



Effects of spatial wind variability on currents and temperature dynamics of shallow lakes

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ABSTRACT

The wind above lakes can vary spatially due to their large surface and because an internal boundary layer (IBL) may develop, but its effect is often neglected. We aim to find changes in the temperature dynamics caused by the spatially varying wind using observations and a validated 3D lake model. We apply different wind forcings: interpolated with IBL considered, interpolated without IBL, and a vector-averaged uniform distribution to examine the modeled temperature structures. We found that considering the IBL is necessary for the wind forcing to accurately model not only the flow conditions and circulation patterns, but also the temperature dynamics. Depth-averaged and surface temperatures are affected through the turbulent heat fluxes, while the stratification is impacted through the mixing due to the wind variability. The predicted temperature difference along the fetch could be 100% higher when we apply simpler wind models without considering the IBL.

1. Introduction

In many lakes, the wind drives the hydrodynamic transport processes almost exclusively, without significant throughflow or density currents (Torma and Wu, 2019; Xue et al., 2024). Directly, the wind generates surface waves and currents. The waves are responsible for stirring and resuspending different materials from the lakebed, including sediments and nutrients (Jin and Ji, 2005; Li et al., 2017), while currents transport the suspended materials or contaminants horizontally and vertically (Jalil et al., 2017). For example, the spatial distribution and patchiness of zooplankton substantially depend on the current patterns, depending on wind strength and lake size (Sprules et al., 2022). Wind-driven currents and waves both generate mechanical turbulence, leading to further mixing and altering, for example, stratification conditions (Rueda et al., 2009). Temperature dynamics are also influenced by surface heat exchange processes, which are driven by wind-dominated atmospheric turbulence (Fairall et al., 2003). The accuracy of wind forcing as a boundary condition plays a crucial role in lake modeling, due to the above mentioned processes and their complex interactions (Fig. 1).

The wind field can be highly variable spatially in addition to its temporal variation, immensely affecting the lake's hydrodynamics. It is

an important source of uncertainty since wind stress is a surface boundary condition for lake circulation models (Rueda et al., 2009), and circulation is sensitive to both the magnitude and nonuniformity of the forcing. In Lake Michigan, for instance, the wind stress curl arising from the spatial variability of wind plays a predominant role in the seasonal circulation patterns (Schwab and Beletsky, 2003). Podsetchine and Schernewski (1999) showed that the flow patterns drastically change in the case of a spatially varying wind field compared to a uniform wind forcing. Inhomogeneity may alter the number of horizontal gyres and reverse the flow direction along some parts of the shoreline, just where hydrodynamic studies are often focused. Similar findings were reported in the model-based sensitivity analysis of Morales-Marín et al. (2017). Furthermore, the discrepancy between measured and computed vertical flow profiles can be significant in some cases due to small-scale wind inhomogeneities (Rubbert and Köngeter, 2005). The non-uniformity of the wind field can arise from several sources, at different scales, from which some processes are prevailing and frequent, while others are strongly site-specific. These features arise from the orography (Valerio et al., 2017), the shoreline canopy through sheltering (Markfort et al., 2010), or thermally driven breeze events (Wang and Zhuo, 2022). In short, we can conclude that the common characteristic inhomogeneities

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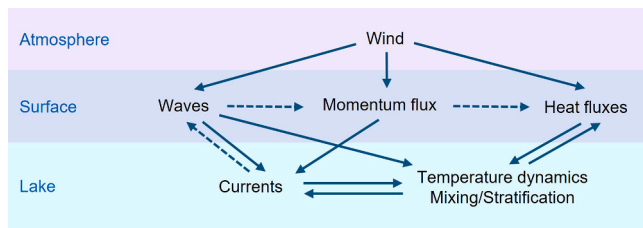


Fig. 1. Effects of the wind to lake temperature dynamics through different processes. Arrows show the direction of the influence; dashed lines show the processes that are usually not included in lake models.

of the lake's wind field, besides site features, must be explored and considered in lake modeling.

Several methods exist to realistically determine the spatial variability of the wind field for lake modeling, employing wind measurement stations with different statistical and interpolation techniques, atmospheric models, or empirical corrections. These methods vary highly in their complexity. One simple representation of the wind forcing acting on the water surface can be achieved by horizontally vector averaging the wind data measured at different locations; however, the wind field is still uniform in this case (Laval et al., 2003). Depending on the lake's size, shape, and features, several interpolation techniques were employed and tested, such as linear, natural neighbor, Barnes, or Kriging interpolation (Beletsky et al., 2003; Juntunen et al., 2019; Rueda et al., 2009; Stepek and Wijnant, 2011; Valerio et al., 2012). However, spatial interpolation schemes may not be accurate enough if, for example, the lake's shape is complex due to a convoluted shoreline (Juntunen et al., 2019). In contrast, atmospheric numerical models can describe many governing processes in detail, such as the impact of canopy sheltering, complex orography (Ashrafi et al., 2022; Santo et al., 2017; Valerio et al., 2012) or large-scale atmospheric processes (Xue et al., 2015). Since typical numerical weather models have limited spatial resolution, computationally costly nested models must be applied to resolve micrometeorological features above small and mid-size lakes (Auger et al., 2021). Nevertheless, these numerical models can be used to validate simplified approaches for long-term lake simulations (Valerio et al., 2017).

The substantial surface roughness reduction is a typical process that modifies the wind as the atmospheric flow arrives from rough land to a smooth water surface. The reduced surface resistance increases the wind velocity, which can be considered with local corrections, namely increasing the land-based wind speeds by a multiplier of the wind-drag coefficient (Chen et al., 2020). However, the wind profile's adaptation to the new surface condition are gradual, leading to a continuously growing internal boundary layer (IBL) along the fetch of the lake (Savelyev and Taylor, 2005; Taylor and Lee, 1984). As a result, the wind stress over the lake surface is not homogeneous, but it increases along the fetch, influencing the flow field in the lake (Józsa, 2014). This land-to-water airflow transition and IBL development phenomena were analyzed by numerical simulations and over-lake wind speed measurements for fetches of the order of a few kilometers by Józsa et al. (2007), who also proposed a semi-empirical formulation to estimate surface roughness transition-related wind speed and stress increment along the fetch together with the IBL height development. This empirical formulation can be applied to determine a wind forcing field, as wind measurements are commonly not available from above the open lake but rather recorded over the land around it.

The goal of this study is twofold. First, we aim to validate a simple analytical wind interpolation technique that includes the effect of the rough-to-smooth surface transition involving IBL development. We employ wind data from several operative anemometers around the shoreline and a short-term field campaign when wind stress was also directly measured by the eddy covariance technique at several points along the fetch. Second, we aim to detect the effect of IBL development-

related spatially varying wind forcing on circulation patterns, thermal structures, and surface heat exchanges in the case of a large and shallow lake. We perform numerical simulations with different wind-forcing scenarios. We simulate circulations and temperatures and evaluate them against profile measurements.

2. Hydrometeorological data

2.1. Study site

Lake Balaton (Hungary) is a shallow but large lake with a mean depth of 3.8 m and a surface area of 596 km². The lake is elongated and can be divided into four main basins regarding circulation and water quality patterns (Istvánovics and Honti, 2018), from west to east, Keszthely, Szigliget, Szemes, and Siófok basins, respectively. The lake's main inflow is the Zala River, which accumulates runoff from half of the total lake catchment (5180 km²) and flows into the westernmost Keszthely Basin (surface area: 39 km², mean depth: 2.9 m). Inflows can be neglected in the case of short-term analyses because they only locally affect currents and temperatures, and the wind and other meteorological variables generate large-scale hydrodynamics and temperature structures. The lake's only outflow is to the Sió Canal, in the eastern Siófok Basin, and it is regulated. The region's prevailing wind direction is the north-northwesterly (NNW), whereas southwesterly (SW) and easterly (E) winds are also frequent. Stronger windstorms predominantly arrive from NNW, which results in about 2–10 km of fetches in the lake.

Open-water measurements have been collected in Lake Balaton for about five months between 18 May and 10 October 2019, focusing on turbulent surface fluxes and energy balance, and between 31 August and 10 October 2018, measuring the spatial variability of the momentum flux across the water surface. Hydrodynamic modeling needs supplementary data, as also shown in this study. All measurement locations are shown in Fig. 2a. The hydrometeorological stations are located in the Keszthely Basin, shown with A, B, and C. Continuous wind and water level data are available at several locations around the lake. By post-processing the time series, we used 20-minute averages of all data variables for our analysis.

2.2. Open-water measurements

In 2019, a hydrometeorological station (denoted as B in Fig. 2a) was operating in the middle of the Keszthely Basin (46° 43' 60" N, 17° 16' 35" E, local depth of 3.5 m). The station's instrumentation is presented in Fig. 2b. Northerly and southerly winds both reach the station with a fetch of 3–4 km, and in the case of easterly winds, the fetch may increase to 20 km. The turbulent momentum, sensible, and latent heat fluxes in the atmospheric surface layer have been measured with the eddy covariance (EC) technique. We used a three-dimensional (3D) anemometer (Campbell CSAT3) and a gas analyzer (Campbell EC150) with synchronized wind, temperature, and humidity measurements at a frequency of 10 Hz. We calculated the fluxes and applied the necessary corrections and filters using the Turbulence Knight 3 software (Mauder and Foken, 2011). The net radiation was measured with a four-component net radiometer (Kipp & Zonen CNR4), including incoming and outgoing short- and longwave radiation. The measurement frequency was 20 s with a 5-minute averaging. The water skin surface temperature was calculated from the outgoing longwave radiation using the Stefan-Boltzmann constant. The water temperature profile along the depth was measured with a thermistor string (Campbell T107) with about 0.5–1 m vertical resolution and a measuring frequency of 5 min. About 2 cm below the water surface, a thermistor was installed on a floating buoy. The sediment heat fluxes at the lake bottom were also measured with heat flux plates (Vaisala Hukseflux). Three T107 thermistors were placed at various depths in the sediment. Data are available in BME (2019).

