Summertime evaporation over two lakes in the Schirmacher oasis, East Antarctica

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Abstract. The evaporation over two glacial lakes in the Schirmacher Oasis (70 °S, Antarctica) was measured by the eddy-covariance (EC) technique during the austral summers (December – February) in 2017 – 2018 and 2019 – 2020. These direct measurements showed that summertime evaporation over two lakes varied from 0.3 to 5.0 mm d^{-1}. Depending on the ice cover presence, the average evaporation varied from 1.5 ± 0.1 mm d^{-1} in December to 3.0 ± 0.2 mm d^{-1} in January – February. In summer, both lakes were warmer than the ambient air, and the largest day-to-day variations in evaporation were associated with variations in the wind speed. The EC measurements were used as a reference for evaluating the uncertainties of the combination formulas and bulk-aerodynamic method. The bulk-aerodynamic method shows the most accurate estimates of lake evaporation (of 6 – 8 %). The combination formulas underestimated the evaporation by 27 – 73 %.

1 Introduction

Antarctica hosts numerous water bodies that store meltwater within a complex hydrological system of lakes and streams. Every summer, approximately 60,000 supraglacial lakes are formed on the ice shelf (Corr et al., 2022; Dirscherl et al., 2021), the largest lakes found in East Antarctica are up to 80 km long (Shen et al., 2025; Stokes et al., 2020). These lakes accelerate the marginal glacier calving (contributing to the global sea level rise), and increase risk of hydrofracture-induced collapse of ice shelves (Rignot et al., 2004; Banwell et al., 2013; Arthur et al., 2022). Providing a moisture source to the atmosphere through evaporation, lakes affect local weather by increasing the precipitation and fostering fog formation (Blanken et al., 2003; Gultepe et al., 2003; Rouse et al., 2005; Su et al., 2020).

Antarctica hosts approximately 120 permanent settlements in Antarctica, most of them located along the continental coast, in rock oases and sub-Antarctic islands (Antarctic station catalog, 2025). Since the COVID-19 pandemic, the number of people visiting Antarctica has increased by 40%, and is expected to grow further by approximately 12.5% per year (Bastmeijer et al., 2023). The nearest lakes provide a freshwater supply for the settlements, and evaluation of their water resources is a crucial task for their managers (Kaup, 2004; Sokratova, 2011). The managers plan on the maintenance and infrastructure investments based on assessment of freshwater resources and hydrological risks (Lan et al., 2025). This assessment requires reliable hydrometeorological observations and robust models to evaluate hydrological stability. Furthermore, the Antarctic lakes house unique lifeforms (Rothschild and Mancinell, 2004; Andersen et al., 2011; Keskitalo et al., 2013). Understanding the lake hydrological regime is important to save these ecosystems (Faucher et al., 2019).

The lake water balance equation describes changes in a lake volume as the difference of inflow (surface/underground runoff, precipitation, etc) and outflow components (evaporation/sublimation, artificial water withdrawal, surface/underground runoff, etc). Depending on the specific case and the integration time scale—whether daily, seasonal, or decadal—the contribution of certain components may be negligible, while others remain essential (Chebotarev, 1975). Evaporation, in particular, is a crucial component of the water balance for lakes in polar regions (Le et al., 2016; Leppäranta et al., 2020; Wang et al., 2020; Shi et al., 2024). However, because evaporation is difficult to measure directly, it is often estimated using indirect methods that require only limited hydrometeorological observations (Morton, 1990; Finch and Hall, 2001; Spence et al., 2003).

A wide diversity of approaches exists to estimate lake evaporation indirectly from meteorological and hydrological observations (Spank et al., 2025; Keijman, 1974). The indirect approaches are divided into mass and energy balance, bulk aerodynamic (mass-transfer), combination, equilibrium temperature and empirical formulas, pans or sticks and tracers i.e. water isotopes (Finch and Hall, 2001; Finch and Calver, 2008; Abtew and Melesse, 2013; Bellagamba et al., 2021). The use of evaporation pans is a traditional approach that can provide high accuracy when properly calibrated (Stanhill, 2002). In the energy balance and mass (water) balance approaches, the evaporation is typically estimated together with discrepancies of a balance equation provided all other components are known (Singh and Xu, 1997; Stannard and Rosenberry, 1991; Abtew and Melesse, 2013).

The bulk-aerodynamic approach estimated the evaporation on the basis of data on a land surface properties (whether a land surface type is an ice or a lake or rock or a forest, surface temperature and surface roughness) and atmospheric variables (wind speed, specific humidity and air temperature) in the lowermost part of the atmospheric boundary layer (Brutsaert, 1982). It needs meteorological observations on the different levels above the land surface allowing it to calculate heat fluxes (sensible and latent) from their gradients. The latent heat flux divided by the latent heat of vaporization yields evaporation.

The bulk aerodynamic approach employs the transfer coefficients (turbulence, heat and moisture) derived from flux-gradient relationships and semi-empirical profiles for wind, temperature and humidity (Yang and Bai, 2023); in particular situations, it adequately estimates the surface evaporation on daily or shorter timescales (Moore, 1983). Consequently, this method has been widely applied to assess evaporation and sublimation over lakes and glaciers in Antarctica (Clow et al., 1988; Bliss et al., 2011; Leppäranta et al., 2016).

Combination approaches integrate elements of both energy budget and mass-transfer principles (Penman, 1948). A more comprehensive form is the Penman-Monteith equation (Monteith, 1965), which was originally developed to describe evapotranspiration from vegetation. Simplified versions of the combination formulas have been applied to Antarctica lakes even though their uncertainties are unknown (Borghini et al., 2013; Shevnina and Kourzeneva, 2017; Dhote et al., 2021). In the empirical approach, daily and monthly lake evaporation rate is correlated with variables such as surface area, incoming solar radiation, diurnal minimum and maximum of air temperature (Zhao et al., 2013, Spank et al., 2025), or it may solely only the air temperature and relative humidity (Konstantinov, 1968). The disadvantage of the empirical, mass transfer and combination approaches is that their performance is restricted to the specific geographical locations where their empirical coefficients were originally derived (Finch and Hall, 2001). To overcome this, the mass transfer coefficients are often calibrated based on measurements from eddy covariance (EC) systems (Ala-Könni et al., 2022, Nordbo et al. 2011).

This study evaluated the evaporation over two lakes in Antarctica from direct (eddy-covariance) and indirect (mass-transfer and combination) methods. The study quantifies the uncertainties of five combination equations and bulk-aerodynamic formulae against the eddy-covariance measurement collected on two lakes in the Schirmacher Oasis in two austral summers in 2017–2018 and 2019–2020.

2 Study area

The Schirmacher oasis (70°45′ S, 11°38′ E) is situated approximately 80 km from the coast of Lazarev Sea, Dronning Maud Land (SA, Fig. 1 a). The oasis is an ice-free rock outcrop approximately 20 km long and 3 km wide (Fig. 1 a), elongated in the west–northwest to east–north-east direction (Simonov and Fedotov, 1964). The relief is hillrocks with absolute heights up to 228 m above sea level (asl). The rock hills mostly consist of gneiss mixed with basalts (Sengupta, 1991). The Oasis was discovered in 1939, and first settlements emerged primarily for science in the late 1960s. Now, the oasis now shelters two scientific observatories operated year round and two tourist camps occupied seasonally (Fig. 1b). The observatories are crewed by 25–30 overwintering personnel, and up to 50 personnel are visiting the site during summers. Two tourist camps accommodated up to 20 people, and operated from November to February. Two ice runways support transportation of people and cargo. Fuel and goods are mostly delivered by ships to coastal bases located on the ice shelf, and then transported to the settlements by vehicles. The settlements, ice runways and coastal bases are connected with year-round ice roads (yellow lines in Fig. 1 b). In summertime, the transportation along the ice roads is hindered by meltwater lakes and streams formed over the ice. Figure 1 c shows the melted ice road to the White Desert Camp in December 2019.

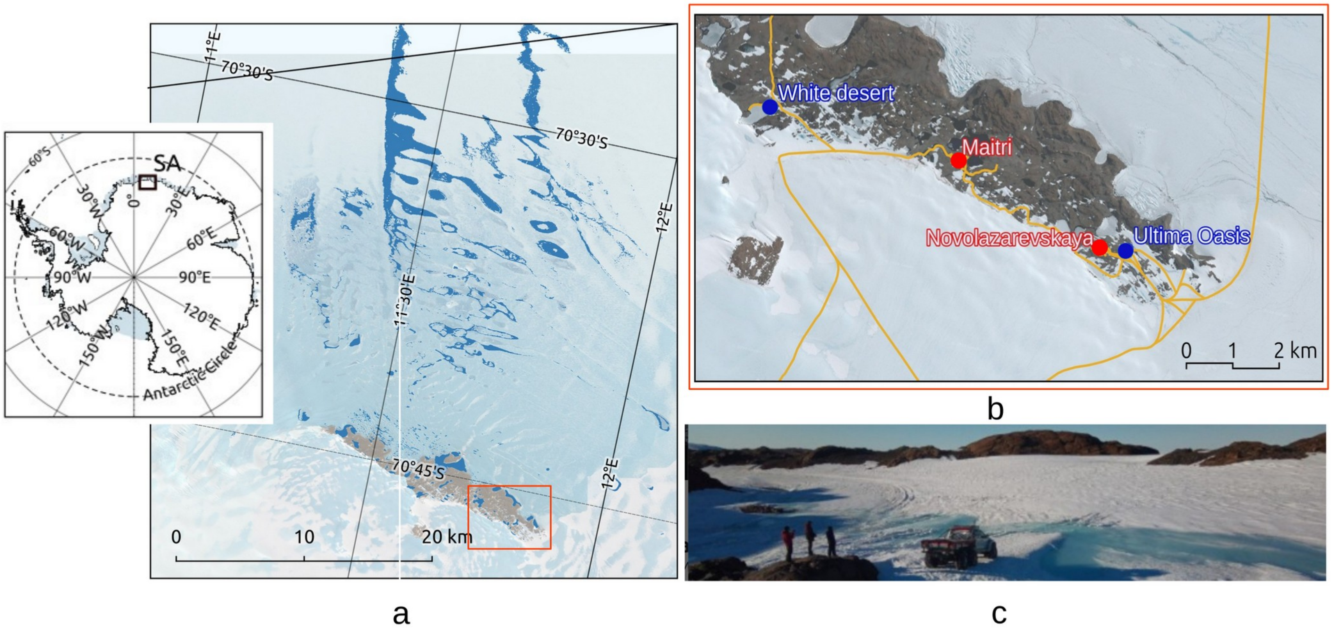


Fig. 1. Location of the Schirmacher oasis (SA) (a) and its social infrastructure (b): year round (red dots) and seasonal (blue dots) settlements connected by roads (yellow lines, © Humanitarian OpenStreetMap Team (HOT), 2020. Distributed under the Open Data Commons Open Database License ((ODbL) v1.0.). © Google Maps, 2019. The red boxes in (a) and (b) outline the area with the main infrastructure in SA, and the image (c) shows the melted ice road to the White Desert Camp (photo by D. Emelyanov, December 2019).

