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Key Points:

- The bulk transfer coefficients exhibit a substantial increase at low wind speeds in lakes which could partially be associated with gusts
- Average drag coefficients significantly correlated with lake surface area at winds exceeding 3 m s^{-1}
- An empirical function describing the dependence of the transfer coefficients on wind speed could be beneficial when modeling lakes

Abstract

The drag coefficient (C_{DN}), Stanton number (C_{HN}) and Dalton number (C_{EN}) are of particular importance for the bulk estimation of the surface turbulent fluxes of momentum, heat and water vapor at water surfaces. Although these bulk transfer coefficients have been extensively studied over the past several decades mainly in marine and large-lake environments, there are no studies focusing on their synthesis for many lakes. Here, we evaluated these coefficients through directly measured surface fluxes using the eddy-covariance technique over more than 30 lakes and reservoirs of different sizes and depths. Our analysis showed that generally C_{DN}, C_{HN}, C_{EN} (adjusted to neutral atmospheric stability) were within the range reported in previous studies for large lakes and oceans. C_{HN} was found to be on average a factor of 1.4 higher than C_{EN} for all wind speeds, therefore, likely affecting the Bowen ratio method used for lake evaporation measurements. All bulk transfer coefficients exhibit substantial increase at low wind speeds (< 3 m s^{-1}), which could not be explained by any of the existing physical approaches. However, the wind gustiness could partially explain this increase. At high wind speeds, C_{DN}, C_{HN}, C_{EN} remained relatively constant at values of $2 \cdot 10^{-3}$, $1.5 \cdot 10^{-3}$, $1.1 \cdot 10^{-3}$, respectively. We found that the variability of the transfer coefficients among the lakes could be associated with lake surface area or wind fetch. The empirical formula $C = b_1 \left[1 + b_2 \exp(b_3 U_{10})\right]$ described the dependence of C_{DN}, C_{HN}, C_{EN} on wind speed well and it could be beneficial for modeling when coupling atmosphere and lakes.

1 Introduction

The major process that governs the interaction between the atmosphere and surface waters is the turbulent exchange of momentum, heat and gases at the air-water interface. Although lakes and reservoirs occupy only about 3% of the land surface area (Downing et al., 2006), they are known to have an impact on local weather and climate. For example, lakes affect the stability of the atmosphere above (Sun et al., 1997), leading to the formation of clouds and precipitation on the shores (Changnon & Jones, 1972; Kato & Takahashi, 1981; Eerola et al., 2014; Thiery et al., 2016). Furthermore, lakes and reservoirs are recognized as significant contributors to the global carbon cycle by emitting significant amounts of carbon dioxide and methane (DelSontro et al., 2018; Rosentreter et al., 2021).

The past three decades have seen a rapid development of lake models (Stepanenko et al., 2014) and their incorporation into numerical weather and climate prediction models (Ljungemyr et al., 1996; Salgado & Le Mogne, 2010; Mironov et al., 2010). Experiments on the coupling of lakes and the atmospheric model revealed their beneficial impact on the weather prediction quality (Balsamo et al., 2012). A number of case studies have demonstrated the importance of lakes for extreme local weather phenomena, such as lake-effect snow over Great American lakes (Fujisaki-Manome et al., 2020), deep hazardous convection over Great African lakes (Thiery et al., 2016), wind speeds over Lake Superior (Desai et al., 2009), or stratiform cloudiness in winter over Lake Ladoga (Eerola et al., 2014). Thus, an accurate representation of the exchange of momentum, heat and water vapor at the air-water interface in water bodies is essential.

In state-of-the-art, momentum, sensible and latent heat fluxes are usually determined based on gradient approaches utilizing transfer coefficients (bulk transfer coefficients) and easy to measure meteorological and limnological variables, i.e., wind speed, air temperature, air humidity and water surface temperature (Stull, 1988). The exchange at the air-water interface and therewith the bulk coefficients are controlled by the boundary-layer turbulence. The bulk exchange coefficient of momentum, known as the drag coefficient (C_D, C_{DN}) (Garratt, 1977), is of particular importance for all air-water fluxes. The coefficients of heat (C_H, C_{HN}) and water vapor exchange (C_E, C_{EN}) are also known as Stanton and Dalton numbers, respectively. Here, "N" stands for "neutral" transfer coefficients, corresponding to the neutral thermal stability of the atmosphere. The transfer coefficients depend on the measurement height of the mean wind speed, air temperature and humidity, respectively, and for this reason, they are usually reported for the reference meteorological height of 10 m.

A considerable amount of studies has been published on the momentum flux and the drag coefficient starting from the early 1950s when the fundamental work, presenting the theory later on named as Monin-Obukhov similarity theory, was published (Monin & Obukhov, 1954; Obukhov, 1971). The theory aims at describing the structure of turbulence in the atmospheric surface layer about several tens of meters thick with the assumption of the fluxes being constant and independent of height. Similarity laws introduce functional relations to derive the universal shapes for the vertical profiles of different quantities for atmospheric thermal stability other than neutral. During the past decades, considerable effort has been devoted to define the exact form of these similarity functions (Paulson, 1970; Businger et al., 1971; Högström, 1988; Zilitinkevich & Calanca, 2010).

As the drag coefficient is one of the key parameters in atmospheric and lake models, the errors in its parameterization lead to errors in the bulk flux estimates. Therefore, numerous early studies focused on exploring different parameterizations of the drag coefficient over the land and oceans in terms of wind speed, atmospheric stability, and surface roughness, which could be a function of the surface wave field (for oceans) (Garratt, 1977; Kantha & Clayson, 2000).

Most of the extensive field measurement campaigns over the oceans have been conducted during the last 30 years of the 20th century (Large & Pond, 1981; Godfrey & Beljaars, 1991; Smith et al., 1996; Fairall et al., 1996). Several of these studies agreed that the drag coefficient linearly increases with increasing wind speed ignoring the state of the wave field. More recent parameterizations of the drag coefficient (e.g., the COARE algorithm, (Edson et al., 2013)), however, include a wave dependence. There is still an ongoing scientific discussion concerning the importance of waves and how their impact could be included in the models (Wu et al., 2019).

Along with the studies in the marine environment, the research started to focus on the drag coefficient estimated from measurements over large and medium-sized lakes (e.g., Hicks, 1972; Donelan, 1982; Graf et al., 1984; Simon, 1997). To date, in total, about two dozen studies focusing on lakes have been published since the beginning of the 1970s. In reviewing these studies below, we separated them by the wind speed regime they were interested in. It is usually assumed that surface wave development starts when the wind speed exceeds 3-4 m s^{-1} (Ataktürk & Katsaros, 1999; Kantha & Clayson, 2000). This is also supported by wave measurements in several lakes (Simon, 1997; Guseva et al., 2021). Therefore, we intend to separate the two wind speed regimes using this threshold.

At the “high” wind speed regime (wind speed exceeds 3 m s^{-1}), in the most simplified way, the surface waves are assumed to be fully developed, and the surface roughness length is described as a function of wind stress, which is commonly known as Charnock relationship (Charnock, 1955). However, this assumption might not hold for lakes with limited wind fetch (Donelan, 1990; Geernaert, 1990). Thus, some research has been made to study the drag coefficient as a function of the surface wave state, for example, taking into account wave characteristics such as the wave age (Donelan, 1982; Ataktürk & Katsaros, 1999). Vickers & Mahrt (1997) reported that for a given wind speed the drag coefficient tends to be larger for younger steeper waves representative of short wind fetches than for longer fetches. Ataktürk & Katsaros (1999) could significantly reduce the scatter in the estimated drag coefficients by considering waves in the parameterization of the surface roughness length. However, these studies mainly examined large lakes and only a few were performed in lakes with short fetch and young wave states (Babanin & Makin, 2008; Lükő et al., 2020). Given the fact that the surface wave measurements in lakes are not often available, their effect still could be investigated via analyzing the relationship between the drag coefficient and fetch length.

