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Near-inertial wave driven dissolved oxygen transfer through the thermocline of a large lake

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ABSTRACT

The offshore regions of large temperate lakes are characterized by strong summer temperature stratification that limits vertical mass flux, and enables near-inertial internal wave motions. Here, we investigate the contribution of near-inertial baroclinic velocity shear on enhancing the vertical transport of dissolved oxygen (DO) through the thermocline of the central basin of Lake Erie. The lake is prone to severe annual hypoxia in the hypolimnion and also has strong near-inertial Poincaré wave activity. Using field measurements, analytical arguments and a numerical model under idealized conditions, we show that the near-inertial waves drive baroclinic shear instabilities that enhance the vertical turbulent diffusivity and reduce the rate of DO depletion in the hypolimnion by up to 12% over the entire basin. Results from modeling large-spatial variability in the enhanced thermocline flux to match the distribution of near-inertial wave energy density, indicate that the observed oxygen budgets will vary as a function of location of sampling.

Introduction

Lakes at mid-latitude typically exhibit strong density stratification during summer leading to the creation of thermoclines which act as a physical barrier between the warm epilimnion and the cold deep hypolimnion (e.g., Boegman, 2009). Consequently, the vertical mixing (i.e., vertical diffusivity) of nutrients and dissolved oxygen (DO) through the thermocline is low and may be negligible into the lake interior for the majority of the stratified period (Wüest and Lorke, 2003). An important consequence is that an isolated hypolimnion can become hypoxic (DO < 2 mg/L) over the summer (Charlton, 1980). It is, however, well established that the vertical diffusivity through the thermocline frequently exceeds molecular scales (Saggio and Imberger, 2001). The mechanism most responsible for increased mixing above molecular scales is the velocity shear associated with baroclinic motions (Boegman, 2009; Wüest and Lorke, 2003), which result from basin-scale internal waves. These internal waves are a common response to wind forcing in both small (Boegman, 2009) and large stratified lakes (Austin, 2013; Bouffard and Boegman, 2012). Unlike small lakes, the earth's rotation affects baroclinic motions in large lakes, defined by a Burger number, $S = c/Lf < 1$, where L is a typical length scale (e.g., half of the lake width), f is the Coriolis force and $c = \sqrt{g' \frac{h_1 h_2}{h_1 + h_2}}$ is the phase speed of the internal wave (Merian's formula), with h_1 and h_2 the thickness of the epilimnion and the hypolimnion, respectively, $g' = g\frac{\rho_2 - \rho_1}{\rho_2}$ is the reduced gravity due to the difference in density in the epilimnion (ρ_1) and the hypolimnion (ρ_2), and g is the acceleration due to gravity.

There are typically two basin-scale internal waves in response to the wind forcing: (1) the longitudinal internal seiche, which becomes a coastally trapped cyclonic Kelvin wave (i.e., rotating counterclockwise in the Northern Hemisphere) with a period much longer than the inertial period, $T_i = 2\pi/f$; and (2) the transverse internal seiche, which becomes an anticyclonic Poincaré wave (i.e., rotating clockwise in the Northern Hemisphere) with a maximum period limited by T_i (Bouffard and Boegman, 2012). In the case of very large lake, $S \ll 1$, Kelvin waves are observed only in the nearshore region (within an internal Rossby Radius, $R = c/f$, of locations where the thermocline intersects the lake-bed), whereas Poincaré waves span the entire basin (Rao and Schwab, 2007), but their amplitudes decrease strongly near the coast (Antenucci and Imberger, 2001). Instability in Poincaré waves increases mixing locally, and can be modeled according to the local gradient Richardson number,

$$Ri_g = \frac{N^2}{\left(\frac{\partial u}{\partial z}\right)^2 + \left(\frac{\partial v}{\partial z}\right)^2} \qquad (1)$$

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where N^2 is the Brunt–Väisälä frequency, u, v are the longitudinal and transverse horizontal velocities, and z is the depth (Bouffard et al., 2012). Ri_g is associated with the seasonal stratification through N^2, and Poincaré wave induced baroclinic currents, and if $Ri_g > 1/4$ everywhere in the flow, the flow will be stable (e.g., Miles–Howard condition). Low Ri_g (e.g., <0.25; Troy and Koseff, 2005) will cause the velocity shear to overcome the stratification and allow growth of shear instabilities that degenerate into turbulence and increase the rate of vertical mixing. The magnitude and spatial extent of this enhanced vertical turbulent diffusivity on DO transfer into the hypolimnion of stratified lakes remains largely unknown.

Lake Erie is an excellent system in which to examine the relationship between hydrodynamics and mass transfer of DO through the thermocline for a number of reasons. It is a large lake (i.e., $S \sim 0.05$) where vast offshore regions of the lake are dominated by Poincaré waves during the summer (Bouffard et al., 2012; Boyce and Chiocchio, 1987; Rao et al., 2008). Moreover, Lake Erie experiences severe hypoxic conditions in the hypolimnion in most years (Loewen et al., 2007; Mortimer, 1987; Zhou et al., 2013), so determining the magnitude and spatial extent of DO transfer through the thermocline is relevant both ecologically and economically. The purpose of this study is, therefore, to estimate the spatial and temporal variations in the transfer of DO through the thermocline caused by Poincaré wave-driven turbulence in central Lake Erie. This work extends Bouffard et al. (2013) where the physical processes contributing to oxygen depletion were analyzed by means of field measurements. We first describe the instruments used in the field and the post-processing tools for the data analysis. Then we present the computed daily vertical transfer of DO through the thermocline associated with the Poincaré wave activity at two field stations. Finally, we apply a numerical model to extend our result to the entire central basin.