Lakes have existed since the Late Quaternary in the Schirmacher Oasis (Phartiyal et al., 2011), and today, the oasis is home to approximately 300 lakes (Gerrish et al., 2008). Most of these are freshwater waterbodies, nourished by ice melt and seasonal snow cover, and remain ice-free for 6 – 12 weeks in summer (Simonov, 1971). The largest lakes are attached to the ice sheet (glacial), and are connected by temporary streams into chains that end up in epishelf saline lakes (Loopman et al., 1988). Lakes with no connection to ice are landlocked: they are fed by only snow melt, and evaporate through time forming hollows and dry beds (Fedorova et al., 2011). This study focuses on two lakes differing in volume and depth, and both becoming ice free almost every summer for 2–8 weeks (Khare et al., 2008; Tolstikov et al., 2018).

Lake Zub/Priyadarshini is the second largest lake in the Oasis with a volume of 1.02 x10^6 m^3. Its maximum depth is 6.0 m (mean depth is 2.9 m), and a surface area of 35,000 m^2 (Dhote et al., 2021). The lake was ice free from 29–31 December 2017 to 8–12 February 2018. Lake Zub/Priyadarshini was reported to be thermally homogeneous during the period spanning from mid-January 1996 to mid-February 1997 (Sinha and Chatterjee, 2000). Lake Glubokoe is the deepest lake in the Oasis with a maximum depth of 34.5 m (mean of depth is 13.1 m), a surface area of 147,000 m^2, and a volume of over 1.93 x10^6 m^3 (Loopman et al., 1988), Until the late 1990s, it was ice-covered year round (Kaup, 2005). This lake was free of the ice from 26 – 28 January 2019 to 23 – 25 February 2020.

The climate of the Schirmacher Oasis is characterized by low air humidity and temperature, and persistent (katabatic) wind blowing most of the year. There are two scientific observatories operating in the Schirmacher oasis (Fig. 1 a): Novolazarevskaya (Novo) observatory (70° 46′36″S, 11° 49′21″ E) starts the observations in 1961, and Maitri observatory (70° 46′00″S, 11° 43′53″ E) operates since 1989. They are located at a distance approx. 5.5 km, their meteorological measurements following the WMO’s standards (Turner and Pendlebury 2004). Table 1 shows the monthly air temperature, relative humidity, averaged over the period of 1961–2010 according to the observations at Novo site (the data given by the Arctic and Antarctic Research Institute at http://www.aari.aq/default_ru.html, last access 07.12.2021).

Table 1. The monthly minimum, mean and maximum values for the meteorological parameters calculated for the period of 1961 – 2010 (the values are separated by a slash).

Meteorological parameter	1961 – 2010		
	December	January	February
Air temperature, °C	–3.9 / –1.0 /1.5	–2.5 / –0.4 / 1.4	–4.7 / –3.3 / –1.0
Relative Humidity,%	47 / 56 / 69	49 / 56 /66	41 / 49 / 59
Atmospheric pressure, Pa	965 / 975 / 991	964 / 976 / 986	964/ 973 / 987
Wind speed, ms^{-1}	4.3 / 7.4 / 10.3	3.1 / 7.0 / 10.4	5.8 / 9.4 / 13.1
Soil surface temperature, ºC	3.0 / 6.7 / 10.0	3.0 / 6.7 /11.0	–2.0 / 0.2 / 4.0
Precipitation, mm	0.0 / 5.3 / 54.8	0.0 / 2.6 / 38.0	0.0 / 2.9 / 25.9

The region of the study is featured with the persistent katabatic winds blowing from the continental interior (Bormann and Fritzsche, 1995). To fetch an area covering eddy-flux measurements, it is important to place an instrumentation toward prevailing wind directions (Burba, 2013). Fig. 2 shows the wind directions and wind speed anomalies for December and January for the period 1998–2016 calculated from the 6-hour synoptic observations at the Novo site available from the British Antarctic Survey Dataset (https://www.bas.ac.uk, last accessed 14.12.2018). The prevailing wind directions range from 110 to 140º (marked with the black arrows in Fig. 2), and these winds come from a direction that would be the direction of katabatic winds. However, it is not guaranteed that all these winds are entirely of katabatic origin, and some winds may be driven by a combined effect of katabatic and synoptic forcing. The wind speed anomalies were estimated for each 10 degree range in the wind direction, and they are given with the different colors in the legend to Fig. 2. The positive wind speed anomalies are often observed within the range of the prevailing wind directions (marked with orange, yellow, red, brown and black in the legend of Fig. 2). Therefore one can expect the majority of strong winds from these directions.

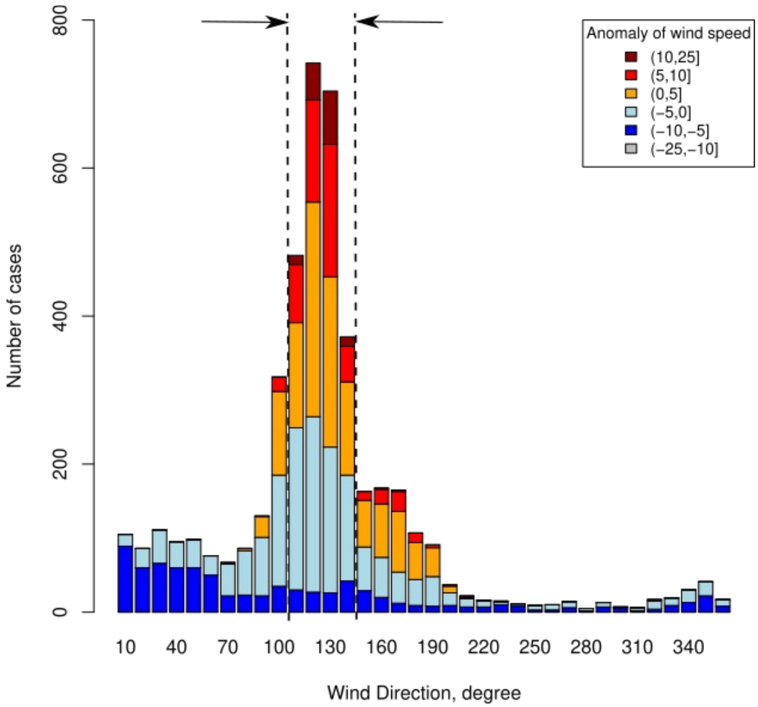


Fig. 2: Wind direction and wind speed anomalies for two austral summer months (December and January).

3 Data and methods

3.1 Micrometeological and hydrological observations

The measurements were collected on two lakes (Fig. 3 a) in two different experiments covering 38 days in 2018 and 33 days in 2019 – 2020. In the first experiment on Lake Zub/Priyadarshini, the evaporation was measured during the period when the lake was free of ice. The second experiment on Lake Glubokoe covers the period of lake ice break-up; the experiment, however, was planned to cover the whole austral summer but was terminated due to the instrumentation failure in mid-January.

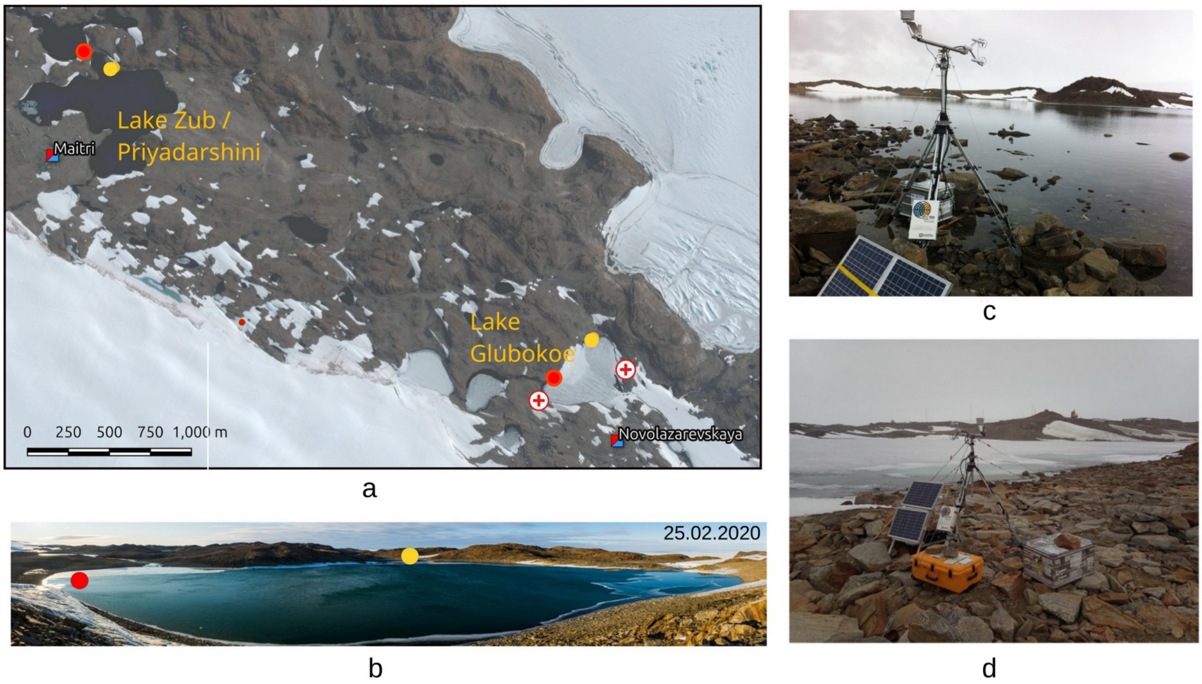


Fig. 3: Location of the instrumentation on the shores of two lakes (a): temperature and water level logger (red dots), the EC system (yellow dots), and digital cameras collecting the panoramic images (rounded red crosses), © Google Maps 2019. The photo (b) shows the instrumentations installed on Lake Glubokoe on the panoramic image from 25 February 2020. The photos on (c) and (d) show the EC system installed on the shore of Lake Zub/Priyadarshini (6 January 2018) and Lake Glubokoe (12 December 2019).

