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Numerical modeling of the autumnal thermal bar

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A B S T R A C T

The autumnal riverine thermal bar of Kamloops Lake has been simulated using atmospheric data from December 1, 2015, to January 4, 2016. The nonhydrostatic 2.5D mathematical model developed takes into account the diurnal variability of the heat fluxes and wind on the lake surface. The average values for shortwave and longwave radiation and latent and sensible heat fluxes were 19.7 W/m^2, − 95.9 W/m^2, − 11.8 W/m^2, and − 32.0 W/m^2 respectively. Analysis of the wind regime data showed prevailing easterly winds and maximum speed of 11 m/s on the 8th and 19th days. Numerical experiments with different boundary conditions at the lake surface were conducted to evaluate effects of variable heat flux and wind stress. The results of modeling demonstrated that the variable heat flux affects the process of thermal bar evolution, especially during the lengthy night cooling. However, the wind had the greatest impact on the behavior of the autumnal thermal bar: The easterly winds contributed to an earlier appearance of the thermal bar, but the strong winds generating the intensive circulations (the velocity of the upper lake flow increased to 6 cm/s) may destroy the thermal bar front.

1. Introduction

In temperate lakes, a unique phenomenon of nature—a sinking of maximum density waters in a narrow zone—is observed twice a year, in spring and autumn. This effect was discovered by Forel (1880) in Lake Geneva and called *the thermal bar*. In the 1950s, Tikhomirov (1959) began the first systematic study of a thermal bar, in Lake Ladoga (Russia). Thermal bars occur during the spring heating and autumn cooling of a lake. When bodies of water with different temperatures mix, the maximally dense waters sink, forming the frontal water separation. Under the influence of the Coriolis force, which generates parallel horizontal flows (Holland and Kay, 2003), the thermal bar acts as a barrier between thermoactive (the inshore side of the thermal bar) and thermoinert (the offshore side of the thermal bar) regions (Tikhomirov, 1982). The water temperature in the thermal bar front is nearly 4 °C (maximum density temperature). Thermal bars form at shallow areas of a lake and move to the central part, and they are destroyed by subsequent surface heating in spring or cooling in autumn (Chubarenko and Demchenko, 2008; Naumenko et al., 2012). Shimaraev et al. (1993) have hypothesized that due to processes of vertical circulations that are caused by the thermal bar, there is deep-water renewal. At the same time, the thermal bar prevents horizontal water exchange (Moll et al., 1993), which influences the water quality (Blokhina and Pokazeev, 2015; Tsydenov et al., 2015), the vital activity of the plankton population (Likhoshway et al., 1996; Holland et al., 2003), and the lake ecosystem as a whole (Goldman et al., 1996; Moll and Brahce, 1986).

The primary reasons for the appearance of the thermal bar are the changes in the lake water temperature from meteorological forcing or river runoff, or a combination of these (Holland and Kay, 2003; Moll et al., 1993). Meteorological factors, in turn, include the effects of wind and the diurnal variability of heat fluxes. Numerical modeling has established that the impact of the diurnal variation of meteorological data on the spring thermal bar increases with distance from the mouth of the river (Tsydenov et al., 2016). It was also found that propagation of the spring thermal bar decelerates at nighttime and that thermal bar reverse motion (towards the shore) is possible at night because of lack of solar radiation (Tsydenov et al., 2016). Studies of fluid circulation in temperate lakes during spring show that the flow structure is dominated by wind (Csanady, 1972; Malm, 1995; Scavia and Bennett, 1980).

Most of the existing numerical studies of the thermal bar circulation are focused on the case of spring. But the behavior of thermal bars in autumn is different from spring events due to weaker temperature gradients (Holland and Kay, 2003; Huang, 1972). Furthermore, unstable atmospheric conditions, especially strong winds, have a significant impact on autumnal thermal bars (Huang, 1972; Ullman et al., 1998). It should be noted, however, that numerical experiments in reservoirs of different depths showed that the higher wind speed, the deeper the reservoir should be for the zone of divergence of water masses to be close to maximum density temperature (Blokhina, 2015).

