Evaporation over a glacial lake in Antarctica

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Abstract. The study provides estimates of summertime evaporation over a glacial lake located in the Schirmacher oasis, Dronning Maud Land, East Antarctica. Lake Zub/Priyadarshini is the second largest lake in the oasis, and its maximum depth is 6 m. The lake is among the warmest glacial lakes, and it is free of ice during almost two summer months. The summertime evaporation over the ice-free lake was measured using the eddy covariance (EC) method, and estimated on the basis of the bulk-aerodynamic method and four combination equations. We used meteorological and hydrological measurements collected during a field experiment carried out in 2018. The EC method was considered the most accurate, and the evaporation was estimated to be 114 mm for the period from 1 January to 7 February 2018 (38 days) on the basis of this method. The average daily evaporation was estimated to be 3.0 mm day^{-1} in January 2018. The largest changes in daily evaporation were driven by synoptic-scale atmospheric processes rather than local katabatic winds. The bulk-aerodynamic method suggests the average daily evaporation to be 2.0 mm day^{-1}, which is over 32 % less than the results based on the EC method. This method is much better in producing the day-to-day variations in evaporation compared to the combination equations, which underestimated the evaporation over the lake open water table by over 40–72 %. We also suggested a new combination equation to evaluate the summertime evaporation of Lake Zub/Priyadarshini from meteorological observations from the nearest site. The performance of the new equation is better than the performance of the indirect methods considered. After this equation, the evaporation over the period of the experiment was 124 mm, which is only 9 % larger than the result according to the EC method.

1 Introduction

Liquid water is increasingly more present over margins of glaciers and ice sheets, and over the surface of the Arctic Sea ice and Antarctic ice shelf due to rise of near-surface air temperatures enhancing snow and ice melt. A large part of the melt water accumulates in a population of glacial lakes and streams, which are typical for the lowermost (melting) zone of glaciers and ice sheets, where the amount of liquid water is sufficient for both the surface and subsurface water runoff (Golubev, 1976). The area of the melting zone is evaluated from in-situ data gathered during glaciological surveys or from remote sensing data. The total area of the melting zone over the Antarctic ice sheet was estimated to be over 92.5 ± 13 thousands square kilometers based on the in-situ data collected during the period of 1969–1978 (Klokov, 1979). Estimations of the area of the melting zone in Antarctica are also available from microwave remote sensors for the summers in the period of 1979/80–2005/06, and already during this period the melting zone has covered over 25 % of the entire continent in at least five summers (Picard et al., 2007).

Recently, remote sensors and geophysical surveys have yielded evidence on a large number of glacial lakes in Greenland and Antarctica (Leeson et al., 2015; Arthur et al., 2020). In 2017, remote sensing data allowed the detection of more than 65000 glacial (supraglacial) lakes located over the East Antarctic coast during the peak melting season (Stokes et al., 2019). The total area of these supraglacial lakes was over 1300 km^2, and most of them were located at low elevations. Glacial lakes are connected by ephemeral streams into a hydrological network that may develop rapidly in the melting season (Lehnherr et al., 2018; Hodgson, 2012). During 2007 – 2016, the mass loss from the Antarctic ice sheet tripled relative to 1997 – 2006 (Meredith et al., 2019), and it explains the observed changes in physiographic parameters (volume, depth and surface area) of many of the glacial lakes located in the East Antarctic oases (Levy et al., 2018; Boronina et al., 2020). Glacial lakes are a well-known indicator for climate change (Verleyen et al., 2003; Williamson et al., 2009; Verleyen et al., 2012). The possible effects of the glacial lakes on global sea level rise are not clear because the processes and mechanisms driving meltwater production, accumulation and transport in the glacial hydrological network are not fully understood (Bell et al., 2017; Bell et al., 2019).