The air temperature, atmospheric pressure, wind speed/direction, and water vapour concentration were measured on a tower equipped with the EC open-path system IRGASON by Campbell Scientific (Table 2). The instrument consists of a 3D sonic anemometer and two gas analysers measuring CO_2/H_2O concentrations on a high frequency (10 Hz). It was placed 5–6 m from the lake shoreline and oriented toward the lake in the direction of the prevailing winds. The lake surface water temperature (LSWT) was measured by the HOBO instrumentation (red dots in Fig. 3 a) installed at a depth of 0.02 m. We used the meteorological measurements from the Novo and Maitri automatic weather stations for a 1–2 day intercalibration with our sensors prior to their installation (data not shown).

Table 2. The instrumentation used in the experiments on Lake Zub/Priyadarshini (LZ) and Lake Glubokoe (LG)

Sensor and site in (Fig. 3)	Elevation, m	Measured variables	Accuracy / (Resolution)	Raw data >> processed data
IRGASON Lake Zub/Priyadarshini Lake Glubokoe	 124 84	air temperature, ºC; H_2O concentration, g/m^3; wind speed, ms^{-1}	±0.15 / (0.025) ±0.037 / (0.00350)	10 Hz >> 30 min
HOBO Lake Zub/Priyadarshini Lake Glubokoe	 122 82	water temperature, ºC	±0.44 / (0.10)	10 min >> 30/60 min and daily

3.2 Methods

3.2.1 Eddy-covariance measurements

The EC technique (Burba, 2013) was used to measure evaporation from the two lakes. The raw measurements were post-processed following Potes et al. (2017). The data were processed to remove spikes (Vickers and Mahrt, 1997), and the spikes detected were counted for quality control. The EC instrumentation collects the gas concentration in 0–360° wind directions, but the only data collected from a sector associated with the footprint over the lake was accounted for. The measurements were filtered by the footprints defined according to Kljun et al. (2004). The air relative humidity and saturation vapor pressure deficit (corresponding to the difference between air and water temperatures) were calculated following Hoeltgebaum et al. (2020) and (Stull, 2017), respectively.

3.2.2 Combination and empirical formulas

The general form of the formulas reads as $E = K\, w_z\, (e_s - e_z)$, where E is the evaporation (mm period^{-1}), w_z is the wind speed (m s^{-1}), $(e_s - e_z)$ is the water vapor deficit (kPa), z refers to height (m), and K is an empirical function approximated with a small number of coefficients. In this study, we evaluated evaporation over two lakes with the formulas originally derived for tropical, temperate and cold climate regions (Table 3).

Table 3. The combination and empirical equations applied to evaluate the lake evaporation. Climate is following the Köppen climate classification.

Author(s) (year)	Formula	Climate
Penman (1948), as given in Tanny et al. (2008)	$E = 0.26(1 + 0.54\, w_2)(e_s - e_2)$	Temperate
Doorenbos and Pruitt (1975)	$E = 0.26(1 + 0.86\, w_2)(e_s - e_2)$	Temperate
Odrova (1979)	$\mathrm{E} = 0.14(1 + 0.72 w_2)(e_s - e_2)$	Cold
Shuttleworth (1993)	$E = 2.909\ \mathrm{A}^{-0.05} w_2\, (e_s - e_2)$	Tropical

In these formulas, E is evaporation (mm d^{-1}), w_2 is the wind speed at 2 m height (ms^{-1}), and A is a lake surface area (m^2). We also verified against the independent observations the formula $E = -0.33(1 - 1.82 w_2)(e_s - e_2)$ suggested in Shevnina et al. (2022).

3.2.2 Bulk-aerodynamic method

Evaporation is defined as the vertical surface flux of water vapor due to atmospheric turbulent transport. Lake evaporation is estimated from measurements of a lake water surface temperature, and relevant bulk properties of the air (Venäläinen et al., 1998; Brutsaert, 1982). We calculated the evaporation after the bulk-aerodynamic methods as follows:

$$E = \rho\, C_{Ez} \left(q_s - q_{az} \right) w_z \qquad (1)$$

where, E is the evaporation (kg m^{-2} s^{-1}); is the air density (kg m^{3}), C_E is the turbulent transfer coefficient for moisture (Dalton number), q_s is the saturation specific humidity corresponding to the lake surface temperature, q_z is the air specific humidity, and w_z is the wind speed. The subscript z refers to the observation height. The actual values for the turbulent transfer coefficient of moisture for a reference level z (C_{EZ}) depend on the surface roughness and atmospheric stratification. The dependence is determined either from field experiments (eddy covariance, gradient measurements and radiation) or from values reported in the literature; both methods were applied in this study. From the literature, the C_E was taken as 0.00107 following Heikinheimo et al. (1999) and Venäläinen et al. (1998), who suggested the value based on measurements over a boreal lake. It allows us to better take into account the different regime of turbulent mixing over a small lake compared to the sea (Sahlee et al., 2014).

Since the atmospheric surface-layer stratification is not always neutral, the stratification effects on the turbulent transfer coefficient C_{EZ} were taken into account as follows:

$$C_{Ez}=\frac{C_{DzN}^{1/2}C_{EzN}^{1/2}}{\left[1-\left(\frac{C_{DzN}^{1/2}}{k}\right)\psi_m\left(\frac{z}{L}\right)\right]\left[1-\left(\frac{C_{EzN}^{1/2}}{k}\psi_q\left(\frac{z}{L}\right)\right)\right]} \qquad (2)$$

where C_{DzN} is the drag coefficient for a lake surface (unitless), k is the von Karman constant (0.4), ψ_m and ψ_m are the stability functions for momentum and moisture, respectively, and L is the Obukhov length (metre). The subscript N in Eq. (2) refers to the neutral values of the transfer coefficients. The Obukhov length (Obukhov, 1946) allows adjusting the transfer coefficients (C_{Dz} and C_{Ez}) to actual stratification in the atmospheric surface layer. In our calculations, the neutral drag coefficient (C_{DN}) equals 0.00181 as suggested by Heikinheimo et al. (1999). The values by Heikinheimo et al. (1999) were given for z = 3 metres, and converted to our observation height (of 2 and 1.8 metres) using Launiainen and Vihma (1990), and the same algorithm was applied to iteratively solve the interdependency of the turbulent fluxes and the Obukhov length. For the stability functions, we used the classic form by Businger et al. (1971) for unstable stratification, and that of Holtslag and de Bruin (1988) for stable stratification. The actual values for the turbulent transfer coefficient (C_E) and the drag coefficient (C_D) were also defined on the basis of the eddy covariance measurements. The wind dependent value for the C_E was defined on the basis of the EC measurements on Zub/Priyadarshini (2017 – 2018), and applied in the calculation of evaporation using Eq. (1) from the independent observations collected on Lake Glubokoe (2019 – 2020).

The C_{Dz} and transfer coefficient (C_{Ez}) were also estimated on the basis of the EC measurements on Lake Glubokoe (2019 – 2020) and applied to calculate the evaporation from the observations on Lake Zub/Priyadarshini (2017 – 2018). Following Arya (1988) the coefficients were calculated as:

$$C_{Dz}=\frac{k^2}{\left[\ln\left(\frac{z}{z_{0m}}\right)\ast\Psi_m\left(\frac{z}{L}\right)\right]^2}. \qquad (3)$$

$$C_{Ez}=\frac{k^2}{\left[\ln\left(\frac{z}{z_{0m}}\right)*\Psi_m\left(\frac{z}{L}\right)\right]*\left[\ln\left(\frac{z}{z_{0q}}\right)*\Psi_q\left(\frac{z}{L}\right)\right]}. \quad (4)$$

where z_{0m} is the momentum roughness length [m] and z_{0q} is the roughness length for water vapour [m] according Ferdorovich et al. (1991):

$$z_{0m}=\frac{0.135\,v_a}{u_*}\ ;\ u_*\le 10.856\left(\frac{cm}{s}\right), \quad (5)$$

$$z_{0m}=\frac{u_*^{\,2}}{69\,g}\ ;\ u_*>10.856\left(\frac{cm}{s}\right), \quad (6)$$

$$z_{0q}=\frac{9.072\times 10^{-2}}{u_*}\ ;\ u_*\le 10.856\left(\frac{cm}{s}\right), \quad (7)$$

$$z_{0q}=8.357\times 10^{-3}-1.904\times 10^{-4}\left(u_*-10.856\right)\ ;\ 10.856<u_*<26.622\left(\frac{cm}{s}\right), \quad (8)$$

$$z_{0q}=1.092\times 10^{-4}\,u_*^{\,2}\,e^{-0.228\,u_*^{3/4}}\ ;\ u_*>10.856\left(\frac{cm}{s}\right). \quad (9)$$

where u_* is the friction velocity (m s^{-1}). The stability functions ψ_m and ψ_m assume the following forms:

$$\Psi_m\left(\frac{z}{L}\right)=\ln\left[\left(\frac{1+x^2}{2}\right)\left(\frac{1+x}{2}\right)^2\right]-2\tan^{-1}x+\frac{\pi}{2}\quad \text{for}\ \left(\frac{z}{L}<0\right), \quad (10)$$

$$\Psi_q\left(\frac{z}{L}\right)=2\ln\left(\frac{1+x^2}{2}\right)\quad \text{for}\ \left(\frac{z}{L}<0\right), \quad (11)$$

$$\Psi_m\left(\frac{z}{L}\right)=\Psi_q\left(\frac{z}{L}\right)=-5\left(\frac{z}{L}\right)\quad \text{for}\ \left(\frac{z}{L}\ge 0\right), \quad (12)$$

where $x=\left[1-16\left(\frac{z}{L}\right)\right]^{\frac{1}{4}}$. The value of the transfer coefficient of moisture (C_E) was further referred to Arya 1988.

