0.122 | N/A | 60 | 2650 |
| Fukuda and Lick (1980) | | 0.21* | 0.20* | 1.24* | Room temp. (inferred, 20) | 10 | 2650 |
| Lou et al. (2000) | | 0.17* | 0.13 | 0.28 | 2 | 30 | 2650* |
| Quaresma et al. (2007) | | 0.21* | 0.20* | 0.06 | 14 | 214 | 2650 |
| Gloor et al. (1994) | | 0.07 | 0.007 | 0.05 | 5 | 10 | 2650 |
| Tsai and Lick (1986) | | 0.24* | 0.25 | N/A | N/A | N/A | N/A |
| Araujo et al. (2008) | | 0.32* | 0.48 | 3.71–0.74* | ~15* | 8–40 | 2650* |
| Mier and Garcia (2011) | | 0.97* | 4.23 | 87.1–5.2* | ~20 | 3–50 | 2650 |
| Aghsaee and Boegman (2015) | | 0.074* | 27 $\times 10^{-3}$ | 0.28* | ~20 | 107 | 1180 |
| Boegman and Ivey (2009) | | 0.035* | 5.5 $\times 10^{-3}$ | 1.51–1.08* | ~20 | 106–150 | 1003.5 |

* Denotes parameter is indirectly inferred from the other available parameters in the reference or using $C_D = 4.5 \times 10^{-3}$ (Valipour et al. 2015a).

$^f$ Assumed the same properties as in our study site.

N/A denotes data are not available.
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(val the time from the formation of shear-induced high frequency waves until the billow starts growing; Fig. 10e). For the first resuspension event there is a ~ 7 h lag between the elevated $u_{max}$ and the induced resuspension (doy 227.6 Fig. 10h), but there is no such lag during the second resuspension event (doy 228.5) resulting from elevated $u_{max}$ (doy 228.2–228.5). A lag could result from the need for eddies to mix the weak stratification near the bed prior to the first, but not the second resuspension event. The first resuspension event is consistent with the time scale of decay of turbulence associated with a turbulent event over the period $\Delta t = 226.7-227.2$ doy (12 h), $\langle TKE \rangle = 3.2 \times 10^{-4}$ m$^2$ s$^{-2}$ and $\langle \varepsilon \rangle = 2.7 \times 10^{-10}$ m$^2$ s$^{-3}$ where $\langle \rangle$ denotes averaging over $\Delta t$ (Fig. 10f). The ratio of TKE to dissipation $\langle TKE \rangle / \langle \varepsilon \rangle < 1.2$; $\Delta t > 23$ (1.2 is for the effect of the buoyancy flux in the TKE budget) suggests that 1/23 ~ 4% of the turbulent energy has been dissipated, and the remnant energy has the potential to reach the bed (after overcoming the weak stratification) and resuspend bed material. The second resuspension event on doy 228 reached the bed with zero lag between the gravitational degeneration of overturns, created in the hypolimnion (Fig. 10b,e), and caused resuspension on doy 228.5 because there was no stratification preventing the downward propagation of eddies ($\Delta t = 227.3-228.6$ doy, $\langle TKE \rangle = 2.0 \times 10^{-3}$ m$^2$ s$^{-2}$ and $\langle \varepsilon \rangle = 2.1 \times 10^{-10}$ m$^2$ s$^{-3}$ and the ratio $\langle TKE \rangle / \langle \varepsilon \rangle < 1.2$ $\Delta t > 74$; the last TKE is dissipated below 1 mab).

The alternative mechanism driving HFIWs induced resuspension is that the shear instabilities generate linear waves that propagate through the weakly stratified hypolimnion and break on the bed (Ivey et al. 2000); thus resuspending sediments (Fig. 10f). The waves will radiate toward the bed at a fixed angle $\beta$ to the horizontal with a period ($\omega = 2\pi T_{HFIW}$) according to $N$ where $\omega = N \sin \beta$ (Ivey et al. 2000; Boegman 2009), which in our case $T_{HFIW} = 17$ min (Fig. 11a–d). In the hypolimnion (below thermocline, Fig. 10b) we obtained $N^2 = 0.5-1 \times 10^{-3}$ s$^{-2}$ (Valipour 2012; Valipour et al. 2015a). On doy 226.9 and 228, we estimate $\beta = 11^\circ-16^\circ$ for $T_{HFIW} = 17$ min and $N^2 = 0.5-1 \times 10^{-3}$ s$^{-2}$. Alternatively, we can estimate $\beta$ by multiplying the time vector by a background 0.3 m s$^{-1}$ near-inertial wave speed (Valipour et al. 2015b) to transform to a spatial coordinate under the frozen turbulence hypothesis. We then measured an observed angle $\beta = 15^\circ$ for the 12°C isotherm (Fig. 10b). The radiating linear waves will shoal on the lakebed and potentially resuspend sediment (Fig. 10b,h). The bed slope at Sta. 341 $z = 0.1%$ (Fig. 1) and so we do not expect critical breaking ($\alpha = \beta$) with associated near-bed turbulence (Ivey and Nokes 1989; Maclntyre et al. 1999; Hawley 2004). However, for our case $(\sin \beta)/(\sin \alpha) \gg 1$ and so near-bed mixing from linear wave reflection can be more intense, with larger overturn scales, than the critical case (Ivey et al. 2000). Both HFIW resuspension mechanisms are theoretically possible, and may be simultaneously occurring, strongly suggesting that HFIW are driving occasional resuspension events in central Lake Erie. Similar results may be expected in other aquatic systems.