We measured the waves with a submerged Nortek Signature 1000

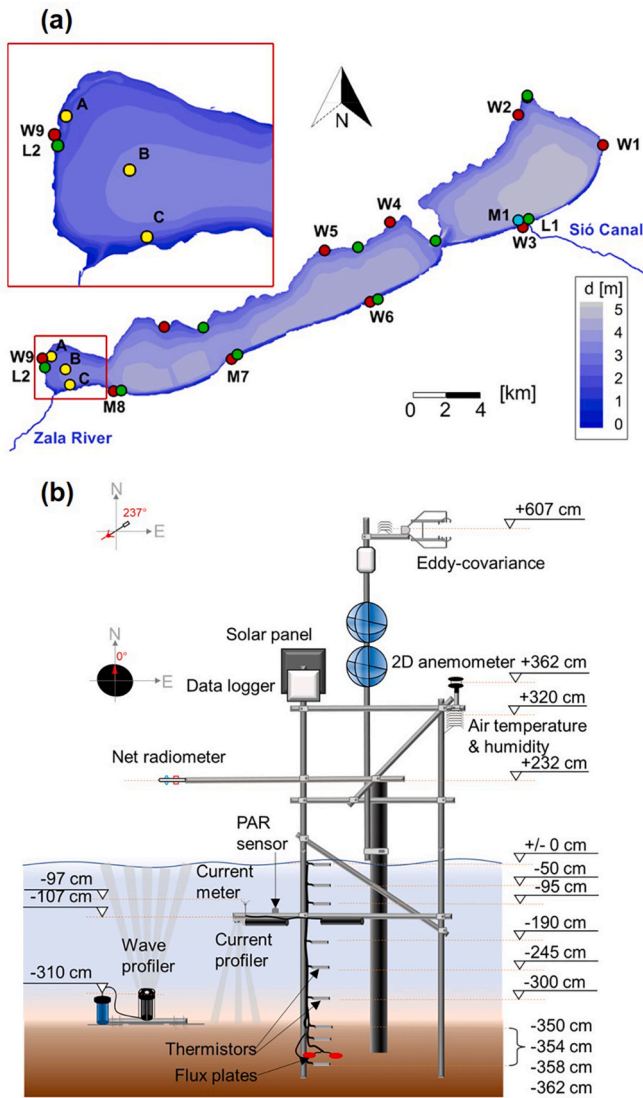


Fig. 2. (a) Lake Balaton bathymetry map with measurement locations, including hydrometeorological stations (A, B, C with yellow) in the Keszthely Basin, wind measurements (W1–W9 with red), and water level gauges (with green). (b) Instrumentation of the open-water station B with measurement heights on the installation day (18 May 2019). (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

current meter using acoustic surface tracking. It sampled the water levels at 4 Hz with 20-minute measurements in bursts. The processing consisted of detrending, spike filtering with linear interpolation, and deriving bulk wave parameters with spectral analysis following [Holthuijsen \(2007\)](#) using 20-minute average calculation. The significant wave height was calculated from the zero-order moment of the wave amplitude spectrum. The zero-crossing method ([Homoródi, 2012](#)) was also used as a control, but there was very little deviation between the results of the two methods. The wavelength and wave period were then calculated from the dispersion relationship.

The current profiles were measured along the depth with a downward-looking Nortek Aquadopp instrument fixed about 1 m below the water surface. The instrument sampled the 3D velocities at 8 Hz with 10-minute bursts between 10 June and 7 August 2019. Profiles were resolved in vertical cells with 8-cm resolution, reaching the bottom of the lake. The data processing of the velocity components at each cell included: (i) filtering requiring instrument-provided correlation values (0–100%) higher than 70%; (ii) spike filtering of less than 2.5 times the

standard deviation from the average using a 15-value moving window; and finally (iii) linear interpolation. Mean values have been determined for the velocity components, from which the horizontal velocity magnitude and directions were calculated along the profile.

The application of the time series measured by the B station was twofold. First, the meteorological variables provided the boundary conditions of the lake model; second, the current and temperature time series formed the core of the lake model's calibration, validation, and evaluation for the different wind forcing scenarios.

2.3. Wind measurements and data

We give more details on the wind observations in this section. The First Micrometeorological research within a Croatian-Hungarian (FIMO-CROHUN) collaboration carried out a measurement campaign between August 31 and October 10, 2018 in Lake Balaton to reveal the characteristics of the wind development over the lake ([Lükő et al., 2020](#)). Three stations have been set up in the Keszthely Basin to explore the spatial variability of wind speed and stress along the fetch for the prevailing wind direction. They are shown on the map as A ($46^{\circ} 45' 14''$ N, $17^{\circ} 14' 59''$ E), B ($46^{\circ} 43' 60''$ N, $17^{\circ} 16' 35''$ E), and C ($46^{\circ} 42' 50''$ N, $17^{\circ} 17' 21''$ E) in [Fig. 2a](#). Stations A, B, and C had fetches of 50 m, 3.5 km, and 6 km in the NNW wind direction, and water depths of 1.4 m, 3.5 m, and 1.5 m, respectively. Station A and B both had Campbell CSAT3 3D sonic anemometers. Station C had a Gill Windmaster 3D sonic anemometer until 26 September 2018, when a heavy storm destroyed it. Since 13 September 2018, simultaneous wave measurements have been done, similar to those at station B in 2019. The onshore Station A in [Fig. 2a](#) belongs to the HUN-REN-BME Water Management Research Group, and it has been monitoring water quality for the past 20 years, including water temperatures along the depth of 1.5 m for the 2019 summer period as well, which was also used for the calibration of the hydrodynamic model.

2.4. Supplemental data

The wind data around the lake was used to determine the wind forcing of the hydrodynamic model. Several anemometers operated by the National Directorate General for Disaster Management are located in ports and on beaches to monitor storms (shown with W1–W9 in red in [Fig. 2a](#)). The measurement heights are between 10 and 21 m, and data included 10-minute averages of wind speed and direction.

The water level data around the lakeshore were used to validate the hydrodynamic model. The Western Transdanubia Water Directorate measures water levels at an interval of 10 min, and the gauges are located inside ports (shown in green in [Fig. 2a](#)).

The Landsat 8 satellite's Level 2 land surface temperature data were collected for the simulation period as well, which has 16-day temporal and 30-m spatial resolution ([Jenkerson, 2020](#)).

3. Methods

3.1. Hydrodynamic model

We used the Finite Volume Community Ocean Model (FVCOM) 4.1 model developed by [Chen et al. \(2003\)](#) primarily for numerical simulations of 3D hydro- and thermodynamic processes in the ocean, but it is regularly used in lake environments ([Anderson et al., 2010](#)). The model solves the hydrostatic Reynolds-averaged Navier-Stokes, temperature transport, and density equations of incompressible fluids using an explicit finite-volume method. The terms resulting from Reynolds averaging are calculated using the Smagorinsky model horizontally, and the Mellor-Yamada 2.5 turbulence model vertically ([Chen et al., 2003](#)). FVCOM solves the equations on a horizontally unstructured triangular mesh. The cell size was set to 50 m near the shore, and it was gradually increased to 100 m in the offshore direction in the Keszthely Basin, while

the computational mesh was less detailed in the other basins of the lake. Vertically, a standard σ -coordinate transformation is used, dividing the water column into eight layers non-uniformly, with higher resolution near the surface and the lakebed. The set of these equations is solved by specifying appropriate boundary conditions, that is, the turbulent momentum and heat exchanges on the surface of the lake, which are calculated by FVCOM using the Coupled Ocean Atmosphere Response Experiment (COARE) algorithm (Fairall et al., 2003). The COARE algorithm calculates the heat fluxes by solving the stability-corrected logarithmic profile equations in a couple of iteration steps based on Monin-Obukhov similarity theory. It uses the Charnock coefficient for the friction velocity and roughness lengths for the sensible and latent heat fluxes. The input data for the FVCOM model are wind speed, incoming long- and shortwave radiation, air temperature, and humidity.

In addition to the specified wind forcing and other meteorological boundary conditions, it is essential to calibrate the model. In Lükő and Torma (2024), we performed a sensitivity analysis on the most important parameters of the model, which were: the roughness length of the lakebed; the background mixing coefficient (K [$\text{m}^2 \text{s}^{-1}$]); the roughness lengths of the air-lake heat fluxes (z_{0H} [m] and z_{0q} [m]); and the light extinction coefficient (k [m]). The roughness length of the bed showed no significant effect on the flow velocities, even when we changed it by three orders of magnitude. At the same time, the model was sensitive to the background mixing coefficient of the turbulence model, and similarly to the roughness lengths of the heat exchanges and the light extinction coefficient. The mixing coefficient is usually taken in the order of $K = 10^{-4} - 10^{-5} \text{ m}^2 \text{s}^{-1}$ (Dong et al., 2020; Durski et al., 2004); however, the velocity magnitudes were sensitive to it, which can be explained by the dissipation of turbulent kinetic energy. In addition, matching the velocity directions and the temperature structure required a similarly low coefficient, so we applied a $K = 10^{-6} \text{ m}^2 \text{s}^{-1}$ coefficient, with which the model can reproduce Lake Balaton's vertical velocity and temperature profiles. The COARE heat flux algorithm is designed for the ocean surface and uses the same roughness lengths for sensible and latent heat. Lükő et al. (2022a) showed that these lengths can differ by a couple of orders of magnitude, and the latent heat flux roughness length has an intra-seasonal variation. The COARE algorithm significantly underestimates the heat exchanges and consequently overestimates the depth-averaged temperatures in the model; thus, we calibrated the roughness lengths using $z_{0H} = 5 \cdot 10^{-3} \text{ m}$ and $z_{0q} = 5 \cdot 10^{-4} \text{ m}$ values for sensible and latent heat fluxes, respectively. We calibrated the light extinction coefficient as $k = 0.5 \text{ m}$.

When evaluating the accuracy of the model, the modeled results are compared with the measured current profiles, as well as the velocity magnitude [m s^{-1}] and direction [$^\circ$], and the temperature profiles with the depth-averaged temperature (T_{avg} [$^\circ \text{C}$]) and the potential energy anomaly (Wiles et al., 2006) characterizing the strength of the stratification (ϕ [J m^{-3}]). The index has been applied to describe the stratification conditions in shallow lakes (Istvánovics et al., 2022; Török and Torma, 2024), and can be calculated as follows:

$$\phi = \frac{1}{d} \int_{-d}^0 (\bar{\rho} - \rho(z)) g z dz \quad (1)$$

where d [m] is the water depth, ρ [kg m^{-3}] is the water density, and $\bar{\rho}$ is its depth average value. The potential energy anomaly shows how much energy per unit volume is needed to mix the stratified water column completely.