Also, we also estimated the transfer coefficient of moisture as suggested in Andreas (1986):

$$C_{Dz}=\frac{k^2}{\ln\left(r/z_0\right)}, \quad (13)$$

$$C_{Ez}=\frac{\alpha_E\,k\,C_{Dz}^{1/2}}{k\,C_{Dz}^{-1/2}-\ln\left(z_Q/z_0\right)}, \quad (14)$$

where, the values of the z_Q and z_0 were taken as the functions of the roughness Reynolds number, and were depended on the wind speed and on the surface roughness parameter.

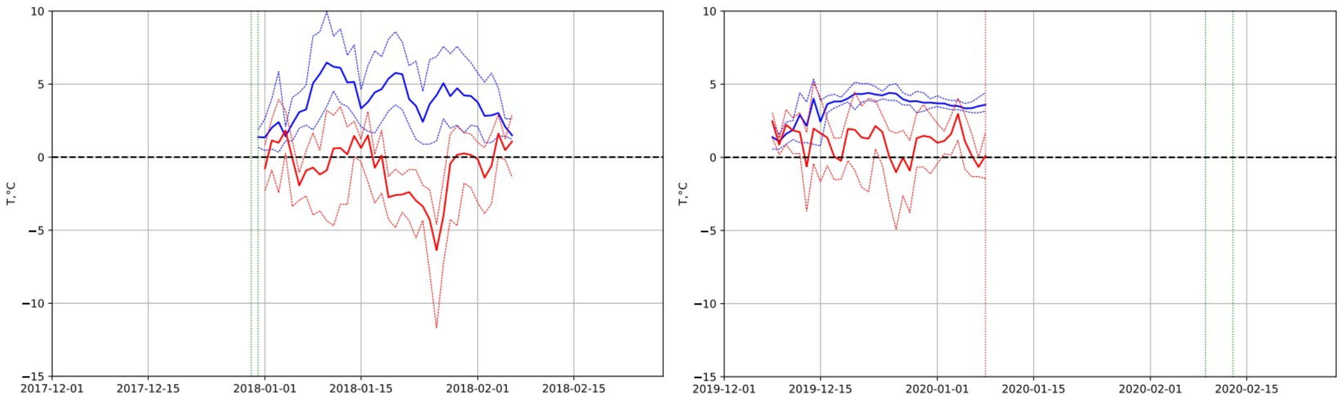
The estimated uncertainties of the indirect methods rely on the eddy-covariance (EC) technique (Tanny et al., 2008; Tran et al., 2023; Shi et al., 2024). To evaluate the skill of the indirect methods we apply the root mean square error ($RMSE = \sqrt{\sum_{1}^{n}\left(E_{EC} - E_{mod}\right)^2}$) following Moriasi et al. (2007). To define if a method is acceptable to be used in hydrological practice or not, we applied the s/σ criteria following to Popov (1979). In s/σ criteria, $s = \sqrt{\sum_{i=1}^{n}\left(E_{EC}^{i} - E_{mod}^{i}\right)^2 / (n-m)}$ and $\sigma = \sqrt{\sum_{i=1}^{n}\left(E_{EC}^{i} - \bar{E}_{EC}\right)^2 / n}$ where, E_{EC} is the evaporation by the eddy covariance method, E_{mod} is evaporation by an indirect method; $\bar{E}$ is the mean evaporation, (mm d^{-1}); n is the length of the series, and m is the number of empirical coefficients in the relationships (equal to 2 in our case). The mean daily evaporation over the observational period and its error ($\sigma_{\bar{E}} = \frac{\sigma}{\sqrt{n}}$) were calculated following Rozhdestvenskiy and Chebotarev (1974).

4 Results

4.1 Weather conditions and lake surface water temperature

In 2017 – 2018, the weather was colder and less windy than the climatology (Table 1): the daily mean air temperatures ranged between –5.6 and 2.4 ºC , and the daily mean wind speed varied between 1.3 to 13.2 m s^{-1}. In January – February 2018, the mean daily wind speed was 6.3 m s^{-1}, and the daily mean relative humidity was 54 %, (varied between 39 and 77 %). The highest relative humidity (up to 90%) was observed on 3 January and 2 December 2018. In 2019 – 2020, the daily mean air temperature ranged from –1.4 ºC to 2.6 ºC, with an average of 1.0 ºC. The daily mean of the air temperature varied between –4.9 ºC and to 5.1 ºC, and the warmest days were observed at the end of December 2019. The daily mean relative humidity was 56%, and it varied between 43% and 77%; and the daily maximums exceeded 90 % on 25–26 December 2019. The daily mean wind speed ranged from 1.3 to 9.5 m s^{-1}, and averaged as 5.9 ms^{-1} for the period of the experiment (Table 2). The strongest winds (weed speed reached up to 13.0 m s^{-1}) were on 9 December 2019 and 1 January 2020, and the weakest winds (wind speed was less than 0.5 m s^{-1}) were on 21–25 December 2019 and 3 January 2020.

During most of the days in both experiments, Lake Zub/Priyadarshinis and Lake Glubokoe were warmer than the ambient air. During the period from 30 December 2018 to 9 February 2019, the mean daily LSWT in Lake Zub/Priyadarshini was 3.9 ºC, which was 4.7 ºC higher than the mean air temperature (Fig. 4 a). The difference between the LSWT and air temperature varied from –0.5 ºC (2–3 January, 2018) to 10 ºC (25–26 January, 2018). From 7 December 2019 to 15 February 2020, the mean daily LSWT of Lake Glubokoe was 3.1 ºC, ranging from 0.6 to 5.3 ºC. During this period, the lake was 2.4 ºC warmer than the air on average (Fig. 4 b). The largest difference between the LWST and air temperature was observed on 25–26 December 2019, when also the relative humidity was the highest, exceeding 90%.



a b

Figure 4. The daily minimum, mean and maximum for the air temperature (red lines) and LSWT (blue lines) measured during the experiments on Lake Zub/Priyadarshini (a) and Lake Glubokoe (b). In (b) the red line shows day 8 in January 2020. The green boxes show the days when the lakes became free of ice.

4.2 Eddy-covariance

Direct measurements of evaporation were a key component of this study. The open-path EC system was installed on a location allowing it to cover the lake surface sector which was between 105 and 240 degrees for Lake Zub/Priyadarshini (green lines in Fig. 5 a and c), and between 90 and 225 degrees for Lake Glubokoe (green lines in Fig. 5 b and d). In Fig. 6 c and d, the X_{90} represents the horizontal distance over the lake surface that encompasses 90% of the flux footprint, and it ranged from 30 to 250 m. The percentage of the filtered data that was excluded did not exceed 20% of total data, and most of the excluded data corresponded to wind directions outside the footprint. These gaps were replaced by the mean of evaporation estimated for the period of the experiment.