Methods

Field observations

Lake Erie is the shallowest of the five Laurentian Great Lakes. The lake can be divided into three basins: a shallow western basin with an average depth of ~10 m; a deep eastern basin with an average depth of ~50 m; and a ~20-m deep central basin with a length and width of 200 km and 50 km, respectively. Due to the relatively shallow 20 m depth in the central basin, the thermocline overlies the lake sediments causing central Lake Erie to be more prone to hypoxia than the other Great Lakes.

We consider data collected from two moorings in 2008: one located near the middle of the central basin, station 84 (Sta. 84: 41° 54′ 58″ N, 81° 38′ 32″ W, or UTM zone 17 T: Northing: 4 640 661, Easting: 446 743), and the other located in the western part of the central basin, station 341 (Sta. 341: 41° 47′ 32″ N, 82° 16′ 33″ W, or UTM zone 17 T: Northing: 4 627 494, Easting: 393 995; Fig. 1). Both stations were equipped with temperature and dissolved oxygen loggers moored through the water column (Table 1). Tidbit temperature loggers (Onset; Bourne, MA) were deployed at Sta. 84 and recorded temperature every 10 min to provide information on the large-scale (low-frequency) motions, whereas TR-1060 temperature loggers (RBR Ltd., Kanata, ON) at Sta. 341 sampled every 10 s (i.e., 0.1 Hz) and therefore provided information on smaller-scale (high-frequency) motions. An upward looking acoustic Doppler current profiler (ADCP; 600 kHz Nortek Aquadopp; Rud, Norway) was also anchored on the bottom at Sta. 341 and recorded velocity over the entire water column at 1-m bin resolution (1 Hz samples averaged over 180 s, burst every 15 min). Dissolved oxygen DO-1050 loggers (RBR Ltd.) were deployed one meter above the bottom at both stations, and sampled at 0.1 Hz. DO data were corrected for biofouling and drift by calibrating clean and fouled loggers in oxygen saturated water both prior and subsequent to deployment, respectively, and through in-situ comparison to DO data from nearby multiparameter loggers (YSI 6600 EDS; YSI Inc) equipped with wipers. More details can be found in Bouffard et al. (2013).

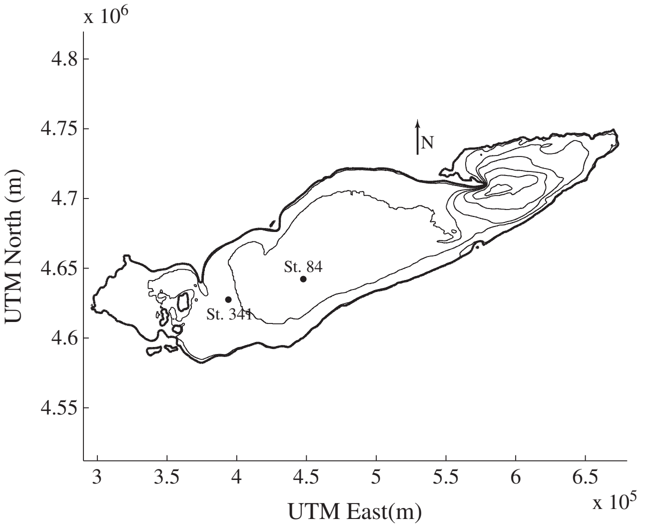

Fig. 1. Map of Lake Erie bathymetry showing depth contours at 10 depth intervals. The two field stations (Sta. 341 and 84) used for this study are indicated.

Temperature microstructure casts were collected using a self contained autonomous microprofiler (SCAMP; Precision Measurement Engineering; Vista, CA), which free-falls through the water column at a velocity of 10 cm s^{-1} and samples water temperature at 100 Hz, thus enabling the recording of temperature microstructure fluctuations as small as 1 mm with an accuracy of 0.001 °C (details provided in Bouffard et al., 2012).

Vertical flux of DO

Our goal is to parameterize the flux of oxygen in the thermocline attributable to the Poincaré wave activity. To do so, we simplify the water column to a two layer system separated by a thermocline of thickness Δz, and express the contribution of the Poincaré wave activity to Ri (Eq. (1)) as

$$Ri_{PW} = \frac{\dfrac{g\Delta\rho}{\rho_0 \Delta z}}{\left(\frac{\Delta u_{PW}}{\Delta z}\right)^2 + \left(\frac{\Delta v_{PW}}{\Delta z}\right)^2} \tag{2}$$

where ρ_0 is the background water density, $\Delta u_{PW} = u_2 - u_1$ and $\Delta v_{PW} = v_2 - v_1$ are the east–west and north–south difference in respective velocity component of the Poincaré wave induced current in the upper (subscript 1) and lower (subscript 2) layers. Mass conservation in the baroclinic flow requires that $u_1 h_1 = -u_2 h_2 = Q_u$, and respectively $v_1 h_1 = -v_2 h_2 = Q_v$ where Q_u and Q_v are the two components of the baroclinic transport, and Eq. (2) can be rewritten as a function of the wave speed c and the equivalent depth h_e,

$$Ri_{PW} = \frac{c^2 h_e \Delta z}{Q_{tot}^2} \tag{3}$$

where the transport (Q_{tot}) is given by $Q_{tot}^2 = Q_u^2 + Q_v^2$.

There are a number of parameterizations of the vertical diffusivity, K_z, as a function of the Richardson number, Ri_g (Lozovatsky et al., 2006; Rao et al., 2008; Yeates, 2003), most of which are of the form,

$$K_{z\_g} = K_o \left(Ri_c^{-1} Ri_g\right)^{-b} \tag{4}$$