Waves and currents interact in several ways, resulting in varying degrees of mutual modification (Niu and Xia, 2017). Waves shape the currents through two processes. On the one hand, if the effect of the waves reaches the bottom, the bed resistance increases through the bed shear stress caused by the waves acting there, and on the other hand, it gives an impulse to the flow through the so-called radiation stresses (Mellor, 2008; Warner et al., 2008). The interactions can mainly modify

the currents, affecting the waves much less, especially in lake environments. To take the interactions into account, we set up a wave model and coupled it one-way to the presented HD model. We used the SWAN (Simulating Waves Nearshore) model for the wave simulation, which is an established 2D model in oceanic coastal environments and lakes (Holthuijsen, 2007; Homoródi, 2012; Kocsis et al., 2022). We calibrated the Lake Balaton SWAN model for open water using wave data measured at the open-water station B in the Keszthely Basin (Lükő and Torma, 2024). The model's input data was the open-water measured wind speed and direction. The results, such as the significant wave height, wave direction, wave period, wavelength, and near-bottom velocity, were transferred to the lake model by one-way coupling. The SWAN calculations were done on the same triangular mesh as the horizontal mesh of the FVCOM hydrodynamic model.

3.2. Wind forcing

The spatial inhomogeneity of the wind field over the lake can be significant. On the one hand, due to the large size and elongated shape of Lake Balaton, both the direction and magnitude of the wind can be very variable. On the other hand, the difference in the aerodynamic resistance of the land and the water surface leads to an internal boundary layer (IBL) in the atmosphere (Józsa et al., 2007) (Fig. 3b). The air flows from a rough to a smooth surface, which increases the wind speed and shear stress along the fetch above the water surface (beyond a limited distance). The effect of the IBL was confirmed by simultaneous measurements on the windward and leeward shores in line with the wind direction (Torma and Krámer, 2017). The wind speed at the windward shore can be twice as high as measured at the leeward shore of Lake Balaton due to the IBL in the prevailing case of NNW winds.

The increased wind speed leads to an increase of wind stress over the lake surface along the fetch. We set up a model to reproduce friction velocity and roughness length over the lake, uniquely using a single wind measurement (Krámer, 2006). In Fig. 3a, we show the wind field calculation steps with equations. All calculations were repeated for each time step. First, we calculate IBL-based wind fields for $i = 1 : N$ number of stations. Using the wind data of a station and the lakeshore geometry, we determine whether wind is from land or water, and calculate the fetch (F) field for the whole lake. Equations 1–5 in Fig. 3a are solved iteratively until the roughness length ($z_{0,w}$) converges, but in different order depending on wind data is from land or water. This IBL model assumes logarithmic wind profiles above land (Equation (1) in Fig. 3a) and within the IBL over the lake (Equation 2 in Fig. 3a). $U_{10,l}$ is the 10-m wind speed for land, $u_{*,l}$ is the friction velocity for land, $k = 0.41$ is the von Kármán constant, $z_{0,l} = 1 \text{ m}$ is the roughness length for land (from sonic radar measurements during the 2018 campaign), $U_{10,w}$ is the 10-m wind speed for water, $u_{*,w}$ is the friction velocity for water, $z_{0,w}$ is the roughness length for water, g is the gravitational acceleration, ν is the kinematic viscosity, and $\alpha = 0.042$ is the Charnock coefficient (Lükő et al., 2022b). The height of the IBL (δ) is calculated using Elliott's (1958) formulation based on the ratio of $z_{0,w}$ and $z_{0,l}$ and on the fetch (Equation 4 in Fig. 3a). The wind speed at the top of IBL over the lake will be equal to the wind speed that is determined for the same height over the land (Equation 5 in Fig. 3a). In other words, this means that above the IBL height, the wind speed profile is determined by the roughness length of the land surface, while in the IBL, it is determined by the roughness length of the water surface according to the Charnock relation (Equation 3 in Fig. 3a). The result of the iteration is the IBL-based wind field for one station. Fig. 3c–e shows different wind fields determined by wind data from three different stations for a simple elliptic lake.

The wind data of N (in our case, 10) stations around the lake were used to calculate the spatially and temporally varying wind forcing fields discretized on the horizontal mesh of the hydrodynamic simulations, incorporating the effect of IBL. All wind fields associated with the stations were interpolated spatially using the inverse-distance weighting

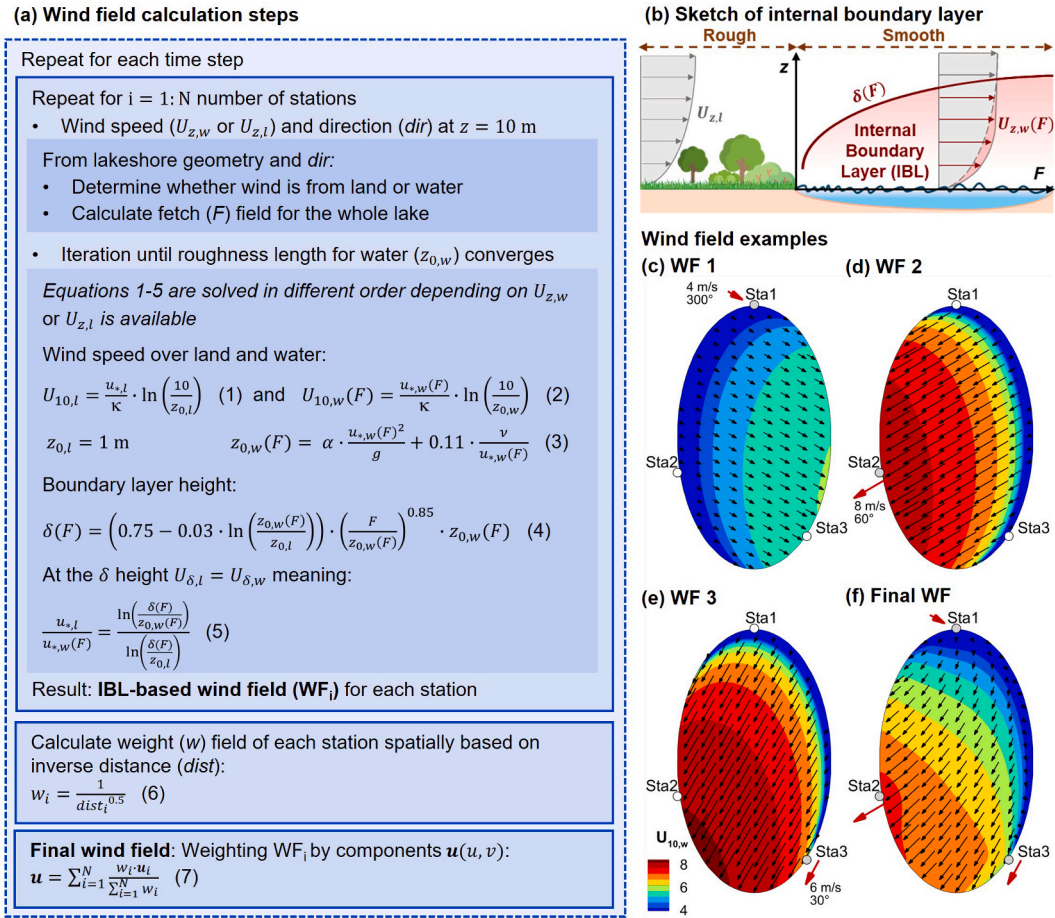


Fig. 3. (a) Wind field calculation steps considering internal boundary layer (IBL) development and inverse-distance interpolation. (b) Sketch of IBL development over the lake. Example IBL-based wind fields using individual stations (c–e) and (f) inverse-distance weighted final wind field.

method. For this, the weight fields were calculated based on the inverse of the distance between a given cell center and a given wind station, using a weighting power of 0.5 (Stepk and Wijnant, 2011) (Equation 6 in Fig. 3a). The final wind field was then obtained by weighing each station's IBL-based wind field (Equation 7 in Fig. 3a). This inverse-distance method was used instead of linear interpolation in the need for extrapolation and due to the wind stations' uneven location and the lake's elongated shape. Further details about the wind interpolation procedure are available in Torma and Krámer (2017). Fig. 3f shows the final interpolated field, where the spatial inhomogeneity of wind speed and direction due to the IBL development and spatial interpolation can be observed.

The wind stress from this semi-analytical IBL model was evaluated against eddy covariance data along fetch from the 2018 measurement campaign as well. The longitudinal profile of the friction velocity along the prevailing NNW wind direction can be compared with the values measured at the three stations, A, B, and C. In Fig. 4a, b, we can see the estimated and measured wind speed and friction velocity for a given wind speed class (e.g., where the speed of a NNW wind is between 9 and 10 m s⁻¹ at station B). Due to the rough-to-smooth transition, the wind speed rises along the fetch from 6 to 10 m s⁻¹. In parallel, we can see that the friction velocity rises from 0.2 to 0.4 m s⁻¹ over the 6 km of fetch. In Fig. 4c, d, scatter plots of the estimated and measured wind speed and friction velocity are shown for all wind speed classes for stations B and C. This simple IBL model could thus reproduce the fetch-dependent variability of the roughness length, and, most importantly, that of the friction velocity associated with the momentum input into the lake. The estimates at the open water station B are promisingly accurate. At the same time, the model was also able to quantitatively reproduce the IBL-

related wind transformation at station C, which is in the shoaling zone where wave-breaking may occur, leading to changes in waves and, consequently, in the apparent water surface roughness. In contrast, friction velocity estimates failed at station A because it is in a transition zone, which is affected by the sheltering of the shoreline canopy. Another possible error source is that the footprint of station A could have been partly on land, as it was so close to the shoreline.

3.3. Wind model scenarios

We chose a representative period to simulate Lake Balaton hydrodynamics in terms of wind speeds and directions and other meteorological variables typical of summer conditions. The time series of the measured meteorological conditions are shown for the modeling period of 23 June 2019 to 18 July 2019 at station B in Fig. 5. During this period, there were several moderate windstorms with the prevailing NNW directions and a few milder events with S winds, and some E winds as well. The air temperature varied between 14 and 33°C, while the relative humidity was between 20 and 90%, with higher relative humidity in the first 5 days. Fig. 5d shows the radiation components; the incoming shortwave radiation peaked over 800 W m⁻² on most days, while the longwave radiation averaged around 400 W m⁻². In the model simulations, besides the wind, all other meteorological variables were spatially homogeneous.