At the “low” wind speed regime, several studies found that the neutral drag coefficient in lakes and oceans tended to increase by an approximate factor of two up to ten compared to the value of $1.3 \cdot 10^{-3}$ (corresponding to a typical value of open water surface roughness (Foken, 2008)) (Wüest & Lorke, 2003; Woolway et al., 2017). Although the wind speed dependence is obvious, many numerical and empirical studies employ a constant value for the drag coefficient, which is often considered as a model tuning parameter (Stepanenko et al., 2014). Despite the fact that there have been many attempts to address the reasons of such increase, there is still no consensus in the scientific community. The low wind speed regime was first described as the aerodynamically smooth flow, when the surface waves are buried within the viscous sublayer and the surface roughness is described as a function of the thickness of this layer (Schlichting, 1968). On the contrary, Wu (1988) proposed that the flow is aerodynamically rough and that capillary gravity waves play a key role at low wind speeds. Surface roughness length was described as a function of the water surface tension. As an additional reason for the increase of the drag coefficient at low wind speed, Godfrey & Beljaars (1991) and Grachev et al. (1998) considered the concept of gustiness, which assumes that at "zero" wind speeds there are dry random convective motions – gusts – in the convective boundary layer (CBL). Thus, the "traditional" formulation of the drag coefficient has been modified using the scalar-averaged wind speed (not the vector-averaged wind speed) to account for gusts. All the possible mechanisms mentioned above were addressed in the recent work by Wei et al. (2016). They concluded that none of them explained the increase of the drag coefficient at low wind speeds. However, they found it can be explained by the increase in the turbulent kinetic energy and enhanced buoyant energy. Similar to (Grachev et al., 1998), Sahlée et al. (2014) and Liu et al. (2020) related the increase of the drag coefficient with nonlocal effects, such as the penetration of large convective eddies into the surface layer from the atmosphere above. Liu et al. (2020) introduced the factor describing this effect and estimated it from two-level measurements of wind speed (however, over the land surface and only for neutral conditions). Another formulation of the drag coefficient at low wind speeds was done by Zhu & Furst (2013) relating the drag coefficient to the turbulent kinetic energy budget. However, their fitting coefficients for the drag coefficient formula were found to be site-specific (Liu et al., 2020).

Other studies on the bulk transfer coefficients in lakes branched off from the main direction – potential physical mechanisms – with a focus on the possible correlation between the bulk transfer coefficients and some lake characteristics. Among them are lake depth at the measurement location (Panin et al., 2006), lake surface area (Read et al., 2012; Woolway et al., 2017), wind fetch at the measurement location (Lükő et al., 2020) and lake biota, e.g, submerged macrophytes (Xiao et al., 2013). All studies showed a strong dependence of the transfer coefficients on these lake characteristics. The drag coefficient tends to decrease with increasing water depth, lake area, fetch and in the presence of water plants at the water surface. It is important to note that although Panin et al. (2006) and Woolway et al. (2017) revealed the correlation between the transfer coefficients and the lake parameters, the estimation of the transfer coefficients was based either on bulk parameterization (Woolway et al., 2017), or was compared to other studies where there were no direct flux measurements (Panin et al., 2006).

Fewer studies have been published on the Stanton and Dalton numbers. Although the measurements in the oceans showed their obvious increase at low wind speeds, both transfer coefficients were considered as fairly constant with a value of $1.1 \cdot 10^{-3}$ (review of these measurements in (Kantha & Clayson, 2000)). First measurements conducted in lakes revealed this value being higher and equal to ~$1.5 \cdot 10^{-3}$ (Harbeck, 1962; Hicks, 1972) or $1.9 \cdot 10^{-3}$ (Strub & Powell, 1987). Harbeck (1962) and Brutsaert & Yeh (1970) reported a dependence of the Dalton number on the lake surface area. Heikinheimo et al. (1999) summarized that the Dalton number is generally known to be less dependent on the wind speed. From the most recent studies (Xiao et al., 2013; Li et al., 2016; Wei et al., 2016; Dias & Vissotto, 2017), there is evidence that both coefficients depend on the wind speed and that the Stanton number is higher than the Dalton number by approximately a factor of 1.3. This indicates that the earlier assumption of the equality of both coefficients may not be valid for lakes.

The eddy-covariance (EC) technique is a micrometeorological method to directly measure momentum, heat, water vapor and greenhouse gas fluxes (Foken, 2008). It is based on the correlation between turbulent fluctuations of vertical wind speed and scalar air properties. Using this technique, one can obtain the spatial and temporal average of turbulent fluxes originating from an area called footprint and a period of meteorological stationarity (Lenschow et al., 1994; Sun et al., 2006; Burba & Anderson, 2010; Foken et al., 2012). Nowadays, the EC technique is commonly used over lakes (Blanken et al., 2000; Vesala et al., 2006; Nordbo et al., 2011; Lee et al., 2014; Mammarella et al., 2015; Spank et al., 2020; Golub et al., 2021). However, several studies reported difficulties in measuring the wind stress at weak winds, which resulted in large uncertainties (Kantha & Clayson, 2000). Low wind speed conditions are more relevant for lakes and specifically small lakes that are the most abundant inland water bodies (Downing et al., 2006).

In this study, we evaluate the first multiple water body estimates of bulk transfer coefficients and their dependencies on wind speed and water body characteristics using EC data measured above lakes. The analysis aimed at answering the following research questions: 1) what are the typical values for the bulk transfer coefficients and their variability among lakes and reservoirs? 2) how do the values compare with the reported transfer coefficients for oceans and other lakes? 3) can the mechanistic approaches mentioned above describe the transfer coefficients at low wind speed regime? 4) is there a consistent dependence of the transfer coefficients on lake characteristics, such as water depth, lake area and wind fetch? In the sections below, we examine possible answers.

2 Materials and Methods

2.1. Eddy-covariance dataset

For this analysis, most of the existing EC data measured by various researchers over lakes and reservoirs were extracted from open access databases and repositories of published papers. The fluxes that are reported in the datasets were calculated using different software (e.g., EddyPro (LI-COR, Inc, 2021), TK3

(Mauder & Foken, 2015), EddyUH (Mammarella et al., 2016)). In total, we obtained data for 23 lakes and 8 reservoirs located in the arctic, subarctic, temperate and subtropical zones (Figure 1, Table S1). The water bodies are located in different landscapes, including mountains (e.g., Lake Lunz, Austria or Lake Klöntal, Switzerland), forests (e.g., Lake Vanajavesi, Finland), and arctic landscapes. The EC mast at each lake or reservoir was installed either on a floating or bottom-fixed platform, on shore, or on small islands. The measurement height ranged between 1.3 m and 16.1 m with 2 m being the most frequent height among all datasets. Elongated shapes of the lakes or shore/island locations were the subject of wind direction filtering to ensure that the measured surface fluxes were originating from water. Approximately half of the water bodies in this study had a surface area (A_s [km^2]) smaller than 10 km^2 with an average wind fetch (F_{ave} [m]) ranging from 168 m to 1553 m. The fetch grid was estimated from the map as the distance from the measurement location to the shore with the corresponding wind direction. Then, the time series of the fetch was interpolated from this grid using the measured wind directions. The average fetch was calculated as the mean distance for the filtered wind directions. The rest of the lakes and reservoirs were larger: the maximum surface area of $2.6 \cdot 10^4$ km^2 and the maximum mean fetch of $2.6 \cdot 10^4$ m refer to one of the Great Lakes – Lake Erie in the USA. The maximum depth (D_{max} [m]) varied between 1.3 m (Lake Villasjön, Sweden) and 89 m (Rappbode Reservoir, Germany). Each EC dataset contained the estimated variables averaged over 30 min intervals.

The variables included wind speed (U_z [m s^{-1}]), wind direction (WD [°]), friction velocity (u_* [m s^{-1}]) as a quantity characterizing the momentum flux ($\tau = \rho_a u_*^2$ [kg m^{-1} s^{-2}], ρ_a – air density [kg m^{-3}]), air temperature (T_a [°C]), turbulent fluxes of sensible heat (H [W m^{-2}]), and latent heat ($L_v E$ [W m^{-2}]), the latter referred to in this paper as water vapor flux as well. Water temperature was provided either as skin surface temperature (T_s [°C]) or bulk water temperature, measured at 0-0.5 m water depth (T_w [°C]). The skin temperature was observed with an infrared thermometer or calculated from outgoing longwave radiation, both corrected for the reflectance of incoming longwave radiation. Some lakes or reservoirs had only momentum flux data, resulting in fewer estimates of heat and water vapor transfer coefficients. Parameters such as precipitation considered as a factor for filtering the data was not available for all datasets. The duration of the EC measurements ranged from 11 days (Lake Wohlen, Switzerland) to 2243 days (or ~ 6.1 years, Lake Dagow, Germany) with a median duration of 155 days.