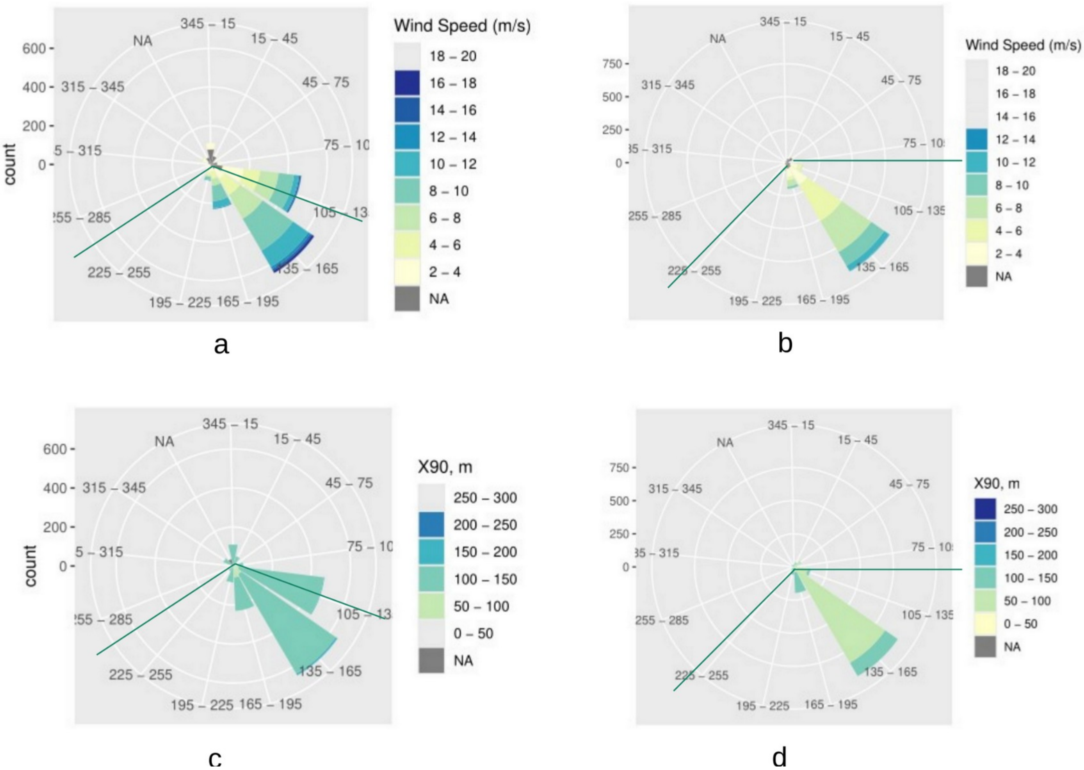
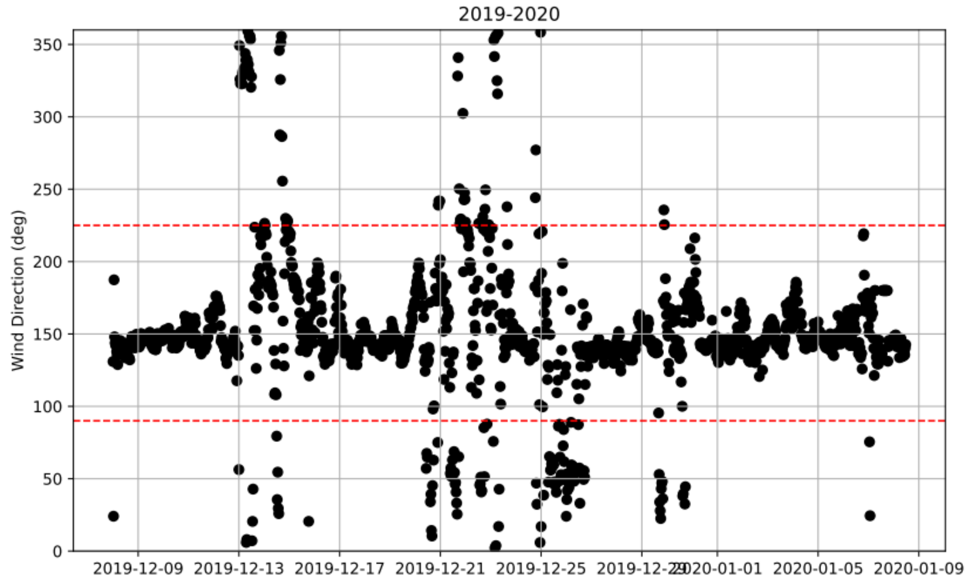
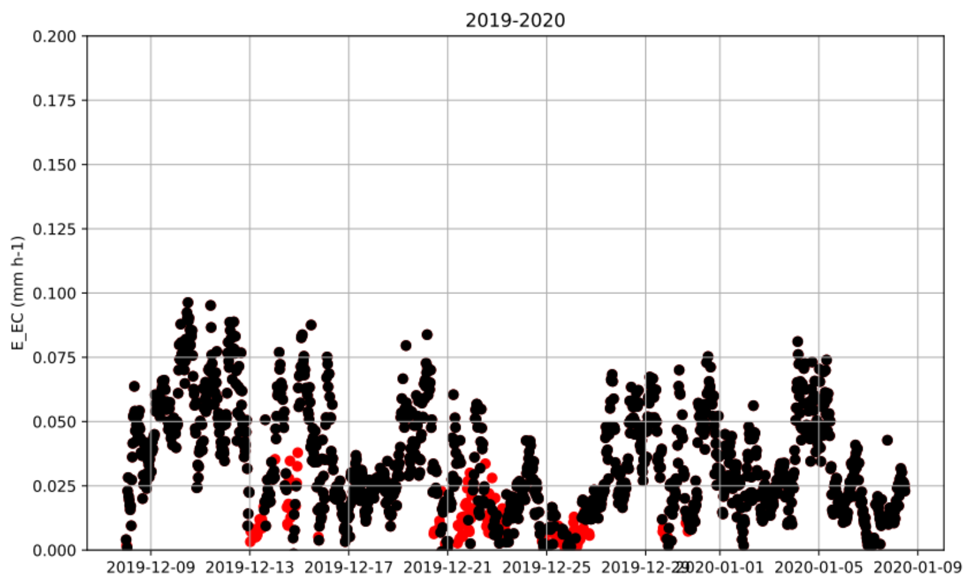


Figure 5. The wind speed and direction measured at Irason site (a: Lake Zub/Priyadarshini, and b: Lake Glubokoe) and the footprint length estimate (X90, m) (c: Lake Zub/Priyadarshini, and d: Lake Glubokoe). The green lines indicate the footprint sector.

The IRGASON height of 1.8 – 2 m generates a blind zone near the instrument, so that the stones on the downwind shore do not affect the fluxes. The low height of the instrument also guarantees that the vertical divergence of the water vapour flux is negligible. The raw 30-minute measurements were filtered by the sensors' signal strengths, a number of gaps in the observations, the footprint and the sector of wind directions covering a lake. The percentage of the filtered out (excluded) measurements did not exceed 20 % of total data, and most of the excluded data corresponded to wind directions outside a lake. The gaps were replaced by the mean of evaporation estimated for the period of the experiment, and the hourly and daily cumulative sum were calculated from the data with gaps filled. Figure 6 shows how the values of evaporation were filtered in the experiment on Lake Glubokoe (2019 – 2020): the red lines outline the sector of wind directions covering the lake surface (a), and the red dots show the filtered values (b).



a



b

Fig. 6. The 30-minute means of the wind direction (a) and evaporation (b) measured by the EC instrumentation on Lake Glubokoe. On (a): the red lines show the limit of the wind directions within the footprint; on (b): the red dots indicate the measurements collected outside the foot.

The daily mean evaporation rate as well as the cumulative sum for the whole period were calculated from high-frequency measurements. For Lake Zub/Priyadarshini, the daily mean evaporation rate varied from 1.5 to 5.0 mm day^{-1} and took an average 3.0 ± 0.3 mm day^{-1}. The average daily evaporation rate was calculated by dividing 114 mm of evaporated water (sum over the whole period of the experiment) by the number of days in the period (which is 38). The evaporation increased with wind speed, with the maximum of 5.0 mm d^{-1} observed during the storm on 2–3 January 2018. For Lake Glubokoe, the daily mean evaporation rate varied between 0.3 and 3.2 mm d^{-1}, and took on average 1.5 ± 0.1 mm d^{-1}. The total sum of the evaporation in this experiment was 54 mm. The largest evaporation (> 2.5 mm d^{-1}) was observed on 9–11 December 2019 and 3–4 January 2020. The smallest evaporation (< 0.9 mm d^{-1}) was observed on 6–7 January 2020.

4.3 Combination and empirical formulas

The evaporation over the was estimated on the basis of four formulas: the mean daily evaporation varied between 0.4 ± 0.1 mm d^{-1} and 1.1± 0.1 mm d^{-1}, which represent 27 – 73% less than the evaporation based on the EC method (Table 4). The formula by Shevnina et al. (2022) gave the daily mean evaporation of over 1.5 ± 0.02 mm d^{-1} over the ice-covered Lake Glubokoe (independent data); it is only 6% less with respect to the evaporation rate calculated from the EC technique. The scores of all formulas (Table 5) are weak: $s/\sigma > 2$, and the RMSE is up to 1.6 mm d^{-1} .

Table 4. The evaporation over two glacial lakes on the basis of the combination formulas and their skill scores. The daily mean evaporation in mm d^{-1} , and the sum is given in mm $period^{-1}$

Author(s) in Table 3	Lake Zub/ Priyadarshini			Lake Glubokoe		
	Mean	Sum	*RMSE / s/σ*	Mean	Sum	*RMSE / s/σ*
Penman (1948)	1.3 ± 0.2	48	1.8 / 2.0	0.6 ± 0.1	20	1.3 / 2.1
Doorenbos and Pruitt (1975)	1.8 ± 0.2	68	1.3 / 1.6	0.8 ± 0.1	28	1.5 / 2.4
Odrova (1979)	0.8 ± 0.1	32	2.2 / 2.4	0.4 ± 0.1	13	1.3 / 2.2
Shuttleworth (1993)	1.0 ± 0.1	38	2.1/ 2.3	1.1 ± 0.1	37	1.3 / 2.1

4.4 Bulk-aerodynamic method

We estimated the lake evaporation on the basis of the observed wind speed, lake surface temperature (controlling q_S), and air specific humidity, applying three literature based formulae for C_{ENz} (Heikinheimo et al., 1999; Arya, 1988; Andreas, 1986). In addition, the actual values of the transfer coefficient were derived from the direct observations on evaporation, wind speed, air specific humidity, and surface saturation specific humidity collected during the experiment on Lake Zub/Priyadarshini (2017 – 2018).