We modeled and compared three different wind forcing scenarios:

- V1-IBL: The instantaneous wind field is calculated using the IBL model and inverse-distance interpolation, as presented in the previous section.

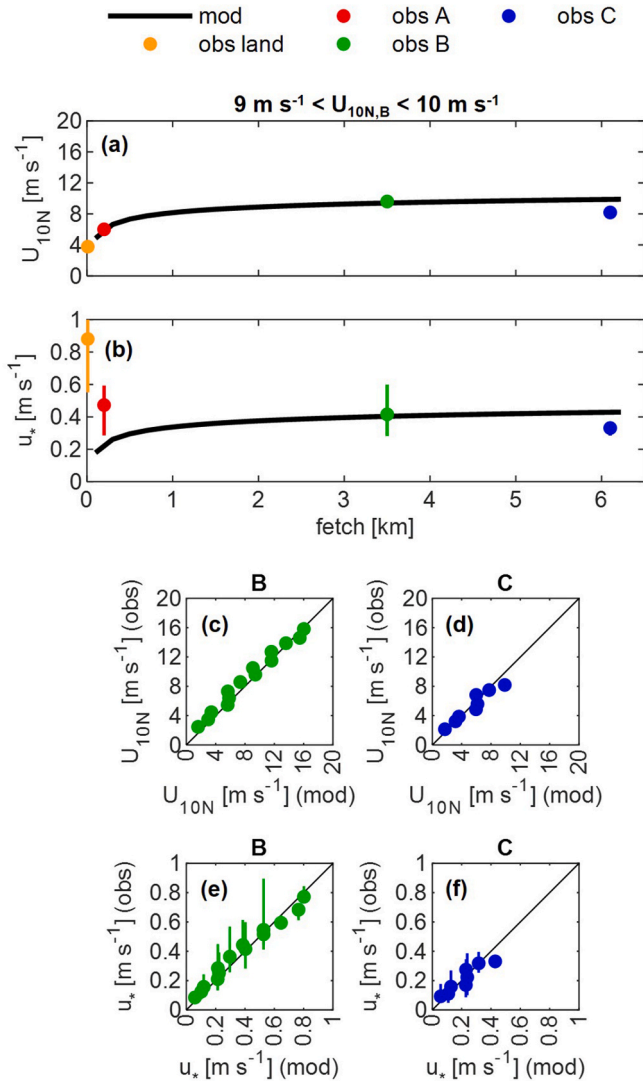


Fig. 4. (a) Longitudinal profiles of wind speed and (b) friction velocity along the fetch of the Keszthely Basin for the $U_{10N,B} = 9\text{--}10 \text{ m s}^{-1}$ wind speed range and NNW wind direction. The black line shows the model prediction, while filled circles show the measurements (wind speed and friction velocity means and roughness length medians) with 25–75th percentile lines over land surface (yellow) and over water surface at stations A (red), B (green), and C (blue). Scatter plot of observed and estimated (c, d) wind speed and (e, f) friction velocity for each 1 m s^{-1} wind speed class at stations B (green) and C (blue). Filled circles show means with 25–75th percentile lines. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

- (ii) V2-INP: The instantaneous wind field is obtained by simple inverse-distance interpolation directly from the station data. We use the same weight fields for stations as for V1-IBL.
- (iii) V3-AVG: The instantaneous wind field is spatially homogeneous and calculated as a vector average of the station data. We average all stations' wind data by components.

The wind field estimation used the wind data measured around the lake at nine stations (W1–W9) and the open-water station B in Fig. 2a. In Fig. 6, we show examples of the different wind fields at a single moment in time (more precisely a 3-hour average of the preceding period). In Fig. 6a, we show V1-IBL, where the spatial inhomogeneity of wind due to the IBL development can be observed. In this case, the wind directions showed little spatial variability, similar to most times when stronger winds occur from NNW. In Fig. 6b, we show V2-INP, where there is also

high spatial variability between stations, compared to V3-AVG, which would be spatially completely homogeneous.

4. Results

4.1. Water levels

We analyzed the sensitivity to the choice of the wind forcing method through water level fluctuations in the three wind model scenarios (V1–V3). We compared the modeled water levels with the measured ones in the eastern basin near Siófok and in the western basin near Keszthely. Generally, all three wind model scenarios resulted in similar water level fluctuations, confirming the similar momentum flux input with the different wind forcings. For the whole simulation period, the water level root mean square deviations (RMSEs) were 0.019, 0.020, and 0.020 m at Keszthely, and 0.014, 0.015, and 0.015 at Siófok for V1, V2, and V3, respectively. There were very low phase errors at both locations. The models slightly underestimated the amplitudes, but we could not achieve a better match with the additional calibration of the Charnock coefficient for the roughness length. Reassuringly, the accepted coefficient was the same as the value derived from the eddy covariance data measured at the open-water station B, which was $\alpha = 0.042$ (Lükő et al., 2022b). We note that using data from one open-water wind station resulted in significant phase errors and opposite fluctuations in many instances (Lükő and Torma, 2024). The larger amplitudes may also come from the fact that measurements are taken inside harbors with possible interference impacts, although we have not explored this any further.

4.2. Currents

When water level fluctuations are not accurate, the circulation patterns cannot be well modeled in the lakes since the velocity directions and magnitudes are influenced by the seiche motion in shallow lakes (Ji and Jin, 2006). Although water levels and seiche were both well simulated in Lake Balaton, the currents were more sensitive to the wind forcing. We evaluated the simulations against vertical velocity profile observations for a qualitative assessment. In Fig. 7, the time series of 20-minute-average velocity magnitudes and directions at 2 m depth at the open-water station B can be seen for the different wind model scenarios together with the observations. The agreement with measurements is generally moderate. Model errors at the beginning of the period are likely caused by the model's spin-up from a cold start. Despite water level fluctuations being similar, open-water currents significantly differ in the wind model versions based on the velocity magnitudes and direction along the vertical profile. Generally, higher velocity magnitudes can be achieved with the V1-IBL model. This resulted in some overestimation at lower depths (around 1 m) for a few periods, but the other two forcings resulted in lower velocities at all depths. This is especially visible during the wind event starting on 14 July, where the IBL-related wind variability can be highlighted during this period. The velocity magnitudes and directions are significantly better reproduced with the V1-IBL wind model. Simulated flow directions match the observations at different depths, while the V2-INP and V3-AVG wind forcings have about 150° error in directions in the lower depths (not shown in a figure). The RMSEs for velocity magnitudes are 0.011, 0.017, and 0.016 m/s for this event in case of V1-IBL, V2-INP, and V3-AVG, respectively, while they are similarly 0.015 m/s for the whole simulation period. The angular RMSEs (the RMSE of the shortest angular difference) for the velocity directions are 49.0 , 77.4 , and 99.6° for this period, in the case of V1-IBL, V2-INP, and V3-AVG, respectively. The angular RMSEs for the whole period decreased from 106.4 to 78.0° when applying the IBL in the wind forcing. The IBL-based wind forcing V1-IBL also resulted in a better match with observations during the periods when the flow direction changed continuously and gradually, for example, between 26 and 28 June, when the flow fully overturned three times. All three

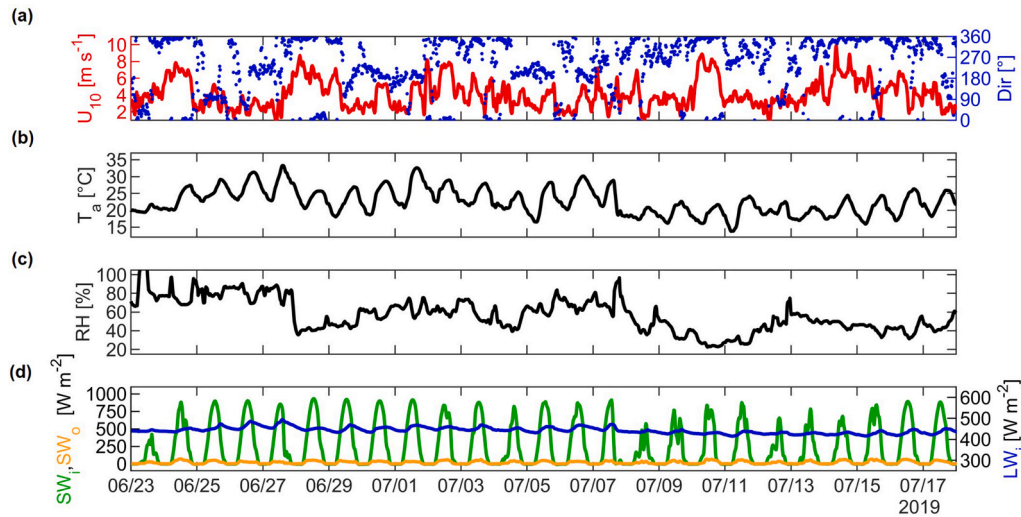


Fig. 5. Time series of measured (a) 10-m wind speed magnitude and direction, (b) air temperature, (c) relative humidity, and (d) incoming and outgoing shortwave and incoming longwave radiation components and at the open-water station B used for the modeling period of 23 June–18 July 2019.

(a) V1-IBL

July 14, 2019 03:00 (UTC+2)

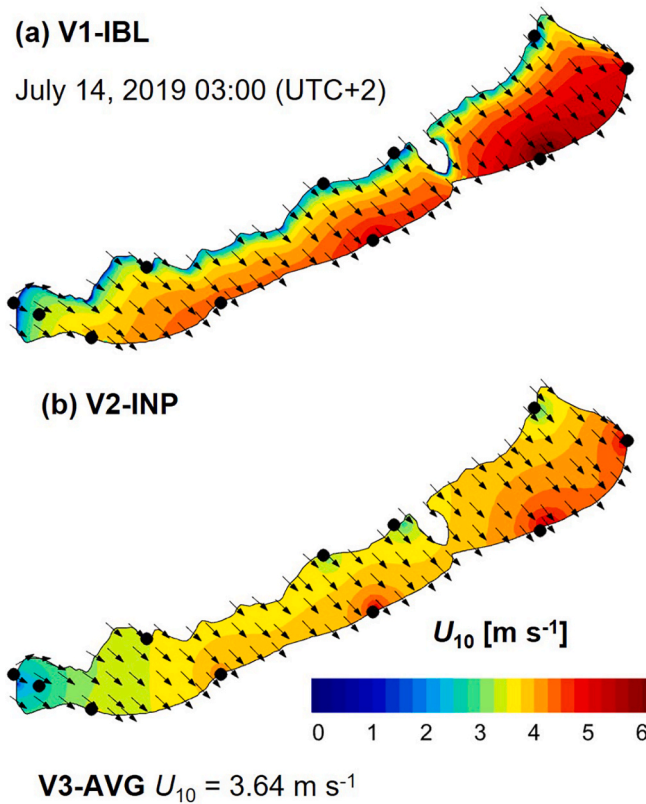


Fig. 6. Calculated 3-hour average wind speed fields with the (a) internal boundary layer development and interpolation (V1-IBL) and (b) only interpolation (V2-INP) methods on 14 July 2019, 03:00 (UTC + 2) (3-hour average) using the data of ten wind measurement stations (black filled circles).

forcing methods reproduced this phenomenon to some extent, but V2-INP and V3-AVG resulted in a high phase error. Finally, during other shorter periods with various wind directions, the flow direction deviations from the observation were reduced with the V1-IBL model (e.g., on 29 June and 2 July). Interestingly, after SSW winds, only on 26 July, V3-AVG agrees better with the observations than the other two models, which could be due to the fact that, in the case of low winds, the IBL does not develop or had not developed for the stronger NNW winds yet.