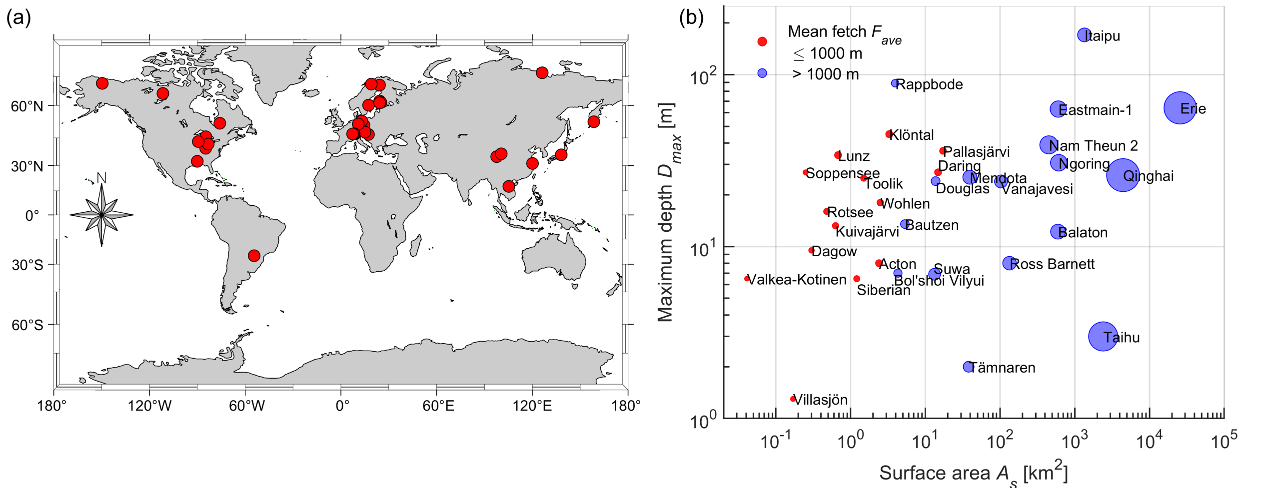

Figure 1. (a) Geographical distribution of the eddy-covariance measurements over the lakes and reservoirs u

2.2. Data filtering and averaging

The individual datasets used in the analysis were subject to filtering with the following different criteria:

1. filtering based on stationarity and integral turbulence test quality flags;
2. restriction of the wind directions to ensure >90% of footprint was originated from water;
3. removing periods with ice cover;
4. removing periods with precipitation (if data on precipitation was available);
5. removing periods with low fluxes $u_* < 0.05$ m s^{-1}, $|H|$, $|E| < 10$ W m^{-2} following, e.g., Li et al. (2016) and Wei et al. (2016);
6. removing periods with floating vegetation on the water surface (only for Lake Suwa, Japan).

Quality screening of EC data is known to be site- and instrument- specific (Burba & Anderson, 2010). The data were either available in filtered form, or they contained the quality flags provided by the software. Non-filtered datasets included quality flags for each flux value (momentum, sensible and latent heat fluxes) to ensure the stationarity of the time series (homogeneity of the flow) and developed turbulent conditions (Foken et al., 2004; Foken & Wichura, 1996).

Removing wind directions was site-specific and we carefully studied each individual site. We only accepted the data from periods when wind was blowing from the lake with sufficient fetch. We specified the accepted wind directions for each site in Table S1. We focused on open-water conditions and discarded ice-covered periods either using the water temperature time series or interval camera data. For Lake Suwa we removed the approximate periods when floating vegetation appeared on the water surface using interval camera data, however, for other sites this kind of data was not available. For some sites, all erroneous data due to rain interference and site maintenance were filtered by data providers, or we removed periods with precipitation (if data were available). There is no common convention for selecting the thresholds of flux values for filtering the fluxes and they were taken from the literature (Wei et al., 2016). We describe the effect of these filters on the data as well as we compare different types of averaging applied to data in Text S1. This pre-analysis revealed that the logarithmic bin averaging for the derived quantities is an adequate measure to use in the following sections.

2.3. Transfer coefficients

Turbulent fluxes of momentum ( ), sensible heat (H) and latent heat (E) at the water surface are expressed as:

$$\tau = -\overline{u'w'}\rho_a = \rho_a u_*^2 = \rho_a C_D U_{10}^2, \tag{1a}$$

$$H = \rho_a c_p \overline{w'T'} = -\rho_a c_p T_* u_* = \rho_a c_p C_H U_{10}\left(T_s - T_{10}\right), \tag{1b}$$

$$L_v E = \rho_a L_v \overline{w'q'} = -\rho_a L_v q_* u_* = \rho_a L_v C_E U_{10}\left(q_s - q_{10}\right), \tag{1c}$$

where $\overline{u'w'}$ is the covariance of horizontal (u') and vertical (w') wind velocity fluctuations; $\overline{w'T'}$ [m s^{-1} K], $\overline{w'q'}$ [m s^{-1} kg kg^{-1}] are the covariances of vertical wind velocity and air temperature (T') and specific humidity (q') fluctuations. U_{10} is wind speed at 10 m height, T_s and T_{10} [K] are the surface water temperature and the air temperature at 10 m height, respectively, q_s and q_{10} [kg kg^{-1}] are the specific humidity at the air-water interface (estimated from surface temperature) and at 10 m height, respectively. c_p [J kg^{-1} K^{-1}] is the specific heat of air at constant pressure, and L_v [J kg^{-1}] is the latent heat of vaporization. $T_* = \frac{-\overline{w'T'}}{u_*}$ and $q_* = \frac{-\overline{w'q'}}{u_*}$ are temperature and specific humidity scales, respectively. The standard sign convention is that the momentum flux is defined as positive downward, while sensible and latent heat fluxes as positive upward (Kaimal & Finnigan, 1994). Using measured flux data from the obtained EC datasets, the transfer coefficients can be derived from Eq. (1a-c) as follows:

$$C_D = \frac{u_*^2}{U_{10}^2}, \tag{2a}$$

$$C_H = \frac{\overline{w'T'}}{U_{10}(T_s - T_{10})}, \tag{2b}$$

$$C_E = \frac{\overline{w'q'}}{U_{10}(q_s - q_{10})}. \tag{2c}$$

Wind speed, air temperature (T_z) and specific humidity (q_z) measured at a certain height z were converted to a standard height of 10 m considering stability of the atmosphere following the equations:

$$U_{10} = U_z - \frac{u_*}{\kappa}\left[\ln\left(\frac{z}{10}\right) - \psi_u\left(\frac{z}{L}\right) + \psi_u\left(\frac{10}{L}\right)\right], \quad (3a)$$

$$T_{10} = T_z - \frac{T_*}{\kappa}\left[\ln\left(\frac{z}{10}\right) - \psi_T\left(\frac{z}{L}\right) + \psi_T\left(\frac{10}{L}\right)\right], \quad (3b)$$

$$q_{10} = q_z - \frac{q_*}{\kappa}\left[\ln\left(\frac{z}{10}\right) - \psi_T\left(\frac{z}{L}\right) + \psi_T\left(\frac{10}{L}\right)\right], \quad (3c)$$

where κ is the von Kármán constant, L [m] is the Obukhov length, $\psi_u\left(\frac{z}{L}\right)$ is the stability function which is the integral of the empirical universal function for the momentum flux and $\psi_T\left(\frac{z}{L}\right)$ – the same for sensible and latent heat (Businger et al., 1971). In the literature, z/L is usually denoted as the non-dimensional stability parameter ζ. To remove the effect of atmospheric stability on the magnitude of the transfer coefficients, C_D, C_H, C_E are converted to their neutral counterparts $C_{\mathrm{DN}}, C_{\mathrm{HN}}, C_{\mathrm{EN}}$ (i.e. for neutrally-stratified atmospheric conditions) (Large & Pond, 1981):

$$C_{\mathrm{DN}} = \kappa^2\left[\ln\left(\frac{10}{z_0}\right)\right]^{-2} = C_D\left[1 + \kappa^{-1}\, C_D^{\frac{1}{2}}\psi_u\left(\frac{10}{L}\right)\right]^{-2}, \quad (4a)$$

$$C_{\mathrm{HN}} = C_D\left[1 + \kappa^{-1}\, C_D^{\frac{1}{2}}\psi_u\left(\frac{10}{L}\right)\right]^{-1}\left[\frac{C_D}{C_H} + \kappa^{-1}\, C_D^{\frac{1}{2}}\psi_T\left(\frac{10}{L}\right)\right]^{-1}, \quad (4b)$$

$$C_{\mathrm{EN}} = C_D\left[1 + \kappa^{-1}\, C_D^{\frac{1}{2}}\psi_u\left(\frac{10}{L}\right)\right]^{-1}\left[\frac{C_D}{C_E} + \kappa^{-1}\, C_D^{\frac{1}{2}}\psi_T\left(\frac{10}{L}\right)\right]^{-1}, \quad (4c)$$

where $z_0$is the surface roughness length. For our calculations, we used the Kansas-type stability functions (Businger et al., 1971) in the form of Högström (1988), which is the most frequently applied form (Foken, 2008). $C_{\mathrm{DN}}, C_{\mathrm{HN}}, C_{\mathrm{EN}}$ were estimated for 31, 24, 23 water bodies under study, respectively, depending on the flux data availability (see details about each lake or reservoir in Table S1 and in data repository 10.5281/zenodo.6597829). After calculation of C_H, C_E (Eq. 2b-2c), we removed negative values that were a result of the inconsistency between sign of the measured flux and the temperature difference (probably related to the measurements random uncertainty).

In the scientific community, there has been an ongoing discussion on the form of the transfer coefficients to be presented. For example, some studies focused only on neutral values of the drag coefficient (Li et al., 2016) or some considered the drag coefficient non-adjusted to their neutral counterpart (C_D). Other studies addressed the so-called "effective" drag coefficient, which was derived as the slope coefficient for the linear relationship between u_*^2 and U_{10}^2 (Xiao et al., 2013). We examine the difference between C_D, C_H, C_E and $C_{\mathrm{DN}}, C_{\mathrm{HN}}, C_{\mathrm{EN}}$ in Section 3.1.