Applying Eq. (1), 30-minute mean values of C_{E2} were solved on the basis of the observed E, q_S, q_2 and w_2, and the results were converted to C_{EN2} . Finally, to allow comparison against relationships presented in the literature for the standard height of 10 m, the C_{EN2} values were converted to C_{EN10} values applying the iterative algorithm of Launiainen and Vihma (1990). The result demonstrated an approximately piecevise linear relationship between w_2 and C_{EN} (Figure 10). Linear regressions were derived for the wind regimes below and above 13 m s^{-1}), separately for C_{EN10} and C_{EN2} as follows:

C_{EN10}= 0.0000119 w_2 + 0.00114 ($w_2 <$ 13 ms^{-1}), (15)

C_{E10}= 0.0013 (if $w_2 >=$ 13 ms^{-1}). (16)

C_{EN2}= 0.0000193 w_2 + 0.00153 (if $w_2 <$ 13 ms^{-1}), (17)

C_{E2}= 0.0018 (if $w_2 >=$ 13 ms^{-1}), (18)

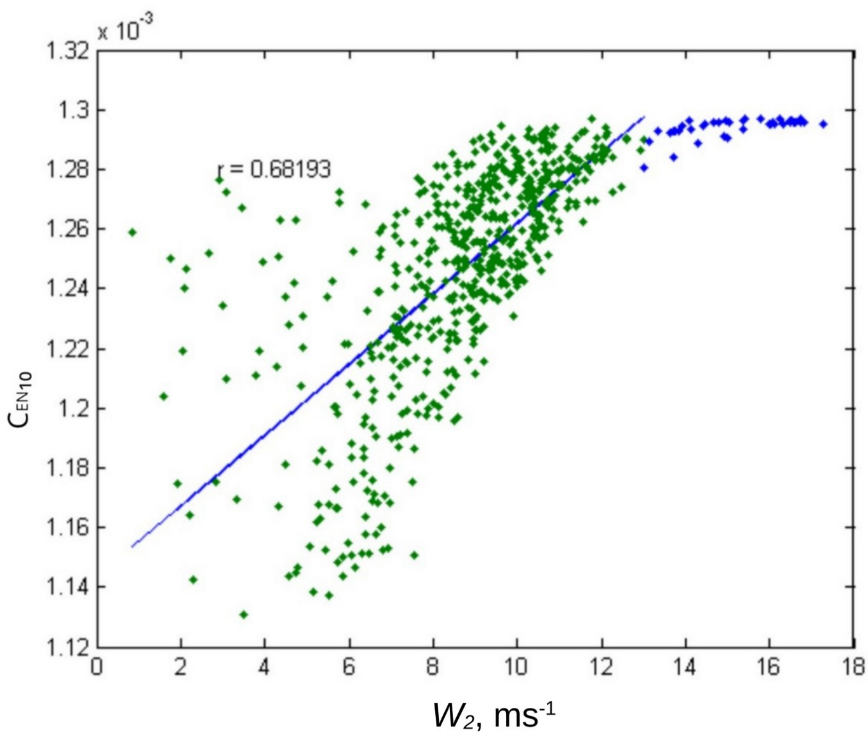


Fig. 7. Dependence of the 10-m neutral transfer coefficient for moisture (C_{EN10}) on the 2-m wind speed (W_2) on the basis of 30-minute mean values observed over Lake Zub/Priyadarshini (2017–2018).

The lake evaporation measured with the EC technique includes the contributions of both interfacial evaporation from the lake surface and evaporation from spray droplets in the air. It is difficult to distinguish between these contributions but, based on observations by Guest (2021) over the Southern Ocean, we may assume that under very strong winds the spray evaporation is some 18 % of the interfacial evaporation. Accordingly, if one wants to apply the wind dependent transfer coefficient (Eqs. 17 and 18) solely for the interfacial evaporation, the contribution of spray evaporation must be subtracted from the results. The stability dependence for the transfer coefficient of moisture was minor. This was because during the period of the experiments, the lakes were continuously warmer than the ambient air (on average by 4.7 C in January – February 2018). Hence, when the wind was strong, the stratification was close to neutral and the transfer coefficient was large and not sensitive to the stratification.

The actual values of the transfer coefficients (C_E and C_D) were estimated from the EC measurement on Lake Glubokoe (2019 – 2020) applying the methods referred to Arya 1988 and Andreas 1986 (Table 2). In the first method (Eqs. 3 - 12), the actual values were estimated for the roughness lengths calculated with literature-based friction velocity (black dots in Fig. 8). In the second method (Eqs. 13–14), the actual values of the transfer coefficients were calculated with independence of stability and friction velocity (red dots in Fig. 8) .

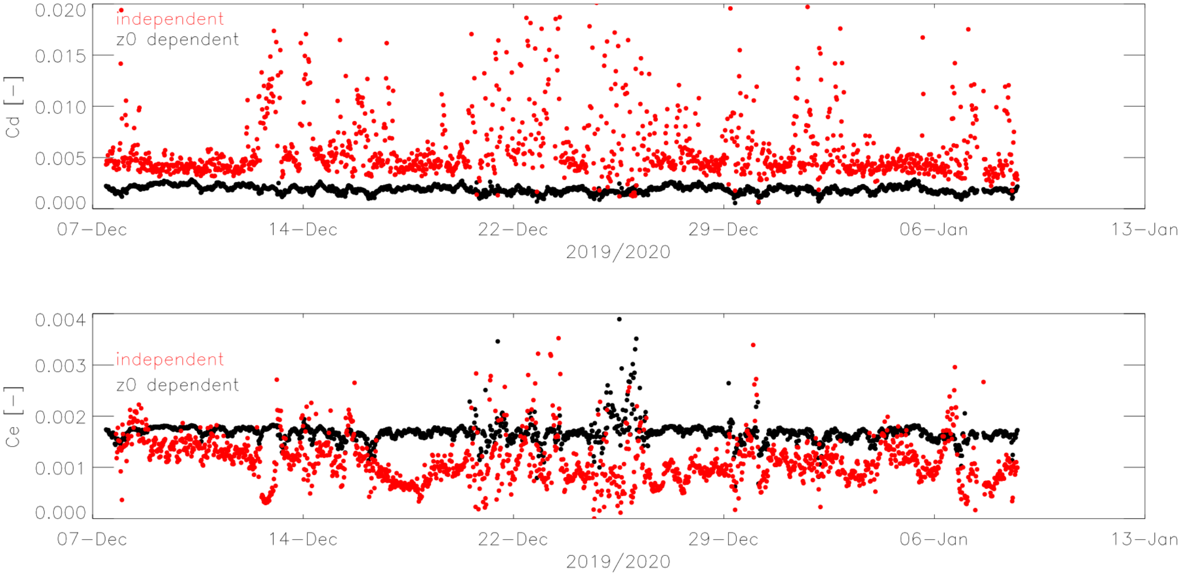


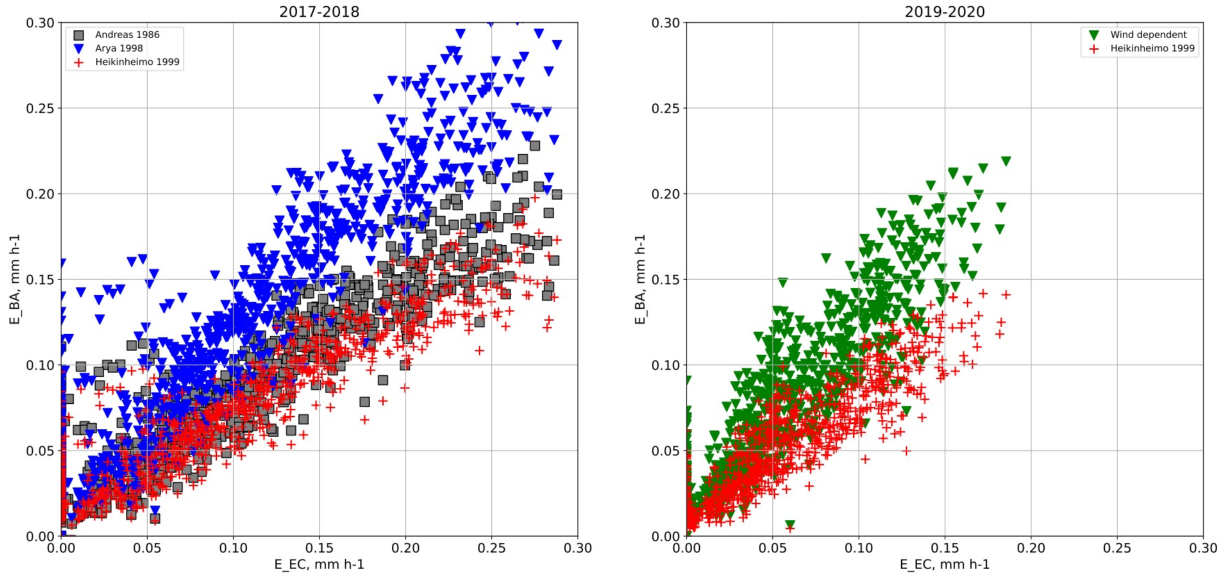
Fig. 8. The actual values of the bulk transfer coefficients calculated following Arya 1988 (black dots) and Andreas 1986 (black dots) estimated from the EC measurement on Lake Glubokoe (2019–2020).

The actual values for the C_{E10} (for the standard height of 10 meters) varied between $1.1\cdot10^{3}$ and $2.1\cdot10^{3}$, in Table 5 they are given for the height of our instrumentation: 2 m for Lake Zub / Priyadarshini and 1.8 m for Lake Glubokoe. The values of the C_{E2} were further used in calculations of lake evaporation with the Eq. (1), and estimations of the scopes for the bulk-aerodynamic method simulating the hourly, daily and seasonally cumulative sums of lake evaporation calculated from the 30 minute data.

Table 5. The equations applied in estimations the transfer coefficient (C_E) in the bulk-aerodynamic method. Notations: N refers to the neutral stratification, and z refers to the height of the instrument.