However, the measurement location in the middle of the basin was not favorable considering it can be the middle of the main circulation gyre, where velocity magnitudes and directions change fast and model errors could be distorted.

4.3. Water temperatures and stratification

We analyzed the temperature dynamics in the lake using depth-averaged temperature and the potential energy anomaly, showing the strength of the thermal stratification of the water column caused by the spatial variability of the wind. In Fig. 8, we show the time series of depth-averaged temperatures and potential energy anomalies for the different model scenarios together with observations at the open-water station B. Fig. 8a shows that the V2-INP and V3-AVG wind model scenarios have trouble with the quick cooling on 28 June 2019. The simulated temperatures became overestimated after that, in contrast to the V1-IBL wind model scenario. Between 29 June and 4 July, a somewhat stronger warming was calculated using the V1-IBL forcing compared to the observations and the other two forcings. Nevertheless, wind forcing of the V1-IBL scenario provided a better match until 11 July, when it resulted in a slight underestimation. Overall, simulated depth-averaged water temperatures do not deviate significantly from the observations for any of the three scenarios, with a slightly better match in the case of V1-IBL. For all the wind model scenarios, higher roughness lengths would have been needed before 9 July 2019, while a lower one would have been needed after that, which agrees with our previous findings of intra-seasonality (Lükő et al., 2022a). Regarding the vertical mixing processes, Fig. 8b shows that V1-IBL reproduces thermal stratification better than the others, especially during the 3-day-long strongly and continuously stratified period between 25 and 28 June. Fortunately, all three scenarios were able to reproduce that stratification did not fully dissolve during the nights. The determination coefficients were $R^2 = 0.78, 0.39$, and 0.41 in the case of the depth-averaged temperatures, and $R^2 = 0.78, 0.64$, and 0.54 in the case of the potential energy anomaly for the V1-IBL, V2-INP, and V3-AVG wind model scenarios, respectively. Based on the R^2 values, the more complex IBL-based wind forcing resulted in a significantly better match with the observation at station B compared to the other two wind forcing scenarios regarding both depth-averaged temperatures and stratification conditions.

The measurement campaigns in the Keszthely Basin allowed the validation of the model locally only, but the spatial variation of the temperature in Lake Balaton is also governed by the wind forcing. Papp

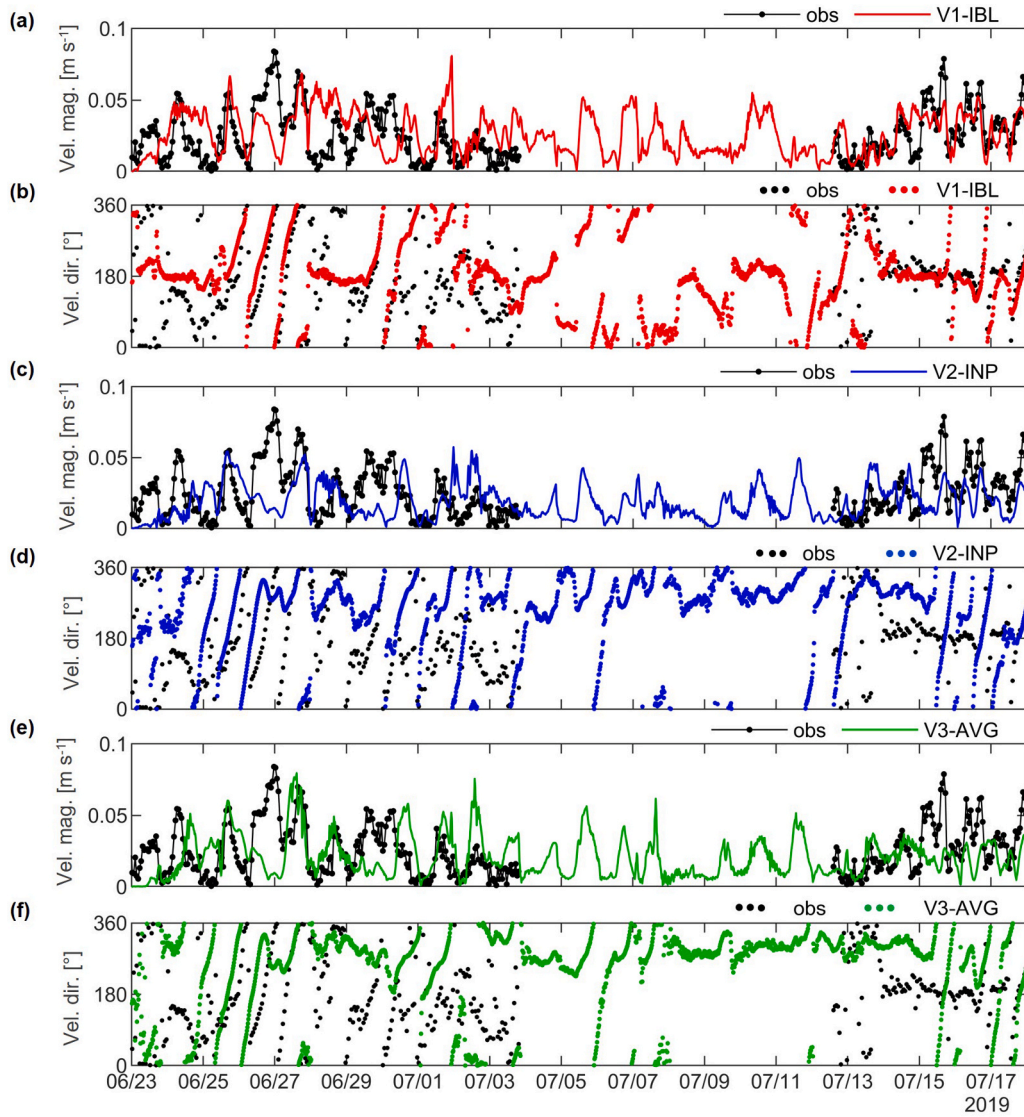


Fig. 7. Time series of (a, c, e) velocity magnitude and (b, d, f) horizontal direction at 2-m depth at the open-water station B, simulated with the (a, b) V1-IBL (red), (c, d) V2-INP (blue), and (e, f) V3-AVG (green) wind models and observations (black). (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

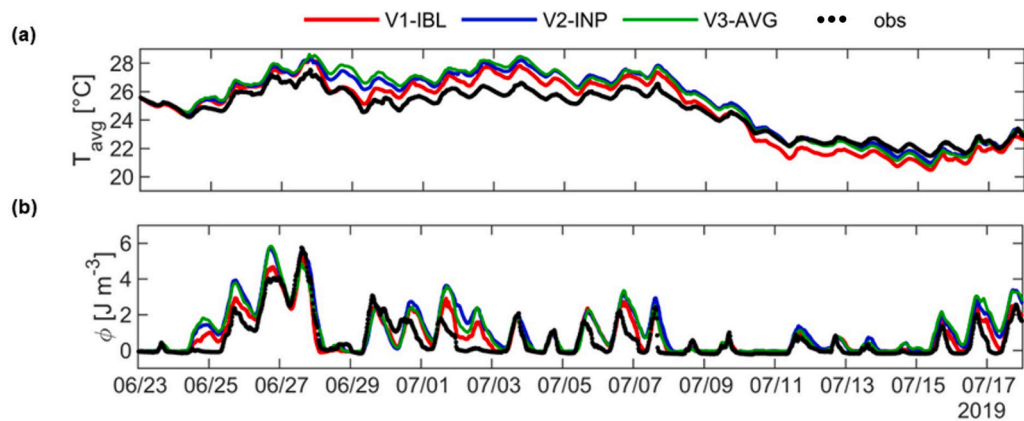


Fig. 8. Time series of (a) depth-averaged temperature and (b) potential energy anomaly at the open-water station B for simulations with the V1-IBL (red), V2-INP (blue), and V3-AVG (green) wind models and observations (black). (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

(2023) analyzed the surface temperature data from Landsat satellite images and evaluated them against the open-water measurements in the

western Keszthely Basin (at station B) and nearshore in the eastern Siófok Basin. The remotely sensed surface temperature product has a 30-m resolution, but the satellite's orbital period is 16 days, and cloud-covered images are also unusable; thus, only one image was useful for the modeled period. Fig. 9 shows the surface temperature variation