Extended Shields diagram

Our particle diameter ($d_{50} = 10 \mu m$) is not within the usable range of the existing Shields diagram. To extend the curve, we plot our observed $\Theta_{cr} = 2.48$ against particle number, $D_r = (d_{50} / (S-1))g^2 / \rho$ where $S = \rho_c / \rho$ (Van Rijn 1993), along with published data (Fig. 12a; Table 3), on the Shields diagram for non-cohesive resuspension (Van Rijn 1993). With the exception of the Gloor et al. (1994) observations in Lake Alpnach, our results are consistent with published data for $d_{50} \sim 10–200 \times 10^{-6}$ m showing an extended trend (Fig. 12a), outside the already well defined range for $D_r > 1$ (Van Rijn 1993). This suggested $\Theta_{cr} = 0.2$ is the threshold for bed mobilization (not resuspension) in the laminar flow regime $Re_c < 1$ (where $Re_c = u_c d_{50} / v$ is the Reynolds number) such as Lake Erie with $Re_c = 0.2$ (Van Rijn 1993). The data propose an inverse relationship between $\Theta_{cr}$ and $d_{50}$ for $D_r < 1$ (Fig. 12a). Mier and Garcia (2011) presented a similar diagram showing that for cohesive fine sediments, $\Theta_{cr}$ significantly increases as $d_{50}$ decreases. The data include unconsolidated fine sediments (the samples were not consolidated as in the field) and non-cohesive Plolite particles with $d_{50} \approx 100 \mu m$ (Boegman and Ivey 2009). The lack of consolidation and presence of cohesive material between the non-cohesive material may explain the proposed relationship in Fig. 12a (e.g., Araújo et al. 2008; Mier and Garcia 2011), because $\tau_{cr}$ required to lift the sediments has to be sufficiently strong (increasing $\Theta_{cr}$) to overcome the cohesive forces. The Shields diagram shows $\tau_{cr}$ is inversely proportional to water temperature for $d_{50} < 600 \times 10^{-6}$ m (Fig. 12b). Again, an extended trend is proposed between the data by Van Rijn (1993) and our observed and collected published data, except for Gloor et al. (1994) and Mier and Garcia (2011). The inconsistency with the Gloor et al. (1994) and Mier and Garcia (2011) $d_{50} \sim 50 \times 10^{-6}$ m data was not unexpected, as it is also inconsistent with Fig. 12a, suggesting different resuspension dynamics (e.g., a laminarizing pressure gradient; Aghsae and Boegman 2015) or significantly different sediments. Our extended Shields diagram captures the Mier and Garcia (2011) $\tau_{cr}$ for $d_{50} \sim 3 \times 10^{-6}$ m, which is well aligned with the other observational trends (Lou et al. 2000; Boegman and Ivey 2009; Aghsae and Boegman 2015) in the Shields diagram (Fig. 12a). We did not have sufficient data to further investigate the Mier and Garcia (2011) observations and so we cannot conclude if the $\tau_{cr}$ is valid for $d_{50} < 10 \times 10^{-6}$ m (Fig. 12b).

Conclusions

We presented and analyzed a unique set of high-resolution physical and biogeochemical data in Lake Erie during two ice-free seasons to understand the physical process driving nearby sediment resuspension. Over the 334 d
of measurement, spanning both years, ~20 cumulative days were observed with significant near-bed turbidity (>10 NTU). Algae, surface waves, storm-induced currents and turbulence generated by HFIWs reflection accounted for 63.7%, 17.7%, 4.9%, and 4.4% of this time, respectively. Therefore, the majority of near-bed turbidity in central Lake Erie, is suspended algae. We present the first evidence of resuspension by HFIWs in a large lake. Further laboratory or non-hydrostatic numerical investigations on HFIW-sediment interaction are recommended to clarify the details of the resuspension dynamics. Our results also suggest an extended Shields diagram for silty bed materials, which can be used to predict the resuspension processes in other aquatic systems.

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Fig. 12. (a) Critical Shields parameter ($\Theta_c$) for initiation of suspension of silt (with ~20% cohesive sediment) on an extended Shields diagram. Particle parameter $D_*=d_{50} (5-1) \rho g \gamma/\rho_s^2$ where $S=q_s/q_s^2$, against $\Theta_c$, both dimensionless, values of $\Theta_c$ on lines for $D_*>1$ are from Van Rijn (1993) Shields diagram of non-cohesive material. (b) Critical shear stress ($\tau_c$) for erosion of silt (with ~20% cohesive sediment) on an extended critical shear stress diagram, values of $\tau_c$, on lines for $d_{50}>100 \mu m$ are from the Van Rijn (1993) diagram of non-cohesive materials. In (a, b) the dashed magenta line is a guide from the end of the Van Rijn (1993) data to our results. See Table 3 for details.
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Acknowledgments
We thank Environment and Climate Change Canada, Ministry of Natural Resources Wheatly Fisheries Station, and the captain and crews of the Keenosay and Limnos for logistical support with the field measurements. We are grateful to R. P. Stumpf at NOAA for providing the Satellite images, J. D. Ackerman for providing the sediment grain size analysis results, and C. Bluteau for providing the inertial fitting routines. RV thanks N. Howley at NOAA for his helpful conversations. We also thank two anonymous reviewers for their constructive comments. The research was funded by a NSERC Strategic Projects Grant to J. D. Ackerman, L. Boegman, K. G. Lamb (PI) and R. E. H. Smith, Environment and Climate Change Canada, and Queen’s University.

Conflict of Interest
None declared.

Submitted 26 April 2016
Revised 19 September 2016
Accepted 25 October 2016

Associate editor: Francisco Rueda