2.4. Parametrizations of the drag coefficient at low and high wind speeds

2.4.1. *Smooth flow*

Previous studies focused on the parameterizations of surface roughness length z_0 (see Eq. 4a) to assess wind speed dependence of the drag coefficient (e.g., Ataktürk & Katsaros (1999). In our study, we compared C_{DN} estimated from measured momentum fluxes with the existing approaches. One of the approaches is based on the smooth flow regime at low wind speed (< 3 m s^{-1}), where the thickness of the viscous sublayer (δ_ν) determines the aerodynamic roughness of the interface (Schlichting, 1968), and not the physical roughness of the water surface:

$$\delta_\nu = z_0 = \alpha \frac{\nu}{u_*}, \quad (5)$$

where $\alpha = 0.11$ and $\nu = 1.6 \times 10^{-5}$ [m^2 s^{-1}] is kinematic viscosity of air. z_0 can be derived from Eq. 4a as:

$$z_0 = z \ \exp\left(-\frac{\kappa}{\sqrt{C_{\mathrm{DN}}}}\right). \quad (6)$$

Substituting Eq. 6 into Eq. 5, replacing $u_* = \sqrt{C_{\mathrm{DN}}} U_{10}$ and taking the standard height as $z = 10$ m, we obtain the following expression for smooth flow approach:

$$\frac{1}{\sqrt{C_{DN\_SF}}} = \frac{1}{\kappa} \ln\left(\frac{z\sqrt{C_{DN\_SF}}U_{10}}{\nu}\right) - \frac{1}{\kappa}\ln(\alpha), \quad (7)$$

We used measured values of U_{10} as input and solved Eq. 7 iteratively for C_{DN_SF}.

2.4.2. *Capillary waves*

As an alternative method to estimate C_{DN} at low wind speeds, we considered the approach proposed by Wu (1994). He suggested that the wind shear stress in the absence of large gravity waves is related to the ripples (capillary waves). For the capillary waves, the roughness length is related to surface tension ( ) as:

$$z_0 = \alpha_{\mathrm{Wu}} \frac{\sigma}{\rho_w u_*^2}, \quad (8)$$

where $_{Wu} = 0.18$ is an empirical constant and ρ_w is water density. Surface tension at a temperature of 20°C is $= 7.28 \cdot 10^{-2}$ N m^{-1}. In analogy to the smooth flow approach, substitution of Eq. 6 to Eq. 8 and replacement of u_* leads to the expression:

$$\frac{1}{\sqrt{C_{DN\_CW}}} = \frac{1}{\kappa} \ln\left(\frac{z\rho_w C_{DN\_CW} U_{10}^2}{\mathrm{w}}\right), \quad (9)$$

Eq. 9 was solved iteratively for C_{DN_CW} using wind speeds U_{10} from the data sets and z = 10 m.

2.4.3. *Charnock relationship*

With increasing wind speed, the thickness of the viscous sublayer becomes smaller, and the aerodynamic roughness of the water surface (z_0) becomes minimal, before surface gravity waves evolve. At wind speeds exceeding 3 m s^{-1}, waves protrude from the viscous sublayer and surface roughness length increases with increasing wind speed, indicating the transition from a smooth to a rough flow regime. Charnock, (1955) proposed the following equation for surface roughness length over fully developed surface waves, which account for typical oceanic conditions:

$$z_0 = \beta \frac{u_*^2}{g}, \quad (10)$$

where ranges from 0.011 to 0.0185 (Garratt, 1994), g is the gravitational acceleration. Substitution of Eq. 6 into Eq. 10 leads to the following implicit equations that was iteratively solved for C_{DN_CH}.

$$\frac{1}{\sqrt{C_{DN\_CH}}} = \frac{1}{\kappa} \ln\left(\frac{\mathrm{gz}}{C_{DN\_CH} U_{10}^2}\right) - \frac{1}{\kappa}\ln(\beta). \quad (11)$$

2.4.4. *The concept of gustiness*

Grachev et al., (1998) suggested that under strong convective conditions, the wind stress at the water surface is predominantly governed by random convective motions - gusts - in the convective boundary layer (CBL), whereas the mean wind speed vector can even become zero (Godfrey & Beljaars, 1991). These large convective eddies embrace the entire CBL and affect the turbulence regime in the atmospheric surface layer. Grachev et al. (1998) formulated a new approach to estimate the drag coefficient using this concept. According to their study, the gustiness could explain the apparent increase of the drag coefficient estimated using the traditional equation (Eq. 2a, 4a) at low wind speeds. The estimated drag coefficient accounting for gusts was a factor of 1.5 to 6 smaller at wind speeds below 2 m s^{-1} in comparison with the drag coefficient calculated from Eq. 2a, 4a. The gustiness concept is widely accepted and used in the COARE algorithm to estimate air-sea fluxes (Fairall et al., 2003).

The effect of gustiness on the drag coefficient can be accounted for by the so-called gustiness factor G, which corresponds to the ratio of the scalar-averaged

($\widetilde{U}_{10}$) to vector-averaged wind speed. Following (Grachev et al., 1998), G can be parameterization in terms of the convective velocity scale w_*:

$$G^2 = \frac{\widetilde{U}_{10}^2}{U_{10}^2} = 1 + \left(\frac{\gamma w_*}{U_{10}}\right)^2, \quad (12)$$

where $\gamma = 1.2$ is an empirical constant (Beljaars, 1995) and w_* is expressed as:

$$w_* = \left(g z_i \frac{\overline{w' T_v'}}{T_v}\right)^{1/3}, \quad (13)$$

where, T_v [K] is the virtual temperature, z_i is the CBL height, defined as the height of the lowest inversion. Previous studies used the fixed height of the CBL equal to 1000 m (Beljaars, 1995). We denote two corresponding types of gustiness factor as G_{wind} and G_{conv}.The new relationship between neutral gustiness drag coefficient C_{DNG} and its gustiness counterpart $\widetilde{C}_{\mathrm{DG}}$ is:

$$C_{\mathrm{DNG}}^{-1/2} = \widetilde{C}_{\mathrm{DG}}^{-1/2} + \frac{\psi_u(\frac{z}{L})}{\kappa}, \quad (14)$$

where $\widetilde{C}_{\mathrm{DG}} = C_{\mathrm{DG}}/G^2$ and $C_{\mathrm{DG}} = \left(\frac{\tilde{u}_*}{U_{10}}\right)^2$, where $\tilde{u}_*$ is the scalar-averaged friction velocity. Akylas et al. (2003) investigated the combinations with different averaging procedures and suggested that vector-averaged friction velocity u_* is more appropriate to use with scalar-averaged wind speed for all wind speed classes. We applied this approach for cases with unstable atmosphere ($\zeta < 0$) using the stability functions described in Grachev et al. (1998).

3 Results

3.1. Transfer coefficients over lakes

Bulk transfer coefficients for neutral atmospheric stability C_{DN}, C_{HN} and C_{EN} (Eq. 4a-4c) were estimated using data from 23 lakes and 8 reservoirs (see data availability details in Table in the data repository 10.5281/zenodo.6597829). The transfer coefficients varied between the water bodies and differed on average by a factor of 2-3 for wind speeds exceeding 3 m s^{-1}. However, we identified three water bodies for which the estimated drag coefficients (C_{DN}) were exceptionally large at all wind speeds (up to a factor of five, Lake Quinghai, China, Nam Theun 2 Reservoir, Laos), or exceptionally low (factor of four, Bol'shoi Vilyui Lake, Russia), when compared to other water bodies with similar surface area. These three water bodies contributed largely to the variability among systems (Figure S4a shows the result without these three sites, Figure S5a shows the estimates for individual water bodies). Similarly, Dalton numbers (C_{EN}) estimated from data measured at Lake Lunz (Austria) were a factor of three higher than for other water bodies (Figure S5c). Stanton numbers (C_{HN}) calculated from the dataset collected at Itaipu Reservoir (Brazil) were a factor of four lower than other estimates. Most (90%) of the C_{HN} estimates were removed by filtering for low flux values at this reservoir. With only one of the peculiar data sets for each of the two transfer coefficients, they did not affect the overall statistics for C_{HN} and C_{EN}. We did not find possible sources of errors and considered these data as outliers. In the overall estimates and in the range of variability shown in Figure 2a, we included the complete dataset.

All transfer coefficients showed a similar wind speed dependence (Figure 2a,b,c). At high wind speeds (> 3 m s^{-1}), C_{DN}, C_{HN}, C_{EN} had relatively constant values of $2 \cdot 10^{-3}$, $1.5 \cdot 10^{-3}$, $1.1 \cdot 10^{-3}$, respectively. All transfer coefficients increased towards the lowest wind speeds. The strongest increase was found for C_{DN}, which was one order of magnitude higher ($2.3 \cdot 10^{-2}$) at the lowest wind speed (0.5 m s^{-1}, the first bin) compared to values at higher wind speeds. A similar, but less pronounced increase was observed for C_{HN} and C_{EN}: their values at the lowest wind speed were $6.5 \cdot 10^{-3}$ and $3.2 \cdot 10^{-3}$, respectively. The mean ratio of C_{HN} to C_{EN} is 1.4 and has its maximum value at low wind speeds and a minimum of 1.2 at wind speeds of 3.5-6.5 m s^{-1} (Figure 3).