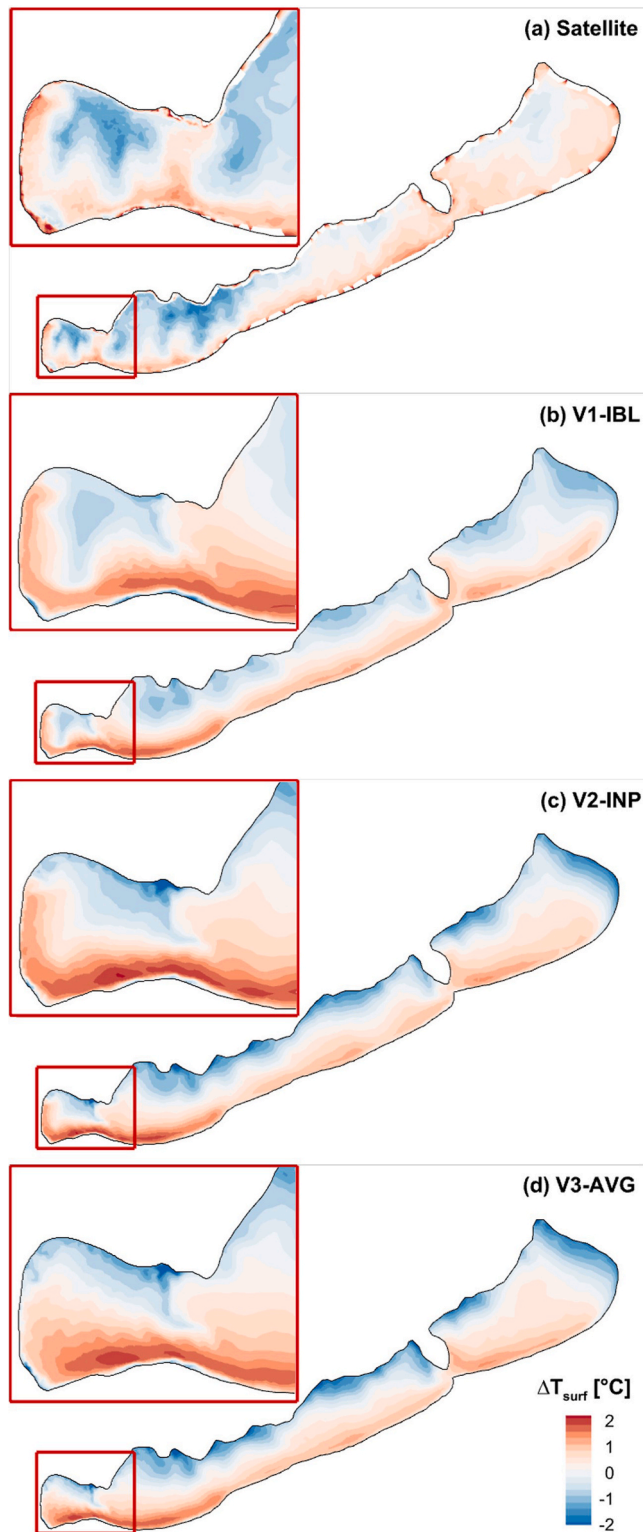


Fig. 9. Lake Balaton surface temperature differences on 28 June 2019, 11:30 (UTC + 2) with (a) the Landsat satellite land surface temperature observation (Papp, 2023) and simulations with the (b) V1-IBL, (c) V2-INP, and (d) V3-AVG wind models.

(surface temperature minus the lake-wide mean surface temperature) on 28 June 2019 at 11:30 (UTC + 2) from the satellite observation and the different model simulations. For the whole lake, the average RMSEs for surface temperature were 1.15, 1.40, and 1.40 for the V1-IBL, V2-INP, and V3-AVG simulations, respectively, which means the IBL-based wind outperforms the others again. We find this a good agreement, considering the compared temperatures are not fully equivalent: the satellite observed the skin temperature, while the model results show the surface layer representing a couple of 10 cm depth, depending on the location and depth. There were a couple of days of low wind and stronger thermal stratification before the moderate wind event on 28 June 2019, when the lake started to mix vertically. The general spatial variation of temperature is captured well by all the model scenarios, which show that the northern and southern littoral zones are colder and warmer, respectively. Overall, it can be stated that the temperature differences between the northern and southern nearshore zones were overestimated with the V2-INP and V3-AVG forcings. For these two scenarios, the northern regions were simulated as colder than the lake-average temperature. According to the thermal image, in the Keszthely Basin, warmer temperatures were present in the western and southern parts of the basin and at the border of the Szigliget Basin from the east. Colder water can be seen in the northwestern part, penetrating the middle open-water area by current-induced plumes. This behavior was reproduced significantly better by the IBL-based scenario, confirming that horizontal transport processes are better simulated. At the same time, in the eastern Siófok basin, the surface temperature was relatively homogeneous according to the satellite image. V2-INP and V3-AVG scenarios simulated a large thermal gradient between the shorelines. In this sense, the V1-IBL forcing performed better, providing fewer horizontal temperature differences. Further images would be necessary for a quantitative assessment of surface temperature distribution results; however, the satellite data from this single time point gave further confirmation of the predictive accuracy of the IBL-based forcing model. In the future, different satellite products could be included in the analysis. Unfortunately, satellites with shorter return periods usually have lower spatial resolution might not be suitable to explore the spatial variability of the temperature field in detail.

5. Discussion

5.1. Horizontal circulation patterns

In Fig. 10, we show 24-hour time-averaged horizontal circulation patterns for the different wind model scenarios on 2 July and 10 July 2019. On those days, there were moderate NNW winds. The circulation patterns are very different depending on the applied wind forcing. On 2 July, in the case of V1-IBL, there is one large counterclockwise circulation in the Keszthely Basin, while V2-INP resulted in two main gyres and V3-AVG in one clockwise circulation opposite to that in V1-IBL. On 10 July, two larger circulations can be seen in the basin at the northeast and southwest parts of the basin for V2-INP and V3-AVG. The circulation patterns were very similar for these two forcings. In contrast, in the case of V1-IBL, one elongated gyre developed at the western shoreline and another one in the middle of the basin. This visual inspection concludes that the flow field is more sensitive to the wind forcings' spatial variability than the seiche motion-related oscillations.

5.2. Spatial temperature dynamics

Since currents and turbulent mixing are mild during calm periods, the primary driving factor of stratification and spatial variation of temperature is the water depth in shallow lakes. As a result, the shallower nearshore regions can warm up more by solar radiation during the day. This can be easily tracked in satellite-based surface temperature fields. In the case of Lake Balaton, it was shown that the nearshore regions are warmer than the open-water areas over monthly and yearly

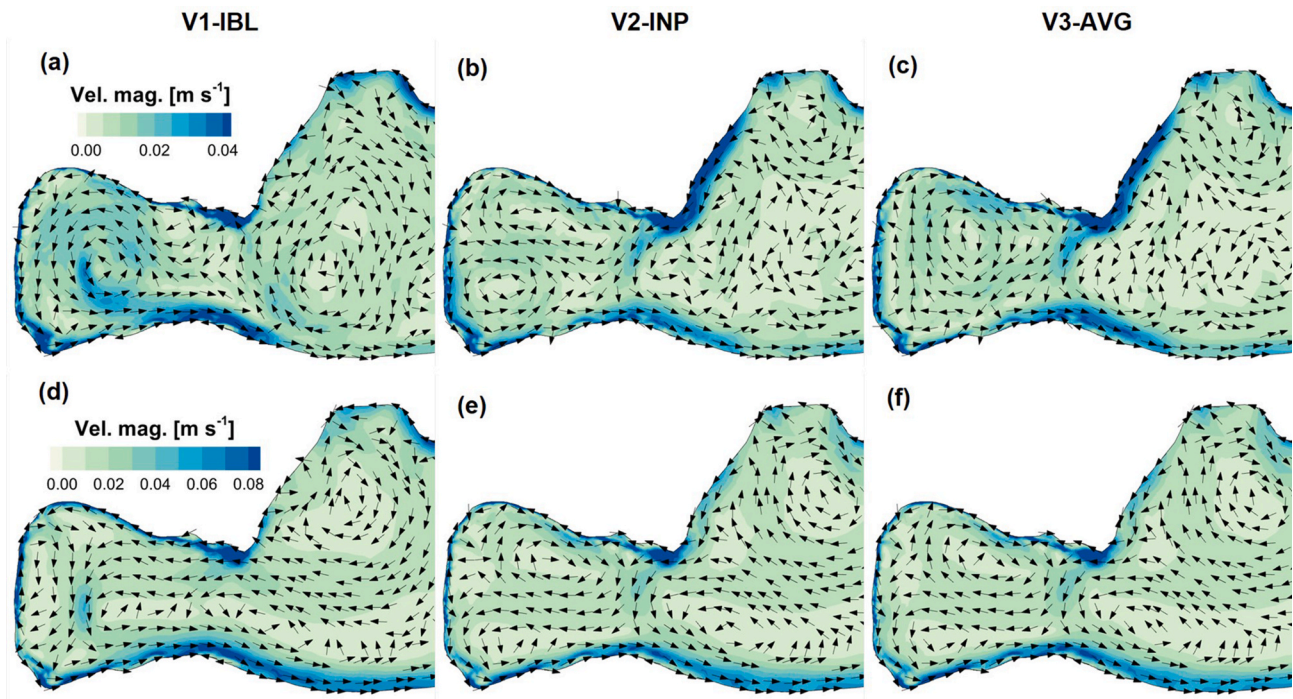


Fig. 10. 24-hour time-averaged velocity fields on (a–c) 2 July 2019, and (d–f), 10 July 2019 in the Keszthely and part of Szigliget basins of Lake Balaton with the (a, d) V1-IBL, (b, e) V2-INP, and (c, f) V3-AVG model simulations.

long-term timescales (Li et al., 2024). However, the spatial temperature and stratification differences are still impacted by the surface heat fluxes and consequently by the wind over the surface (Fig. 1).