The objectives of this study are numerical modeling of the riverine thermal bar during the cooling of a lake and an evaluation of the effects

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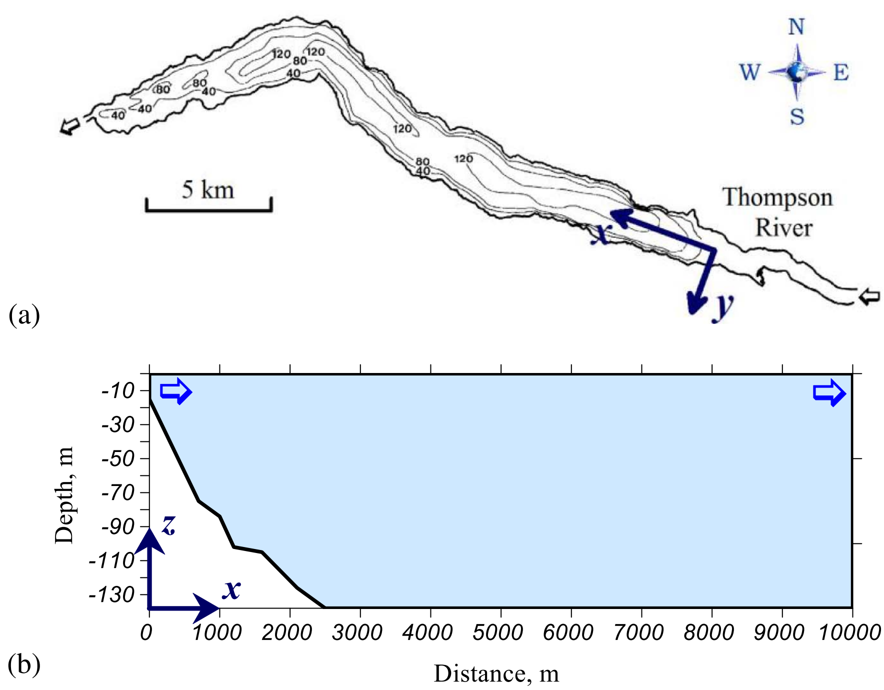

Fig. 1. Kamloops Lake morphometry: (a) bathymetry, (b) calculation domain (longitudinal section).

of wind and surface heat fluxes on the dynamics of the autumnal thermal bar using the 2.5D nonhydrostatic model.

2. Material and methods

2.1. Study area

Because detailed observations of autumnal changes in the temperature distribution of a river-dominated lake are available for Kamloops Lake (Carmack et al., 1979; John et al., 1976), the research area is this lake, located in southwestern Canada, British Columbia, 340 km northeast of Vancouver, between 50°26′ and 50°45′N and 120°03′ and 120°32′W, on the Thompson River. A vertical cross-section of Kamloops Lake along the Thompson River inflow is taken for this study. The origin of the system of coordinates coincides with the river mouth (Fig. 1a). The calculation domain is 10 km long and 138 m deep (Fig. 1b). The depth of the river inflow is 15 m.

2.2. Model description and numerical method

In this study, a 2D section is adopted with the justification that gradients normal to the lake shore are much greater than gradients parallel to the shore, which is neglected. To take Coriolis force related to the Earth's rotation and effect of wind into account, the longshore velocity (v, in the y-direction) was introduced into a mathematical model (as a result, a 2.5D model was obtained). The process to be modeled is assumed to be homogenous along the direction across the river mouth, the *Oy axis*; the *Ox axis* is directed towards the lake, and *Oz* is directed vertically upwards (Fig. 1). The 2.5D nonhydrostatic model for reproducing hydrodynamic processes in a lake expressed in the Boussinesq approximation includes the equations of momentum (Eqs. (1)–(3)), continuity Eq. (4), energy Eq. (5), and salinity balance Eq. (6):