Unstable atmospheric conditions ($\zeta < 0$) prevailed over all water bodies, particular during the evening and at night time, when $> 80\%$ of all data were obtained under unstable conditions (Figure 2d). Stable atmospheric conditions occurred most frequent during the day (12-19 hours). In addition, we estimated the percentage of time when the wind speed was less than 3 m s^{-1} (Figure 2e). Low wind speed conditions prevailed slightly during the evening and at night, when the atmosphere was mostly unstable. This means that the significant increase of the transfer coefficients at low wind speeds frequently coincides with unstable atmospheric conditions, when the water is still warm and the atmosphere starts cooling at the end of the day.

To analyze the effect of atmospheric stability on the transfer coefficients, we compared the transfer coefficients (C_D, C_H, C_E, Eq. 2a-2c) with their neutral counterparts (C_{DN}, C_{HN}, C_{EN}, Eq. 4a-4c, Figure S6). We found that atmospheric stability did not significantly affect the values of C_D, C_H and C_E at wind speeds exceeding 3 m s^{-1}: their values were in close agreement with C_{DN}, C_{HN} and C_{EN}. However, it is evident that at low wind speeds (0-2 m s^{-1}) these transfer coefficients under in-situ conditions were systematically higher (up to a factor of 2-3) than their neutral counterparts C_{DN}, C_{HN} and C_{EN}.

Estimation of C_H and C_E (Eq. 2b, 2c and Eq. 4b,4c) involves water surface temperature, for which the skin temperature is the most appropriate measure. However, these measurements were not available for some sites. Instead, we used water temperature measured at some depth (often varying between 0 and 0.5 m between datasets). We compared three types of calculations of C_H using two subsets which use: (a) only skin temperature (b) only water temperature and (c) the total dataset which includes both types of temperature measurements

(Figure S4b). C_{HN} estimated with water temperature tends to be slightly lower than the estimates using skin temperature (the percentage difference is approximately 10%). As a result, we presented C_H and C_E (Figure 2b,c) calculated using all available data, independent of how water surface temperature was measured. When both the skin and water temperatures were available for one site, the skin temperature was used.

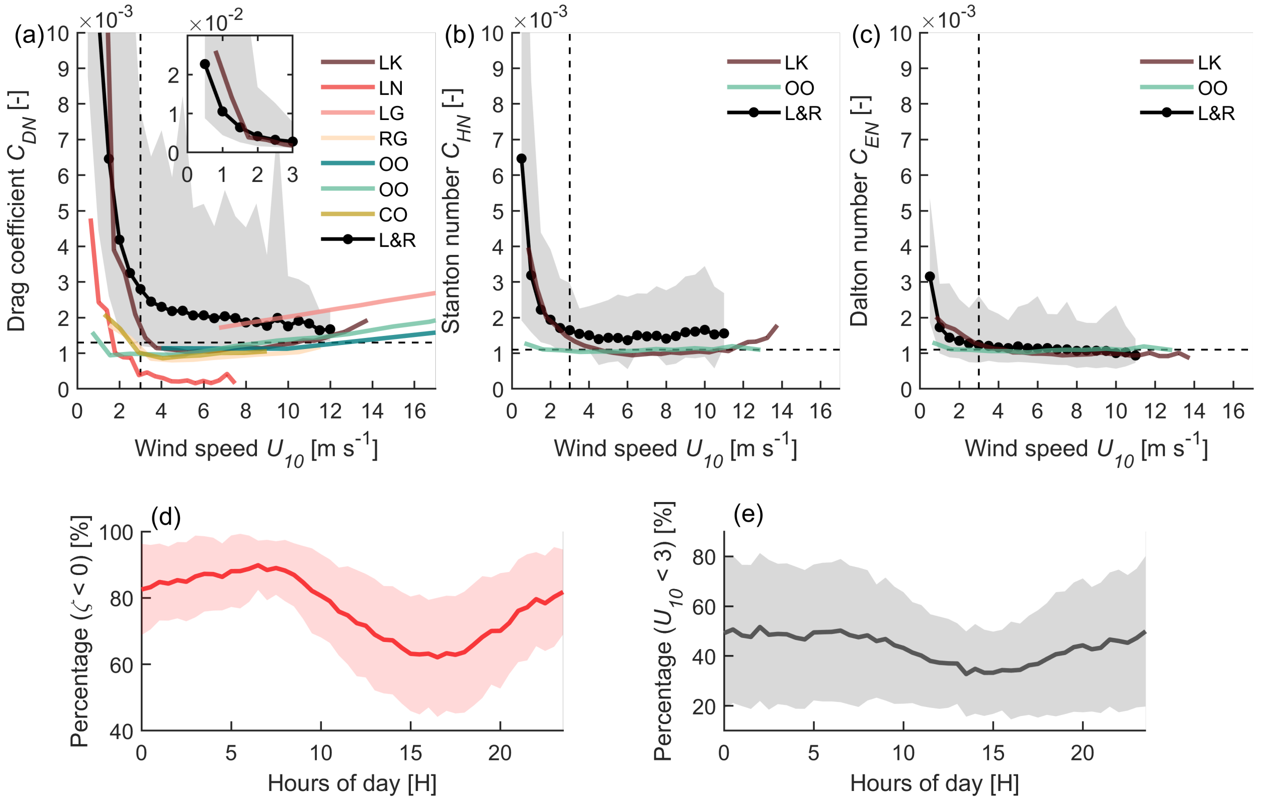

Figure 2. Neutral (a) drag coefficient (C_{DN}), (b) Stanton number (heat transfer coefficient, C_{HN}), (c) Dalton

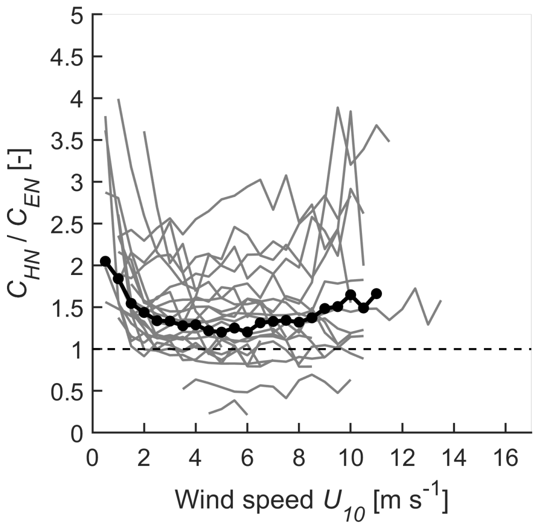

Figure 3. Ratio of bin-averaged C_{HN} to C_{EN} estimated for each individual dataset (21 water bodies, shown l

3.2. Parametrizations of the drag coefficient

We examined the possible mechanisms (Eq. 7, 9, Section 2.4) that could explain the increase of the drag coefficient at low wind speeds and we tested the Charnock relationship (Eq. 11), which describes its wind speed dependence at high wind speeds (Figure 4a). It is evident that our estimates of C_{DN} at wind speeds exceeding 3 m s^{-1} were higher (around factor of two) than the that predicted by the model proposed by Charnock, (1955).

The discrepancy between the EC-derived C_{DN} and theoretical models from the literature showed that neither the concept of smooth flow, nor the consideration of capillary wave roughness could explain the sharp increase of C_{DN} at low wind speeds (Figure 4a). We found that the function describing the wind speed dependence of the drag coefficient proposed by Liu et al. (2020) based on EC measurements over terrestrial surfaces ($C_{\mathrm{DN}} = b_1\,[1 + b_2\exp(b_3U_{10})]$) could successfully describe the relationship over all lakes (Figure 3a). In a similar way, we applied this empirically derived function to C_{HN} and C_{EN} estimates (Figure S7). The fitted coefficients for our data are provided in Table 1.

The concept of gustiness was proven to be relevant for the drag coefficient at low wind speeds at least in the marine environment (Section 2.4.4). We considered this alternative approach to estimate the drag coefficient using one dataset collected in Lake Suwa, Japan, for which scalar averaged wind speeds could be calculated in addition to the commonly provided vector averaged wind speeds (Figure 4b). We calculated the drag coefficient considering gustiness (C_{DNG} , Eq. 14) using the gustiness factor derived from both types of wind speeds (G_{wind}), and from the parametrization using the convective velocity scale (G_{conv}) for unstable atmospheric conditions and used C_{DN} (Eq. 3a) for stable conditions. At wind speeds less than 3 m s^{-1} C_{DN} was on average a factor of 1.3 higher than C_{DNG} (when using G_{wind}). The ratio reached its maximum of 1.9 at the lowest wind speed (0-0.5 m s^{-1}). The ratio of C_{DN} to C_{DNG} estimated using G_{conv} was approximately a factor of 1.2 higher than the one that was estimated using G_{wind}.