Besides the circulation and the water temperature field, we also calculated the potential energy anomalies over the whole lake to show the spatial variation of the thermal stratification. In Fig. 11, the time-averaged potential energy anomaly fields on 26 June 2019 and between 29 June and 1 July 2019 can be seen with the different model

scenarios for the Keszthely and Szigliget basins. While results with V2-INP and V3-AVG show similar patterns, the IBL development caused significant differences. In the case of the first event (Fig. 11a–c), a stronger stratification (up to $\phi \sim 5 \text{ J m}^{-3}$) could develop in the deeper part of the Szigliget Basin compared to V1-IBL. Even though the IBL-affected spatial wind variance should be low during low winds, it also resulted in a slightly weaker stratification in the open water of the Keszthely Basin. In Fig. 11d–f, during the second event, the stratification

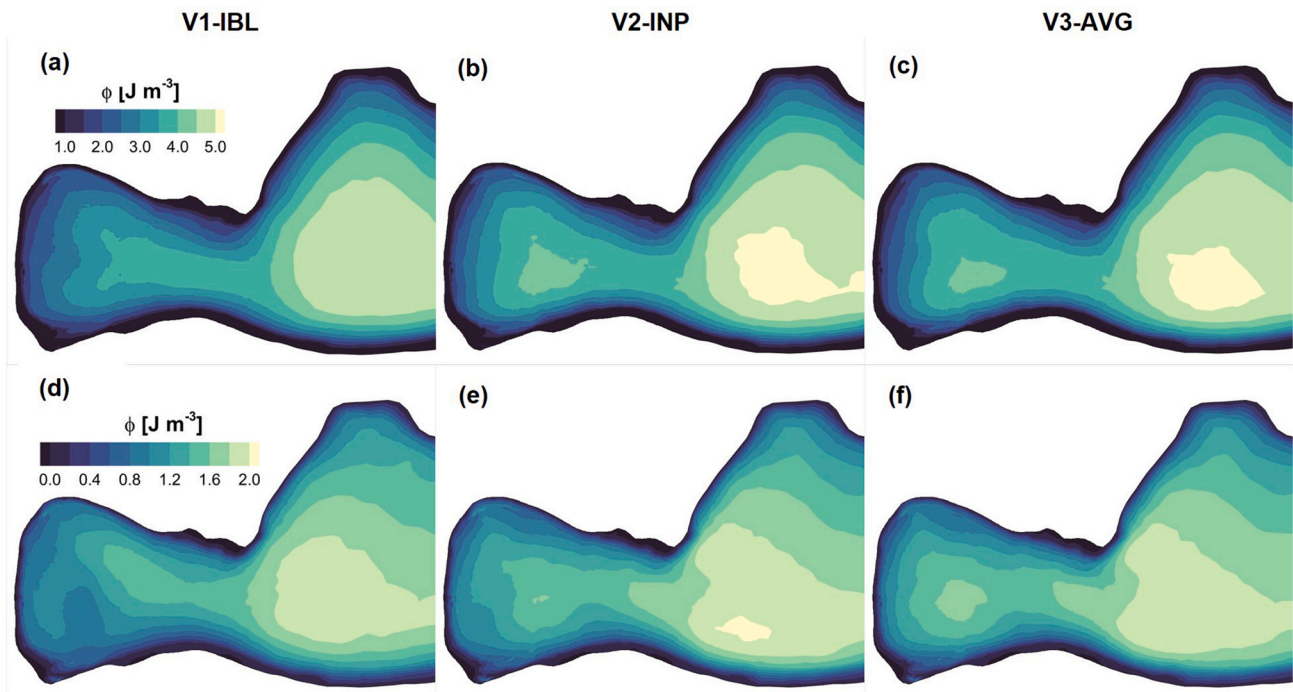


Fig. 11. Time-averaged potential energy anomaly fields (a–c) on 26 June 2019, and (d–f) between 29 June and 1 July 2019 in the Keszthely and part of Szigliget basins of Lake Balaton with the (a, d) V1-IBL, (b, e) V2-INP, and (c, f) V3-AVG model simulations.

was weaker because of low and moderate winds from the southwest. The significant difference is that in the case of V1-IBL, more than 50% lower ϕ can be seen in the open water compared to the other two scenarios in the Keszthely Basin. Furthermore, the horizontal circulation has a higher impact on the spatial distribution of ϕ for V1-IBL, while in the other two cases, ϕ more closely follows the bathymetry. During this period, the average wind was somewhat higher; consequently, the IBL development was more effective.

5.3. Turbulent heat fluxes

The wind speed directly affects the turbulent heat fluxes across the water surface and, consequently, the water temperature field, while the

latter also has feedback on the fluxes. Thus, the spatial inhomogeneity of fluxes is governed by the complex interaction of atmospheric and lake conditions. Fig. 12a-f shows the modeled time-averaged sensible and latent heat fluxes for a stronger northerly wind on 28 June 2019. In Fig. 12a-c, the sensible heat flux shows smaller differences, but results with V1-IBL have slightly higher exchanges near the south shore. This can be explained by the fact that sensible heat flux is more strongly correlated with temperature differences, but there is also a strong correlation with wind, even on larger timescales (Lükő et al., 2022a). In Fig. 12d-f, the latent heat flux could be up to 100 W m^{-2} higher when IBL is included in the wind model. This aligns with the regression analysis results, namely that latent heat flux is primarily governed by the wind rather than the specific humidity gradient on shorter time scales.

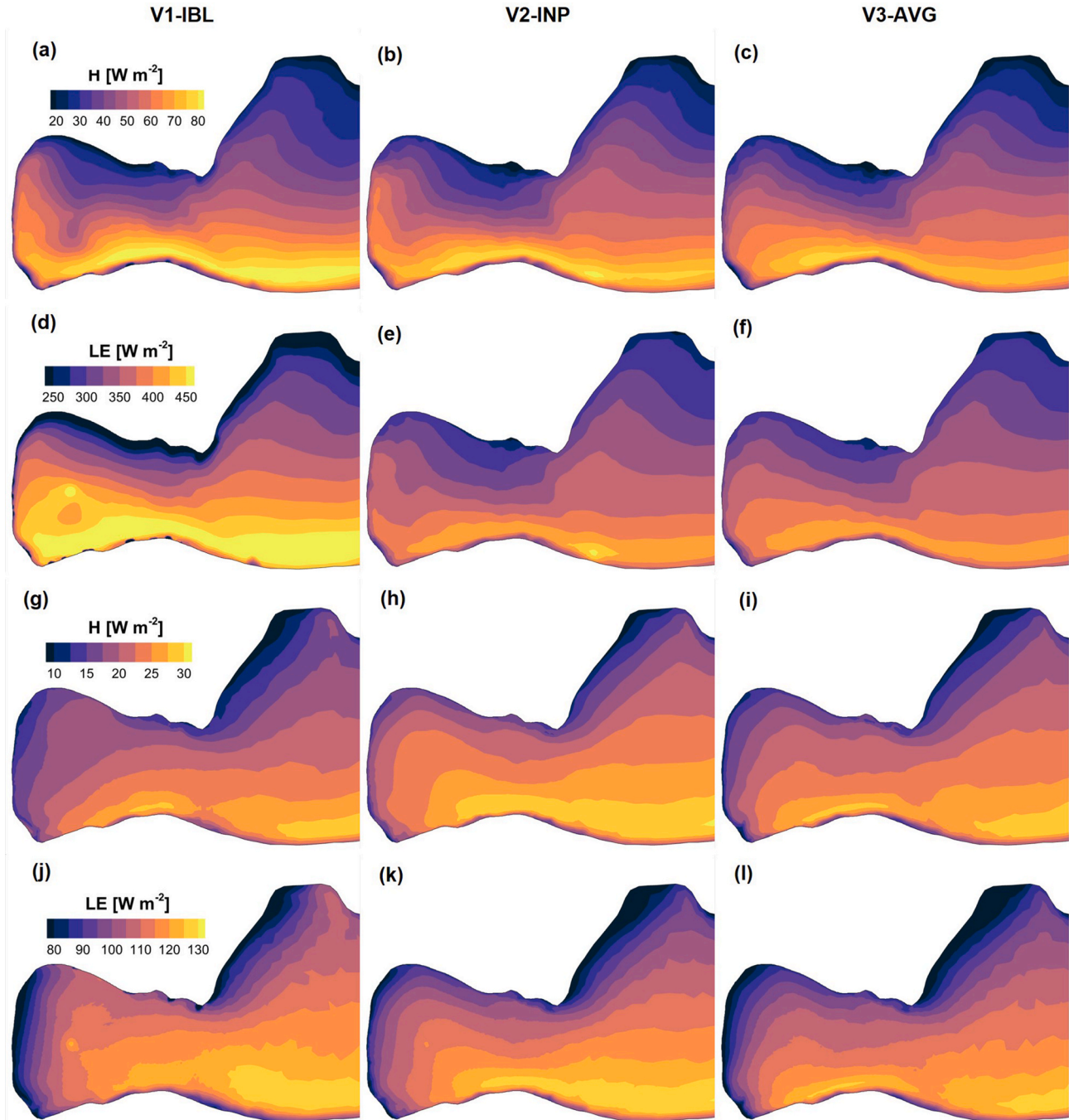


Fig. 12. Time-averaged (a–c, g–i) sensible and (d–f, j–l) latent heat flux fields (a–f) on 28 June 2019, and (g–i) on 1 July 2019 in the Keszthely and part of Szigliget basins of Lake Balaton with the (a, d, g, j) V1-IBL, (b, e, h, k) V2-INP, and (c, f, i, l) V3-AVG model simulations.

Fig. 12g-i shows the time-averaged heat exchange fields of a weak southerly wind event on 1 July 2019. In Fig. 12g-i, the sensible heat fluxes are about 25% lower in the northern part of the Szigliget Basin and most of the Keszthely Basin. In Fig. 12j-l, the latent heat fluxes show high variability with the different wind forcing scenarios. The V1-IBL model shows much lower fluxes in the western part of the Keszthely Basin because of the mentioned two-way interaction. Note that both periods show slightly higher values near station B. At this station, we measured the wind using a sonic anemometer, a fundamentally different instrument from the other standard wind stations, which may result in a bias in measured winds. Based on our experience, the sonic anemometer provided higher values during very low winds than the other stations. We do not show a direct comparison of observed and modeled turbulent heat fluxes here because of the well-known energy balance closure problem and the related underestimation of the heat fluxes of the eddy covariance technique (Lükő et al., 2022b; Mauder et al., 2020).

5.4. Long-term averages

Finally, we evaluate the effect of the wind's spatial variability on the horizontal structure of temperature, stratification, and heat fluxes by calculating long-term averages over the whole, nearly 1-month simulation period between 23 June and 18 July 2019. In Fig. 13a-c, Lake

Balaton's depth-averaged temperatures can be seen with different wind forcings. The effect of the IBL is significant on the spatial variability of the temperatures; the lake is colder along the fetch, especially near the south shoreline. However, the north-to-south temperature difference along the fetch is much lower than in the cases of V2-INP and V3-AVG. The most remarkable deviations are along the south shoreline and in the eastern Siófok Basin. V2-INP resulted in a more similar temperature distribution to V3-AVG than V1-IBL. This means that even if the spatial variability of wind is taken into account, the interpolation alone cannot reproduce the effect of the IBL development-related changes. Fig. 13d-f show the potential energy anomalies, which have similar spatial structures for all the wind forcings; however, V1-IBL resulted in somewhat lower values during this 1-month summer period. On a monthly scale, we can conclude that IBL-related impacts play a lesser role in the spatial structure of stratification, and the bathymetry dominates.