$$\frac{\partial u}{\partial t} + \frac{\partial u^2}{\partial x} + \frac{\partial uw}{\partial z} = -\frac{1}{\rho_0}\frac{\partial p}{\partial x} + \frac{\partial}{\partial x}\left(K_x\frac{\partial u}{\partial x}\right) + \frac{\partial}{\partial z}\left(K_z\frac{\partial u}{\partial z}\right) + 2\Omega_z v - 2\Omega_y w; \quad (1)$$

$$\frac{\partial v}{\partial t} + \frac{\partial uv}{\partial x} + \frac{\partial wv}{\partial z} = \frac{\partial}{\partial x}\left(K_x\frac{\partial v}{\partial x}\right) + \frac{\partial}{\partial z}\left(K_z\frac{\partial v}{\partial z}\right) + 2\Omega_x w - 2\Omega_z u; \quad (2)$$

$$\frac{\partial w}{\partial t} + \frac{\partial uw}{\partial x} + \frac{\partial w^2}{\partial z} = -\frac{1}{\rho_0}\frac{\partial p}{\partial z} + \frac{\partial}{\partial x}\left(K_x\frac{\partial w}{\partial x}\right) + \frac{\partial}{\partial z}\left(K_z\frac{\partial w}{\partial z}\right) - \frac{g\rho}{\rho_0} + 2\Omega_y u - 2\Omega_x v; \quad (3)$$

$$\frac{\partial u}{\partial x} + \frac{\partial w}{\partial z} = 0; \quad (4)$$

$$\frac{\partial T}{\partial t} + \frac{\partial uT}{\partial x} + \frac{\partial wT}{\partial z} = \frac{\partial}{\partial x}\left(D_x\frac{\partial T}{\partial x}\right) + \frac{\partial}{\partial z}\left(D_z\frac{\partial T}{\partial z}\right) + \frac{1}{\rho_0 c_p}\frac{\partial H_{sol}}{\partial z}; \quad (5)$$

$$\frac{\partial S}{\partial t} + \frac{\partial uS}{\partial x} + \frac{\partial wS}{\partial z} = \frac{\partial}{\partial x}\left(D_x\frac{\partial S}{\partial x}\right) + \frac{\partial}{\partial z}\left(D_z\frac{\partial S}{\partial z}\right), \quad (6)$$

where u, v are the horizontal velocity components along the Ox and Oy axes, respectively; w is the vertical velocity component; Ω_x, Ω_y, and Ω_z are the vector components of the Earth's rotation angular velocity; g is the acceleration of gravity; c_p is the specific heat capacity; T is the temperature; S is the salinity; p is the pressure; ρ_0 is the water density at standard atmospheric pressure; and temperature T_L and salinity S_L are a reference temperature and salinity of the lake, respectively. Absorption of shortwave (solar) radiation H_{sol} is calculated according to the Bouguer–Lambert–Beer law:

$$H_{sol} = H_{Ssol,0}(1 - r_s)\exp(-\varepsilon_{abs}\cdot d),$$

where $H_{Ssol,0}$ is the shortwave radiation at the lake surface, $r_s \approx 0.2$ is the water reflectivity coefficient, and $\varepsilon_{abs} \approx 0.3\ \mathrm{m}^{-1}$ is the solar radiation absorption coefficient in water.

A two-parameter k–ω model of turbulence developed by Wilcox (1988) and consisting of equations for kinetic energy and turbulent fluctuation frequency to find vertical eddy diffusivity is used to close the set of Eqs. (1)–(6). The horizontal diffusion coefficients are assumed to be constant as $K_x = D_x = 1.0\ \mathrm{m}^2/\mathrm{s}$.

The equation of state (Chen and Millero, 1986) relates water density to temperature (T), salinity (S), and pressure (p) in the range $0 \le T \le 30$ °C, $0 \le S \le 0.6$ g/kg, $0 \le p \le 180$ bar.

The problem is solved using the finite volume method (Patankar, 1980). The scalar quantities (temperature, salinity, etc.) are calculated