Table 1. Coefficients for the empirical function $C = b_1\,[1 + b_2 \exp(b_3 U_{10})]$ (Liu et al., 2020)), describing the w

C_{DN}
C_{HN}
C_{EN}

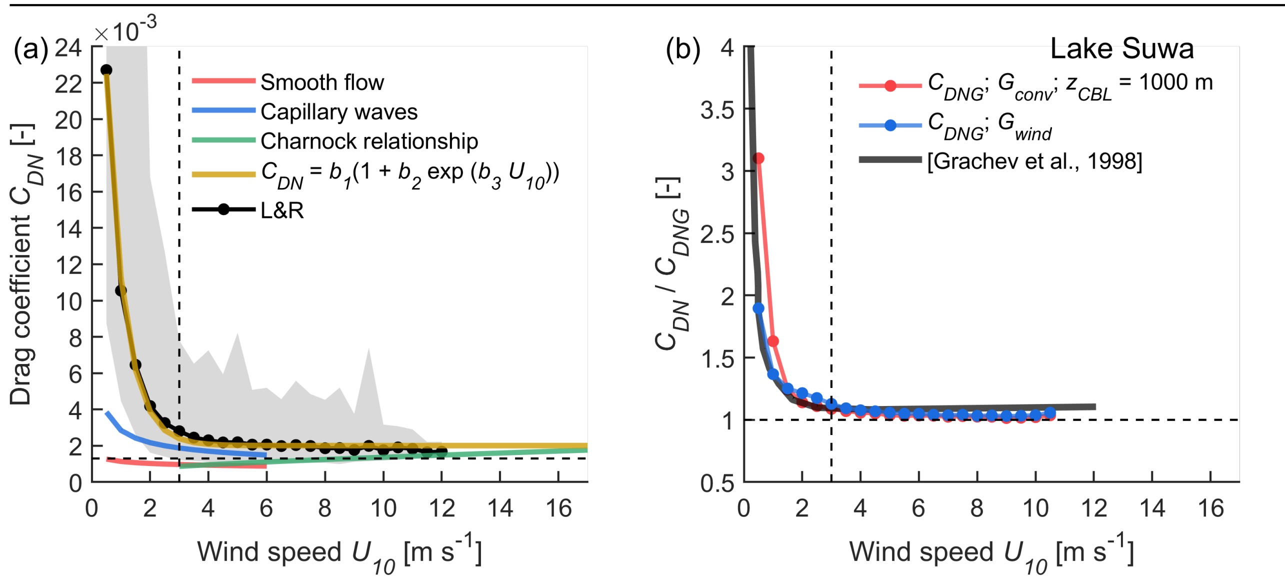

Figure 4. (a) Bin-averaged C_{DN} versus U_{10}. Black line with symbols (L&R): C_{DN} obtained from EC measur

3.3. Dependence of the bulk transfer coefficients on the lake characteristics

We examined the dependencies of the bulk transfer coefficients on lake characteristics, including the maximum and average water depth, water depth at the measurement site, maximum and average wind fetch, and water surface area. As the transfer coefficients at high wind speeds were relatively constant, we first analyzed effects of lake characteristics on the mean values of the transfer coefficients for wind speeds exceeding 3 m s^{-1} estimated for each individual water body.

We found that the mean C_{DN} decreased significantly (Pearson correlation coefficient r = -0.5, p-value < 0.05) with increasing lake surface area and with increasing mean and maximum fetch (Figure 5a, 5b and Figure S8a). These relationships could be expressed as power law dependencies ($y = x^A \exp(\mathrm{B}\ \ln 10)$, where A and B are the slope and intercept of the linear regression $\log_{10} y = A\ \log_{10} x + B$) with exponents of -0.07 and -0.16, -0.12, respectively. Most variability in C_{DN} was found to be explained by the lake surface area (for log-transformed data the coefficient of determination was $\mathrm{R}^2 = 0.3$). The correlation between C_{DN} and mean fetch was slightly higher than for maximum fetch (-0.5 versus -0.4). A principal component analysis revealed that lake surface area has a largest predictive power (Figure S9). We did not find a significant correlation ($r \sim$ -0.2, p-value > 0.05) between the mean C_{HN}, C_{EN} and surface area, mean, and maximum fetch, the stronger correlation could be found when considering data at fixed wind speeds. Significant correlation was found between C_{EN} and the surface area at fixed wind speed of 6 m s^{-1} (Figure 5g). In addition, there was weak negative correlation with mean and maximum fetch ($r \sim$ -0.4, Figure 5h and Figure S8g).

Using the principal component analysis, we identified that there was no significant correlation of the averaged transfer coefficients at high wind speeds with maximum, average or local water depth (Figure 5c,f,i, Figure S8, S9). We used the exponential dependence from Panin et al. (2006) to compare with our results. However, we did not have sufficient sites with larger depth to confirm any dependence.

At low wind speeds ($<$ 3 m s^{-1}), the transfer coefficients were strongly wind speed dependent (Figure 2a,b,c) and their relationships with lake characteristics are examined separately for each different wind speed interval. Here we found that C_{DN} and C_{HN} significantly increased with increasing water surface area for wind speeds between 0.5 m s^{-1} and 2 m s^{-1}. At higher wind speeds these correlations become negative, as in the analysis for wind speed $>$ 3 m s^{-1} presented above. As an example, we show the transfer coefficients for a wind speed of 1 m s^{-1} in Figure S10. Significant correlation ($r \sim$ 0.5) (and its decreasing towards high wind speeds) could also be observed between C_{DN}, C_{DN} and mean and maximum fetch (data not shown). At the same time, we found significant correlations of all three transfer coefficients with measurement height at low wind speeds, which was not present at high wind speeds (Figure S10d-f).

As a final step, we looked at the possible relationship between the averaged wind speed (estimated over entire time series for each individual water body) and surface area. We found a significant correlation between them in a double-logarithmic domain ($r = 0.5$, p-value < 0.05, Figure S11), resulting in a power-law dependence $U_{10} = A_s^{0.05} \exp(0.5\ \ln 10)$.

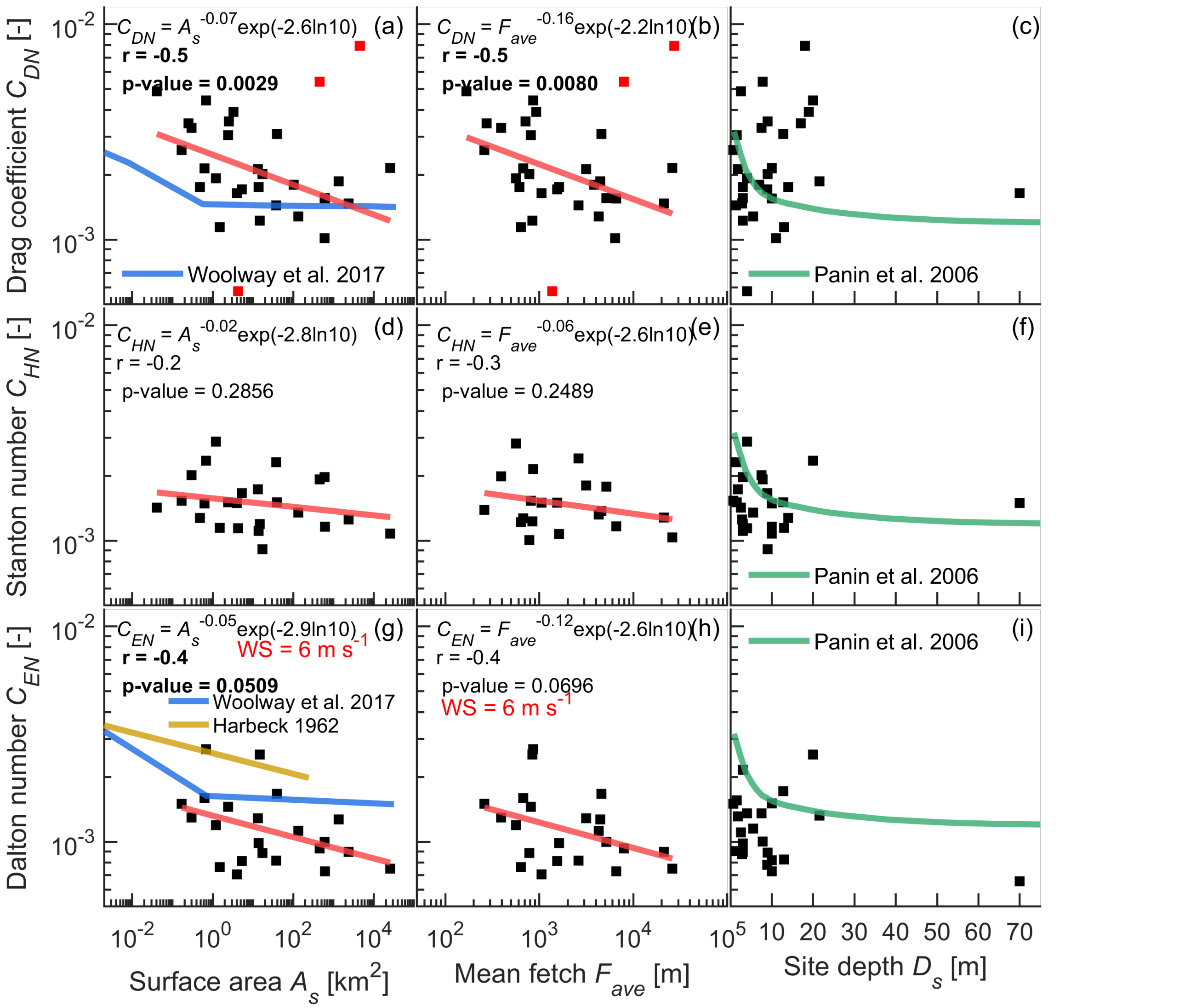

Figure 5. Neutral transfer coefficients (a, b, c) C_{DN}; (d, e, f) C_{HN}; (g, h, i) C_{EN} versus surface area of the w