Lake	Eqs. for the C_E	Methods	$C_{E\ Nz}$	
			z = 10 m	z = [1.8 – 2] m
Lake Zub / Priyadarshini (2017 – 2018) Lake Glubokoe (2019–2020)	taken from literature	Heikinheimo et al., 1999	1.2×10^{-3}	1.1×10^{-3}
Lake Glubokoe (2019–2020)	Eqs. (17 – 18)	Wind dependent	–	–
Lake Zub / Priyadarshini (2017 – 2018)	Eqs. (3 – 12)	Arya, 1988	1.4×10^{-3}	1.2×10^{-3}
	Eqs. (13 – 14)	Andreas, 1986	2.1×10^{-3}	1.7×10^{-3}

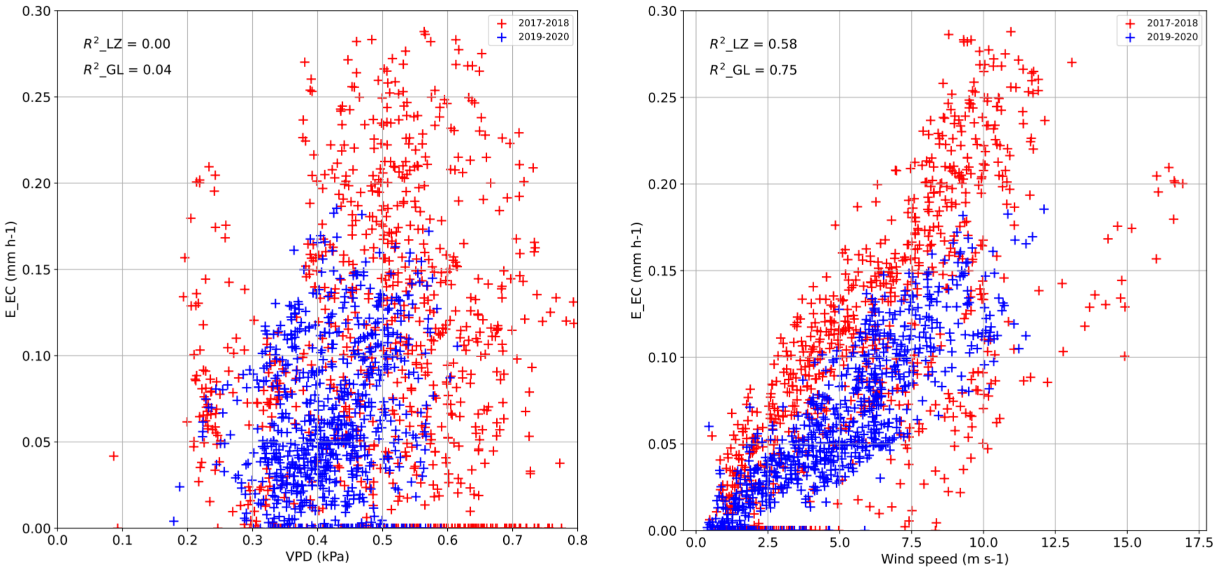
The hourly rate of lake evaporation was simulated with the bulk-aerodynamic method following the different actual values of the E_{C2} (Table 5). Figure 9 shows a good agreement between the bulk-aerodynamic simulations and independent measurements: the Pearson correlation coefficient is 0.92 for the data collected in the experiments on Lake Zub/Priyadarshini, 2017 – 2018 (Fig. 9 a) and 0.89 for Lake Gluboke, 2019 – 2020 (Fig. 9 b).



a b

Fig. 9. The measured (x-axis) evaporation and simulated (y-axis) with the bulk-aerodynamic method applying different actual values for the E_C: (a) Lake Zub/Priyadarshini and (b) Lake Glubokoe. The red dots cross to Heikinheimo et al. (1999), grey box to Andreas (1986), blue triangle to Arya (1988), and green triangle to the wind-dependent coefficient.

There are two factors driving lake evaporation: the vapor pressure deficit (depends on a land surface-air temperature gradient) and wind speed. The results from two glacial lakes suggest no dependency between the lake evaporation and the vapor pressure deficit (Fig. 10 a) with the Pearson correlation coefficient is near zero in both experiments (2017 – 2018 and 2019 – 2020). However, the dependence between the lake evaporation and the wind speed is strong (Fig. 10 b), the Pearson correlation coefficient is 0.58 for the ice-free Lake Zub/Priyadarshini (2017 – 2018) and is 0.75 for the ice covered Lake Glubokoe (2019 – 2020). It implies that the wind speed is the major driver for the evaporation over the local lakes.



a b

Figure 10: The scatterplots of the vapor pressure gradient, (x-axis in a: VPD, kPa) and wind speed (x-axis in b: m s^{-1}) against the hourly rate of lake evaporation (y-axis in a and b) for Lake Zub/Priyadarshini (red) and Lake Glubokoe (blue).”

The daily and seasonal rate of lake evaporation were calculated as a cumulative sum of the hourly rate after the bulk-aerodynamic method (with the various C_{E2}). For the ice free Lake Zub/Priyadarshini (2017–2018), the mean daily evaporation rate varied between 2.0 ± 0.1 mm d^{-1} and 3.0 ± 0.2 mm d^{-1} (Table 6). The best scores for the daily mean evaporation rate gives the C_{E2} = 1.7 x 10^{-3}, and it fits the estimates according to the EC technique (3.0 ± 0.2 mm d^{-1}).

Table 6. The lake evaporation estimated using the bulk-aerodynamic method used different actual values of the transfer coefficient (C_{E2}). The bold numbers indicate the indirect method that yielded the estimates closest to the reference, and the parentheses show the verification with the independent data. Sum is the cumulative value for the period of the experiment (Table 2).

Actual value the C_{E2} in Table 5	Lake Zub/ Priyadarshini (2017 – 2018)			Lake Glubokoe (2019 – 2020)		
	Mean mm d^{-1}	Sum mm	*RMSE / s/σ*	Mean mm d^{-1}	Sum mm	*RMSE / s/σ*
1.1 x 10^{-3}	[2.0 ± 0.2]	[78]	[1.0 / 1.1]	[1.3 ± 0.1]	[42]	[0.5 / 0.8]

Eqs. (17, 18)	2.6 ± 0.2	100	0.7 / 0.7	[1.9 ± 0.2]	[63]	[0.5 / 0.9]
1.2×10^{-3}	[2.1 ± 0.2]	[79]	[0.9 / 0.9]	**1.5 ± 0.1**	**50**	**0.4 / 0.7**
1.7×10^{-3}	**[3.0 ± 0.2]**	**[114]**	**[0.7 / 0.7]**	2.2 ± 0.2	71	0.8 / 1.2

For Lake Glubokoe (2019 – 2020), the best estimate for the evaporation being 1.5 ±0.1 mm d^{-1} (on average) is obtained with the transfer coefficient C_{E2} = 1.2×10^{-3} with the daily lake evaporation of 1.6 ± 0.1 mm d^{-1}; it is approximately 8% less than the reference method. Depending on the value of transfer coefficient (C_E), the bulk-aerodynamic method either underestimated the evaporation up to 20% (1.1×10^{-3}) or overestimated it up to 38% (Eqs. 17). The RMSE varied from 0. 4 to 0.8 mm d^{-1}, and the s/σ criteria was 0.7 when the C_E was 1.2×10^{-3} and 1.7×10^{-3}, and the scores are acceptable for an indirect method (s/σ < 0.8).

It is indeed interesting that the new wind-dependent transfer coefficient, derived on the basis of Lake Zub/Priyadarshini data, has a smaller RMSE for Lake Glubokoe than for Lake Zub/Priyadarshini. This is, however, understandable because the transfer coefficient is only one of the factors that controls the evaporation calculated using the bulk-aerodynamic formula (Eq. (1)). In addition to the transfer coefficient, at least the following factors may generate inaccuracy in the estimated evaporation: (1) errors in the measurements of lake surface temperature (controlling the saturation specific humidity), air specific humidity, and wind speed, (2) contribution of spray droplets to evaporation, (3) role of waves in the lake surface, and (4) validity of the Monin-Obukhov similarity theory, which requires quasi-stationary, horizontally homogeneous conditions. Errors and uncertainties associated with (1) – (4) above may either increase the overall error or partially offset one another. Hence, it is reasonable that the actual value of the transfer coefficient varies among lakes, as factors (1) – (4) may be site-specific, and also depends on the metrics used (e.g., RMSE or s/σ) to evaluate performance (Table 4).

5 Discussion

The EC technique has been recognized as the most precise for directly measuring the evaporation over inland water (Eugster et al., 1997; Spank et al., 2020; Rodrigues et al., 2020), and has taken as a reference to quantify various indirect methods (Wang et al., 2019; Tanny et al., 2008). Its accuracy is, however, sensitive to how a measuring system has been deployed and post-processed (Spank et al, 2025; Burba, 2013). In our experiments, the EC system was deployed to measure fluxes over the lakes on low heights (1.8 – 2m), and the data got low gaps. The system was deployed with accounting the prevailing winds, and during the experiments the instrumentation fetched up to more than 80 % fluxes from the footprint covered by a lake surface. The EC measurements were processed following the standard routine for the open-path systems (filtering, gap filling and footprint). Therefore, we considered the EC measurements given the reference to compare the lake evaporation calculated with the indirect methods.

We estimated evaporation from the EC measurements collected in two lakes in the Schirmacher Oasis. Two field experiments covered the break-up and ice-free periods on the lakes differing in depth, volume and thermal regime. For Lake Zub/Priyadarshini, the direct estimates give the average daily evaporation rate as 3.0 ± 0.2 mm d^{-1} in January – February 2018 when the lake was ice-free. The daily evaporation varied from 1.5 to 5.0 mm day^{-1}; and the cumulative sum of lake evaporation was 114 mm for the period from 1 January to 7 February 2018 (38 days). The evaporation increased with the wind speed, and took the maximum of 5.0 mm d^{-1} observed during storms. The highest evaporation was measured on 2 – 3 January 2018. For Lake Glubokoe, the daily evaporation rate varied between 0.3 and 3.2 mm d^{-1}, with an average of 1.6 ± 0.1 mm d^{-1} during the ice break-up period. The cumulative sum of evaporation is 54 mm of evaporation over the 33-day period in December 2019 – January 2020. The largest evaporation rate (up to 3.0 mm d^{-1}) was observed during the storm 13–15 December 2019. The EC measurements imply the lakes start evaporating immediately after the break up, and this effect is likely caused by the strong winds resulting in the high evaporation rate even in days with low vapour pressure gradient.

To the best of our knowledge, the measurements of lake evaporation in Antarctica with the EC technique were done only on two lakes in the Schirmacher Oasis. However, these measurements are available for natural and artificial lakes globally (Guseva et al., 2022; Liu, 2023). While many studies analyse the surface energy budget of lakes, they typically report evaporation as a latent heat flux expressed in W m^{-2} or J cm^{-2} day^{-1} (Stewart and Rouse, 1976; Stannard and Rosenberry, 1991; Marsh and Verseghy, 2000; Oswald and Rouse, 2004; Blanken et al., 2011). These energy-based units are often impractical for hydrological (water balance) applications, and are difficult to compare directly with depths measured by other techniques (for example, evaporation pans). To ensure comparability of our results, we converted the reported latent heat fluxes (W m^{-2}) into equivalent depths of evaporated water.