In Fig. 13g-l, we show the modeled turbulent heat flux fields for the different wind model scenarios. We recall here that only the wind field was spatially variable for two scenarios, while air temperature, relative humidity, and radiative fluxes were all homogeneous for all the cases. During this 1-month period, the wind direction was variable, having winds from all directions. Nevertheless, both sensible and latent heat fluxes increased along the prevailing wind direction from northwest to southeast, even for the homogeneous wind forcing (V3-AVG). The

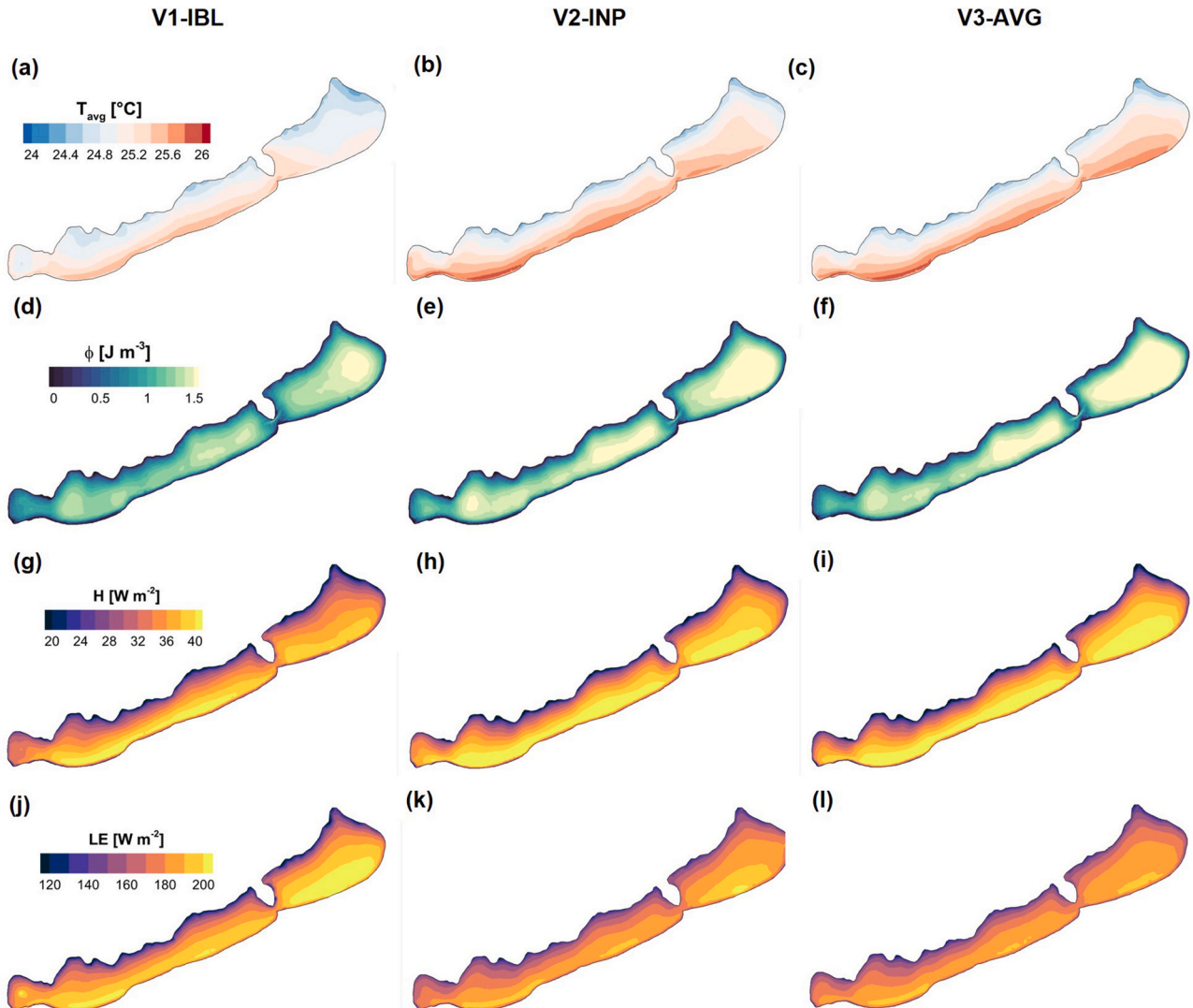


Fig. 13. Time-averaged (a–c) depth-averaged temperature, (d–f) potential energy anomaly, (g–i) sensible, and (j–l) latent heat flux fields between 23 June and 18 July 2019 in Lake Balaton with the (a, d, g, j) V1-IBL, (b, e, h, k) V2-INP, and (c, f, i, l) V3-AVG model simulations.

similarity with the surface temperature distribution means that turbulent surface fluxes were strongly governed by it. As the water was warmer along the south shore, the air-to-water temperature difference also became higher, leading to increased sensible heat loss (Fig. 13g-i). In the case of latent heat, the higher temperature resulted in a higher specific humidity of saturation at the surface, consequently a higher specific humidity gradient, and finally, an elevated latent heat loss (Fig. 13j-l). V2-INP and V3-AVG showed slight differences for both fluxes. In contrast, sensible heat was reduced on the south shoreline for V1-IBL. This is mainly caused by the reduced surface temperature. The latent heat loss in the case of V1-IBL likely caused the colder temperature. Following this, the latent heat loss was increased, implying that it was more sensitive to the wind speed than the sensible heat flux. Due to the wind speed dependency of latent heat, the difference compared to the V2-INP and V3-AVG models could be up to 10% in the open water, which can lead to a significant evaporation error if the wind field is not represented accurately in the model. In short, sensible and latent heat fluxes varied with an opposite sign by including the wind speed development due to the IBL: sensible heat decreased, and latent heat increased along the south shore compared to V2-INP and V3-AVG. Again, this means that interpolation cannot fully reproduce the impact of wind speed acceleration.

5.5. Limitations of the wind model

We showed that applying the IBL-based wind model has many advantages for modeling a shallow mid-size lake, such as improved currents and temperature dynamics. The model's generality is clear, since the surface roughness changes from land to water occurs in theory if the lake surface is large enough. Consequently, it could be applied to other lakes if wind stations are available at the shoreline or over the lake. Nevertheless, the method has some limitations.

The suggested wind model consisted of two parameters to calibrate: the roughness length of the land and the water surface. In the former case, due to simplicity, we used the same value for roughness length at all land stations, which can vary as a function of land cover (e.g., urban areas, tall or low canopy). However, the model can be easily updated with different land roughness values. In our case, wind profile data were available over the land from which the roughness length could be derived. In the latter case, we used the Charnock formulation, and its coefficient was determined using eddy covariance measurements. If the roughness lengths are unknown, the calibration of the IBL model is possible if there are several wind measurements available at upwind and downwind shores or over the lake (leave-one-out cross-validation). Furthermore, roughness length values can be determined by involving water level fluctuations to ensure accurate momentum flux. This is somewhat more time-consuming, as the lake simulation must be involved in the calibration process.

In this study, we used a constant Charnock coefficient, but it can also be wave dependent. These formulations are also available directly for the roughness length, which would be beneficial to apply due to the waves' fetch-dependency. This provides a potential linkage between wave and wind models.

The suggested model has another limitation, as it assumes logarithmic wind profiles both over the land and the lake. This may not be valid in certain cases. In the case of strong orographic effects, such as those found in alpine lakes, the vertical wind component is not zero (Valerio et al., 2017). In the case of lake-land breezes, divergence occurs in the wind field, resulting in upwelling or downwelling air flows over the lake. However, this latter case is characterized by low wind speeds and, consequently, restrained circulation and turbulent heat exchanges. The wind profile can deviate from a truly logarithmic shape due to atmospheric stratification too. In Lake Balaton, Lükő et al. (2022b) showed that stratification does not play a crucial role; however, this is not necessarily true in other lakes. Nevertheless, the wind profile equations can be extended with stability corrections following the Monin-Obukhov

similarity theory.

For large lakes with extensive fetches, the IBL-based spatial variability of the wind and its overall effect can be smaller. Here, the internal boundary layer over water could even fully reach its equilibrium state without further development along the fetch. However, the IBL wind model can be useful for modeling these large lakes' coastal processes, where wave and turbulent exchange processes are still strongly varying in the wind direction.

Finally, our wind model's accuracy may depend on the spatial resolution of the wind stations. We suggest analyzing the variability of the available stations first to determine and select the representative ones to be involved in the wind field estimation, as some stations can represent strongly local conditions.

6. Conclusion

Wind fields over lakes often exhibit strong spatial variability driven by mesoscale processes and the development of an atmospheric internal boundary layer, which are often neglected due to simplicity. We aimed to introduce a simple IBL-based wind field calculation method and quantify the effect of this phenomenon on lake hydrodynamics and thermal structure.

Lake thermodynamics are governed by coupled interactions between wind, waves, turbulent momentum, and heat fluxes and circulation (Fig. 1). Lake Balaton was an excellent case study with available hydrometeorological observations and a validated 3D lake model. We analyzed multiple wind-forcing scenarios using continuously measured wind data from stations around the lake and compared simulations.

Simulated water level fluctuations were similar across different wind forcing scenarios and agreed with observations. In contrast, the open-water currents and circulation patterns differed significantly in both magnitude and direction. The wind field that accounted for IBL development reproduced observed flow conditions significantly better. These circulation differences also altered the spatial structure of lake temperature, which was confirmed by in situ and satellite observations. While horizontal transport contributes to these thermal differences, the dominant mechanism is the spatial variability of the turbulent heat fluxes introduced by the IBL-based modified wind field.

We conclude that accounting for IBL development in wind forcing substantially improves the simulation of lake temperature dynamics for a mid-sized lake. Therefore, studies focusing on transport or vertical mixing should incorporate IBL-based wind models. This approach is computationally simple, semi-analytical, and physics-based yet yields measurable improvements in model performance.

CRedit authorship contribution statement

Gabriella Lükő: Writing – original draft, Visualization, Validation, Software, Resources, Methodology, Investigation, Funding acquisition, Formal analysis, Data curation. **Tamás Krámer:** Writing – review & editing, Software, Funding acquisition. **Péter Torma:** Writing – original draft, Supervision, Methodology, Funding acquisition.

Declaration of competing interest

The authors declare that they have no known competing financial interests or personal relationships that could have appeared to influence the work reported in this paper.

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23).

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