4 Discussion

4.1. Bulk transfer coefficients estimated for lakes and reservoirs

We examined the bulk transfer coefficients describing the transport of momentum, heat and water vapor at the water surface estimated based on EC data collected at 23 lakes and 8 reservoirs of different size, depth, and location. All transfer coefficients tended to increase towards low wind speeds and remained relatively constant at wind speeds exceeding 3 m s^{-1}. This increase was reported in previous studies for lakes (see, e.g., Wei et al., 2016; Xiao et al., 2013) and has been extensively investigated but has remained unexplained up to now. The lower bound for C_{DN}, C_{HN}, C_{EN} among the water bodies at high wind speeds were within the range reported by previous studies – either for large lakes (> 200 km^2, (Kuznetsova et al., 2016; Wei et al., 2016)) or for the marine environment: classical open ocean measurements (Fairall et al., 2003; Large & Pond, 1981) and coastal ocean sites under fetch-limited conditions (Lin et al., 2002). Indeed, we also considered large lakes (Figure 1b) that were expected to have the smallest drag coefficient as they had the largest fetch (e.g., Lake Erie, Lake Taihu, Lake Balaton). The mean C_{DN} for winds exceeding 3 m s^{-1} was equal to $2 \cdot 10^{-3}$ and this value corresponded to an upper bound for the water surface roughness (0.001 m) reported by Foken (2008), but was a factor of two higher than the values reported for oceans and large lakes or reservoirs (Large & Pond, 1981; Fairall et al., 2003). While C_{HN} and C_{EN} are commonly assumed to be equal, we found that C_{HN} was on average by a factor of 1.4 higher than C_{EN} (averaged over all wind speeds and all water bodies under study). The finding of C_{HN} being higher than C_{EN} confirmed the results reported by, e.g., (Wei et al., 2016; Dias & Vissotto, 2017). The mean value of C_{EN} for high wind speeds ($1.1 \cdot 10^{-3}$) was found to be the same as in (Kantha & Clayson, 2000), but C_{HN} was larger ($1.5 \cdot 10^{-3}$) as in (Harbeck, 1962; Hicks, 1972). The fact that $C_{\mathrm{HN}} > C_{\mathrm{EN}}$ may have significant implications, because it results in biased estimates of lake evaporation based on the energy-budget Bowen ratio method.

Values of C_{DN} varied considerably depending on the type of measurements used for its estimation. For example, in (Simon, 1997) C_{DN} was calculated from the dissipation rate measured at the water side for relatively large Lake Neuchâtel (218 km^2, Switzerland). C_{DN} was significantly lower than our estimates (factor of ten) and the estimates from lakes or marine measurements (factor of five). However, these estimates also confirmed the increase of C_{DN} at low wind speeds. C_{DN} at high wind speeds calculated from the wind profile method at the nearshore site in Lake Geneva (Graf et al., 1984) was in close agreement with our estimates. The strong increase of C_{DN}, C_{HN}, C_{EN} at low wind speeds was similar to the one observed for a large lake with the same EC method of estimation the surface fluxes (Wei et al., 2016), but it was not supported by measurements in the marine environment.

4.2. Bulk transfer coefficients at high winds

The estimated C_{DN} at high wind speeds was higher than predicted by Charnock relationship. This result was expected as Charnock relationship is based on the assumption that the water surface roughness is controlled by fully developed surface gravity waves. This may not be the case for many lakes, where wave generation is fetch-limited (e.g., overview in (Ataktürk & Katsaros, 1999)). We could attribute this difference to the lake surface area and the average and maximum wind fetch at the measurement location. To support this, we found that the C_{DN}, C_{HN} and C_{EN} were highest in small water bodies and decreased with increasing surface area and fetch lengths for wind speeds exceeding 3 m s^{-1}. As approximately half of the water bodies under study are relatively small (surface area < 10 km^2), our data indicated that they contributed disproportional to the higher transfer coefficients. For large lakes, the transfer coefficients at high wind speed tended to be lower and closer to the values reported in previous studies and predicted by Charnock relationship. At these higher wind speed, the surface gravity waves could potentially reach the fully developed state in large water bodies. We found a significant correlation between C_{DN}, C_{EN} and the lake surface area. The resulting C_{EN} dependence on surface area with the power of -0.05 confirmed the findings of previous studies (Harbeck, 1962; Brutsaert & Yeh, 1970). However, the values of C_{EN} in our analysis were approximately a factor of two lower. Our results could not confirm a bilinear decrease of C_{DN} with increasing lake size with a weaker dependence for large lakes, as estimated by Woolway et al. (2017) (Figure 6a). The difference between the relationship of C_{DN} with lake surface area reported in Woolway et al. (2017) could be attributed to fact, that they estimated the transfer coefficients from measurements of mean wind speed by applying the parameterizations of surface roughness for smooth flow (Eq. 5) and Charnock relationship (Eq. 10). Nevertheless, the power dependence for lakes with surface area < 1 km^2 (Woolway et al., 2017) looked similar to the one we observed (power of -0.07) for all lakes and reservoirs, suggesting that it could be generalized to many water bodies. In contrast to the results reported in (Panin et al., 2006), we did not find evidence for the existence of an influence of water depth on the bulk transfer coefficients.

4.3. Bulk transfer coefficients at low winds

Low wind speeds are typical conditions for lakes (Woolway et al., 2018), especially for smaller ones (Figure S10), which are most abundant by number (Downing et al., 2006). The most pronounced increase in bulk transfer coefficients at low wind speed was observed for C_{DN}, which was up to one order of magnitude higher at low wind speeds compared to its value at high wind speeds. We found less pronounced increases of C_{EN} and C_{EN} but their values at low wind speed can be larger up to factor of six and three, respectively, in comparison to their constant values at high winds. Periods with low wind speeds mostly corresponded to periods with unstable atmospheric conditions or enhanced convective transport, which is the most prevailing condition for all studied lakes during the ice-free period (Read et al., 2012; Woolway et al., 2017).

None of the tested approaches, including smooth flow and capillary wave parametrizations, could explain the strong increase for C_{DN} at low wind speeds. While Wei et al., (2016) suggested that the contribution of gusts (different formulation of the C_{DN}) was not significant in their dataset, we found that the increase is partially attributed to the way of the calculation of C_{DN}. Different formulation involving the gustiness factor (G_{wind}) could reduce the values of C_{DN} up to a factor of two at wind speeds of 0.5 m s^{-1}. The two different estimates of the gustiness factor (G_{wind} and G_{conv}) should have given similar results, if the correct height of the convective boundary layer was used. However, our estimates of C_{DNG} using G_{conv} were higher than C_{DNG} using G_{wind}, which we consider as a reference. This may indicate that a fixed CBL height of 1000 m was incorrect and should have smaller values. According to Oke (1987) and Stull (1988) the CBL starts to grow during the daytime, when the ground warms the atmosphere. The observed increase of C_{DN} corresponded to the data during evening and night hours, when unstable atmospheric conditions in the surface layer above water were dominant (Figure 3d, e). While the land starts cooling and the atmosphere becomes stable in regular daily cycle, the thermal internal boundary layer (TIBL), several tens of meters thick (Glazunov & Stepanenko, 2015), may grow above the lakes at this time. The TIBL develops due to the temperature difference between land and water. Above this layer (as well as stable layer above the land), there still could be the residual layer which is left from the CBL during the day. Thus, large convective eddies may entrain this air from the residual layer. Tests with the CBL height of 10 m (not shown) led to much lower values of C_{DNG} (approximately a factor of two) compared to the reference. It remains unclear which CBL height should be used for a correct parameterization of C_{DNG}, if the scalar-averaged wind speed is not available.

The wind speed dependence of the bulk transfer coefficients (especially at low winds), could be well described by an empirical function that was originally proposed for the land surface (Liu et al., 2020). This suggests that similar physical processes control the increase in transfer coefficients at low winds, which are independent of the specific roughness conditions of water surfaces. We suggest that the reason for this increase to some extent was a contribution of large convective eddies or non-local effects as described in Liu et al. (2020) and other closely related studies (Read et al., 2012; Sahlée et al., 2014; Ala-Könni et al., 2021).