The EC measurements remain rare on the lakes located north of 60° latitude. Ala-Könni et al. (2022) measured the evaporation over Lake Kuivajärvi (61°N, Finland) during the ice free period, and yielded the mean evaporation rate 1.5 – 2.5 mm d^{-1} during July – August; the evaporation up to 5 mm d^{-1} were observed in days with strong winds, or periods when dry and cold synoptic fronts pass. Blanken et al. (2000 and 2003) measured the evaporation over Great Slave Lake (62 °N, Canada) during the ice free period from June to September (1997, 1998, and 1999) when the evaporation rate reached 2.8 mm d^{-1}. The results imply that both (wind speed and vapor pressure gradient) control the evaporation over an ice free lake. Granger and Hedstrom (2011) estimated that 4–9 day-averaged evaporation rate reaches 3.2 – 3.9 mm d^{-1} during an ice free period for small lakes located next to Lake Great Slave. The northernmost EC measurements were obtained from a small Arctic thermokarst lake (72 °N, Russia) during an experiment spanning from April to August 2014 (Franz et al., 2018). The lake evaporation rate was estimated to be near zero from April through the mid June (covering the ice-covered and break-up periods), subsequently averaging 2.5 mm d^{-1} and peaking at 6 mm d^{-1} during the ice free period from July to September. These estimates of the lake evaporation are comparable with the evaporation over the ice-free Lake Zub/Priyadarshini measured with the EC technique (2017 – 2018).

Evaporation of the high-altitude lakes in the Qinghai-Tibetan Plateau (36°N, 3200 – 4700 m asl, China) has been extensively studied in last decades with the EC measurements spanning more than 5 years (Shi et al. 2024; Gan and Liu, 2020; Meng et al 2020; Wang et al., 2019; Li et al. 2016). Li et al. (2016) found that evaporation over the ice-free Lake Qinghai reaches up to 12 mm d^{-1} during storms, and windy days (wind speed > 4 m s^{-1}) contribute up to 22 % of the annual lake evaporation, and this contribution is comparable for the ice sublimation (23 %). Shi et al. (2024) conclude that evaporation over the ice-free Lake Qinghai is controlled by wind rather than by vapor pressure gradient or solar radiation. It is in accordance with our results shown that wind is the primary factor driving the evaporation over two lakes in Antarctica. We did not measure the radiation during the experiments, however, it is routinely measured with instrumentation installed at Maitri and Novo observatories. It opens opportunities for a new study on the lake surface energy balance.

Faucher et al. (2019) measure the ice sublimation over the surface of perennially frozen glacial Lake Untersee (71° S, Antarctica) with sticks during two years (), and estimate the sublimation (evaporation) between 400 to 750 mm $year^{-1}$ which gives the mean daily rate 1.1–2.1 mm d^{-1}. It is very close to the estimates obtained for the ice-covered Lake Glubokoe (2019 – 2020). Dugan et al., (2013) estimate the lake ice ablation (sublimation) of 5–31 mm d^{-1} for the lake in Tailor Valley (77°S, Antarctica); the authors, however, did not have means to distinguish between the contributions of surface melt and sublimation to ablation during summer. For winter, Dugan et al. (2013) found sublimation values ranging from 0.2 to 0.7 mm d^{-1}, and these values are naturally much smaller than our results obtained for the fully or partially ice-free lakes.

Leppäranta et al. (2016) reported the summertime ice sublimation of 0.7 mm d^{-1} for the lakes in the Vestfjella mountains, Droning Maud Land (73°S, Antarctica), and the ice thinning of 10 – 15 mm d^{-1} on a small pond on top of a nunatak during days with the winds speeds exceeding 30 m s^{-1} . Because of no surface melt during days in such windy days, the ice thinning is likely attributed to sublimation (Leppäranta et al., 2016). No EC observations were available from the site but we checked that according to the bulk-aerodynamic method these values are indeed possible under such a wind speed, if the ice surface temperature is close to 0 °C and the air is very dry.

The bulk-aerodynamic method is sensitive to actual values of the transfer coefficient of moisture (C_E) which is typically considered almost constant with a value of approximately $1.1\cdot10^{-3}$ over the ocean (Kantha and Clayson, 2000) and varying between 1.1 and $2.4\cdot10^{-3}$ over the natural and artificial lakes (Hicks, 1972; Guseva et al. 2022). Our results showed that the transfer coefficient of moisture (C_{ENz} , z = 10 m) varying between $1.4\cdot10^{-3}$ and $2.1\cdot10^{-3}$ (Lake Zub/Priyadarshini and Lake Glubokoe, respectively). With these actual values, the bulk-aerodynamic method underestimated the evaporation by 6 – 8%, and got the scores $s/\sigma < 0.80$ and RMSE of 0.6 mm d^{-1} , which are acceptable to recommend this method to simulate the daily lake evaporation rate for two lakes in the Schirmacher oasis. It aligns to conclusions of the studies focusing on the evaporation over the lakes of Qinghai-Tibetan Plateau (Wang et al., 2019).

The method used in calculation of evaporation is important for the shallow coastal lakes whose volume changes very little over a summer (Shevnina and Kourzeneva, 2017). Gopinath et al. (2020) found that summertime evaporation is a significant component in the water budget of shallow lakes in the Schirmacher Oasis. Lake Zub/Priyadarshni supplies freshwater to the

Maitri research station which is crewed with approximately 60 people at maximum during summers. A new Maitri-II research station is planned to open in 2032; expanding the investigation of Antarctica, it will accommodate up to 140 people (10 October 2025, https://timesofindia.indiatimes.com/). To plan the water supply/utilisation system of the stations, one should investigate (model) the water budget of Lake Zub/Priyadarshni. Dhote et al. (2021) evaluated the component of this lake water balance equation, namely the precipitation, inlet river runoff and wastewater return (as positive); and the outlet river runoff, evaporation and freshwater withdrawal (as negative).

The volume of Lake Zub/Priyadarshini is 1.03x10^6 m^3, and in January – February 2018 it decreased by 40.3 m^3; the discrepancies of the water balance equation was 670.6 m3 (Dhote et al., 2021). The lake evaporation of 58.5 m^3 was calculated following Odrova (1979), which underestimated the evaporation by72 % according to our results. In absolute values it corresponds to 42.1 m^3 (58.5 multiplied by 0.72), which represents approximately 6% of the discrepancy of the water balance equation. It can be reduced by using the bulk-aerodynamic method with the neutral 10-m turbulent transfer coefficient of moisture (C_{ENz}, z = 10 m) equalling 2.1 x 10^{-3}. The volume of precipitation over the surface of Lake Zub/Priyadarshini was 5.3 m^3 in January – February 2018, and it is 0.05% of the lake volume. The rest of discrepancy is likely contributed by methods used to evaluate the inlet/outlet river runoff.

Application of the traditional approaches to measure lake evaporation is limited by the difficulty of their maintenance and operation in Antarctica. Defining the actual values for the mass transfer and empirical coefficients requires detailed micrometeorological measurements on different heights which episodically done in Antarctica. The tracers-based approach may give an accurate estimate of lake evaporation (Fekete et al., 2006); it, also, requires regular water sampling followed by a unified programme. It, however, is relatively easy to organize on lakes located next to the human settlements.

6 Conclusions

We measured the evaporation with the eddy-covariance technique on two lakes in the Schirmacher Oasis (70°S), and estimated the summertime lake evaporation varied from 0.3 to 5.0 mm d^{-1}. Depending on the presence of lake ice, the average evaporation varied from 1.6 ± 0.2 mm d^{-1} in the early stage of lake ice break-up (December) to 3.0 ± 0.3 mm d^{-1} for the lake ice-free period (January – February). During the austral summer, the lakes are warmer than the ambient air on most days. The changes in evaporation were associated with changes in wind speed rather than changes in difference between the near-surface air temperature and lake surface water temperature.

We further quantified uncertainties in the bulk-aerodynamic method, combination and empirical formulas have been applied in estimations of summertime evaporation over two lakes in coastal Antarctica. The bulk aerodynamic method yielded the most accurate estimates of evaporation (biases of 6 – 8 % for the mean values) when the moisture transfer coefficient (C_{ENz}, z = 10 m) was set to $1.4\cdot10^{-3}$ and $2.1\cdot10^{-3}$ for Lake Zub/Priyadarshini and Lake Glubokoe, respectively. This method showed an acceptable skill in estimation of the daily evaporation over both lakes during the ice break-up and open water periods. Hence, the method can be recommended for hydrological (lake water balance) applications required for operational (short term) decision making. The selected combination formulas underestimated the daily evaporation over the lakes by 27–73%. With the correction, however, they can be applied to estimate the cumulative summertime evaporation, which is needed for the hydrological applications required for decisions on sites' maintenance and investment. These indirect methods need the measurements of lake surface water temperature.

Code availability. The code is available at the Github. The code for the figures 3, 4, 6, 7, 8 and 9 was written using suggestions given by the AI Github Copilot.

Data availability. The dataset is available at https://doi.org/10.5281/zenodo.14823402.

Interactive computing environment. Pycharm at Ubuntu (Noble Nombat).

Supplements. The evaporation estimated after the EC (direct) and indirect methods are given in ECH_ZB/GL.csv and BA_daily_ZB/GL.csv.

Author contribution. ES designed and ran the field experiments as well as calculated the evaporation applying the combination equations, their uncertainties, and skill scores. TV estimated the evaporation with the bulk-aerodynamic method. MP designed and analyzed the EC measurements in the field experiments. TN analyzed synoptic situation and weather during the experiments. All authors contributed to writing of the manuscript.

Competing interests. The contact author has declared that none of the authors has any competing interests.

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