Unexpectedly, we found that the transfer coefficients (C_{DN} and C_{HN}) significantly increase with increasing lake surface area, mean and maximum wind fetch for wind speeds less than 2 m s^{-1}. This result is counterintuitive, because at low winds we did not expect a dependence on lake surface area or fetch, as it should only be important for the development of surface waves, which appears to be only around the wind speed of 3 m s^{-1} (Simon, 1997; Guseva et al., 2021). It may be related to the development of the TIBL above lakes but the potential mechanism remains unknown. At the same time, we found the significant positive correlation between the transfer coefficients and the measurement height, which was also unexpected. This finding could be a result of the measurement limitations and it could potentially attribute to the increase of the bulk transfer coefficients at low wind speeds. These issues require a separate detailed investigation.

4.4. Study limitations

The estimated bulk transfer coefficients show large scatter, even after filtering the data. The scatter is particularly high at light winds, i.e., in the first three to four wind speed intervals used for bin-averaging (0.5 – 2 m s^{-1}). It could be associated with limitations of the EC measurements, namely, the validity of the underlying assumptions, including the homogeneity and stationarity of the flow, as well as by increasing random errors. As there are no common thresholds, for example, for the removal of low flux values, and the quality check of the EC results is very specific to each site, these effects may not have been removed completely by data filtering. In our case, applying the filter with low fluxes led to increase in transfer coefficients at low wind speeds (0-0.5 m s^{-1}) up to a factor of 1.6 which was a largest impact on the data among other filters (Text S1). However, the most recent study (Ala-Könni et al., 2021) argue that these thresholds for fluxes are not very important for the data quality filtering. Moreover, our estimates of C_{DN} differ from former results obtained using different types of measurements, such as water-side energy dissipation rates (Figure 3a, e.g., (Simon, 1997)). Thus, the combination of water- and air side measurements could be beneficial for further investigation of the bulk transfer coefficients.

Hwang (2004) suggested that the standard height of 10 m at which the transfer coefficients are reported is inappropriate for analyzing C_{DN} and its dependence on surface roughness under wave conditions. They argue that the only relevant parameter that could serve as a reference height is the wavelength that describes the decay rate of the waves with the distance from the water surface. The adjustment of the transfer coefficients to 10 m height may not be very relevant for lakes and reservoirs and the flux measurements at two different heights should be considered in future measurements. These measurements would additionally provide confirmation for the existence of a constant flux layer, which is another important assumption underlying EC measurements.

4.5. Broader implications

Bulk transfer coefficients are usually applied in numerical models for the atmospheric boundary layer, as well as in hydrodynamic models of lakes and reservoirs. Currently, the global modeling studies focusing on the lake mixing and phytoplankton blooms for climate change predictions use constant coefficients, including C_{DN} (Jöhnk et al., 2008; Read et al., 2014; Woolway & Merchant, 2019; Grant et al., 2021), or consider C_{DN} as a tuning parameter of the models (Stepanenko et al., 2014). Inadequate values of C_{DN} result in biased estimates of the current velocities in lake models (Chen et al., 2020). The increase of the transfer coefficients at low wind speeds observed in our analysis can therefore lead to significant errors, as these conditions are the most prevailing conditions for lakes. The empirical parameterizations of the wind-speed dependence of bulk transfer coefficients provided in Table 1 can potentially be applied in modeling lake-atmosphere interactions and for more accurate estimation of the surface fluxes. The dependence on the lake surface area is more complicated to implement, as we observed contrary dependencies for low and high wind speeds.

We emphasize that the Bowen-ratio method, which is frequently used to estimate evaporation may be biased given our finding that $C_{\mathrm{HN}} > C_{\mathrm{EN}}$. This finding violates the assumption of their equality in the Bowen ratio energy budget and related methods, and implies larger (smaller) sensible heat (latent heat) fluxes than those predicted under that assumption. Both the physical mechanisms underlying their difference and the extent of the differences in the predicted sensible and latent heat fluxes require further investigation.

In state-of-the-art weather and earth system models, lakes are included as separate tiles in the model cells, where the surface fluxes over the tiles are computed via Monin-Obukhov similarity scaling. The models provide constant meteorological variables for each grid cell, which is a so-called blending height concept (von Salzen et al., 1996). To use the bulk transfer coefficients derived in this study to compute fluxes, specific values of wind, temperature and humidity over lakes should be used, which can be obtained in generalization of the tile approach, involving the parameterization of internal boundary layers over contrasting surfaces (Arola, 1999; Molod et al., 2003; de Vrese et al., 2016). MacKay (2019) presents a specific example of such an approach developed for lakes and wind speed only. Our results demonstrate a good potential of wind-gust concept to explain the observed increase of bulk exchange coefficients under low winds. However, direct incorporation of this concept in current weather models is not feasible, as these models normally do not predict scalar-averaged wind speed. A possibility to alleviate this obstacle is to replace the scalar-averaged wind speed by another measure of wind speed variability, e.g., the square root of the horizontal turbulent kinetic energy component (Castelli et al., 2005; Esau et al., 2018), or to use a parameterization of the convective velocity scale. However, the latter requires knowledge of the convective boundary layer height above the lake, which is not well understood.

5 Conclusions

We were the first to analyze the bulk transfer coefficients of momentum, sensible and latent heat from the directly measured surface fluxes above various lakes and reservoirs. We observed a pronounced increase of the transfer coefficients at low wind speeds (< 3 m s^{-1}) and relatively constant values at high wind speeds (> 3 m s^{-1}). At high wind speed, the estimated transfer coefficients generally agreed with the results provided by previous studies for large lakes and oceans, yet the Stanton number was systematically higher than the Dalton number by a factor of 1.4, which has implications for the Bowen ratio method. At high wind speed, the drag coefficient and the Dalton number decreased with increasing surface area of the water body and with increasing fetch length, whereas the opposite was found at low wind speed. The strong increase in the transfer coefficient at low wind speed could not be explained by known mechanisms, including smooth flow and capillary waves. However, it can be partly explained by the existence of gusts under unstable atmospheric conditions, and potentially by additional non-local effects. The bulk transfer coefficients at all wind speeds were well described by an empirical function that has been proposed for the land surface. Using this function could potentially improve the accuracy of the bulk parametrization of surface fluxes in numerical models for lake hydrodynamics and atmospheric dynamics. We underline the need for simultaneous measurements of waterside and airside turbulent fluxes in future investigations, as well as experimental confirmation of the validity of the assumptions underlying eddy-covariance flux measurements at low wind speed.

Acknowledgments

S. Guseva and A. Lorke were financially supported by the German Research Foundation (Deutsche Forschungsgemeinschaft, DFG, grant LO1150/12-1). N. Dias and F. Armani were supported by research project FUNPAR 2882, with funding provided by CHESF (São Francisco Hydroelectric Company). A.R. Desai acknowledges support from the DOE Ameriflux Network Management Project and the NSF North Temperate Lakes LTER (#DEB-2025982). Funding for the AmeriFlux data portal was provided by the U.S. Department of Energy Office of Science. H. Iwata was supported by the Japan Society for the Promotion of Science KAKENHI grant number 17H05039 and 21H02315. J. Jansen was supported by the Swedish Research Council under VR grant number 2020-06460. G. Lükő was supported by the ÚNKP-21-3 New National Excellence Program of the Ministry for Innovation and Technology, Hungary. Data processing for Bol'shoi Vilyui Lake made by I. Repina was partially supported by a grant of the Russian Science Foundation 21-17-00249. T. Sachs was funded by a Helmholtz Young Investigators Grant (VH-NG-821) and infrastructure at Lake Dagow was operated as part of the Terrestrial Environmental Observatories Network (TERENO) of the Helmholtz Association of German Research Centers. K. Scholz was supported by the Austrian Academy of Sciences (ÖAW) as part of the project "Influence of climate extremes on C cycling dynamics across the boundaries of aquatic ecosystems (EXCARB)" and by the Autonome Provinz Bozen-Südtirol (ALCH4 Project) – both grants received by Prof. Georg Wohlfahrt (University of Innsbruck). The measurements in Bautzen Reservoir and U. Spank were supported by the German Science Foundation in frame of the projects TREGATA (project number 288267759) and MEDIWA (project number 445326344). V. Stepanenko was supported by Russian Ministry of Science and Higher Education (agreement No. 075-15-2019-1621). P. Torma was supported by the by the National Research, Development and Innovation Office (grant numbers K134559 and K138176). I. Mammarella and T. Vesala acknowledge funding from ICOS-Finland (University of Helsinki). We thank Christian Wille, Sarah Waldo and Werner Eugster for providing the data. The authors declare no conflicts of interest.

Open Research

Derived quantities such as the bulk transfer coefficients are available in the open data repository Zenodo 10.5281/zenodo.6597829. Original datasets are available from the open data repositories or from the co-authors (see details in Table S1).

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