Among others, a modelling approach can help to understand how climate warming changes the amount of liquid water seasonally formed in the glacial hydrological network including the lakes and streams. The mass (or water) balance equation of a lake is among the models applied to evaluate the volume of a lake from known inflow and outflow terms (precipitation, evaporation, surface/subsurface inflow/outflow runoff, water withdrawal) measured or modelled (Chebotarev, 1975; Mustonen, 1986). In Antarctica, various processes drive the water exchange in the local lakes, and their mass (water) budget is closely linked to the heat budget (Simonov, 1971; Krass, 1986; Shevnina and Kourzeneva, 2017), and different numbers of the terms are important while estimating their volume depending on whether the lake is of the glacial type or the land-locked type. However, for all local lakes, the estimates of the water budget are sensitive to uncertainties inherent in the methods applied to evaluate evaporation (Shevnina et al., 2021).

Performing direct measurements of evaporation is difficult in practice, and therefore various indirect methods are used to evaluate the evaporation over the lakes. Finch and Calver (2008) categorize such methods into seven major models (approaches) needing various meteorological and hydrological measurements, and each approach has inherent strengths and weaknesses. The pan evaporation approach has good accuracy, however the maintenance of instruments is difficult to perform in remote locations, such as Antarctica. The mass (water) balance approach needs observations on the terms of the lake water budget (precipitation, surface/subsurface inflow/outflow runoff, water extraction, etc.) and knowledge of the lake's physiography (volume and surface area) to estimate the evaporation together with the discrepancy term. The discrepancy term depends on the uncertainties inherent in the hydrological and meteorological measurements and in the methods applied to estimate the terms of the lake's water budget (Finch and Calver, 2008). The application of the mass balance method for lakes located in Antarctica is not possible due to the lack of the hydrological observations. In the energy budget approach, evaporation from a lake is estimated as the term required to close the energy budget when all other terms of the budget are known (similarly to the mass balance approach). It needs a large number of observations with a high frequency of the measurements for temperature, wind speed, humidity and radiation fluxes (Finch and Calver, 2008).

In the bulk aerodynamic approach, the evaporation is calculated on the basis of data from the Earth surface properties (surface temperature and specific humidity as well as roughness lengths for momentum and moisture/heat) and atmospheric variables (wind speed, specific humidity and air temperature) in the lowermost part of the atmospheric boundary layer. In addition to observational studies on evaporation and associated latent heat flux, the bulk method is the cornerstone for parameterization of the turbulent fluxes of momentum and sensible heat in numerical weather prediction and climate models (Brunke et al., 2003). For Antarctic applications of the bulk method for evaporation and latent heat flux on the basis of in-situ and remote-sensing observations, see Broun et al. (2001), Vihma et al. (2002), Favier et al. (2011), and Boisvert et al. (2020).

The combination equation approach includes the elements of both energy balance and mass-transfer approaches in the estimation of evaporation. The Penman equation (Penman, 1948) is among the most famous presenting this approach, where evaporation is calculated from the simultaneous solution of diffusion equations for heat and water vapor, and the energy balance equation (Finch and Calver, 2008). A more general form of the combination equation is given by the Penman-Monteith equation (Monteith, 1965), which was developed to describe evaporation from plants (evapotranspiration). There are also a number of empirical formulas that need additional information on lake surface area, radiation, daily minimum and maximum air temperatures, etc. (Hojjati et al., 2020; Zhao et al., 2013) or require only the air temperature and relative humidity to be known (Konstantinov, 1968). The disadvantage of the empirical and combination equation approaches is that their application is limited by the features of the location where the empirical coefficients were estimated, and there are no regional values suggested for the Antarctic continent (Finch and Hall, 2001). The combination equations are also named as the Dalton-type equations in Odrova (1979). In this study, we estimated the uncertainties inherent in four equations while estimating the summertime evaporation over lakes located in Antarctica. The equations that were used were by Borghini et al. (2013) and Shevnina and Kourzeneva (2017), however, the uncertainties inherent in the estimations are not yet known.

Edinger et al. (1968) formulated a basis for the equilibrium temperature approach, where fluctuations in water temperature are driven by heat exchange. The empirical factors approach allows converting the estimations of evaporation from one type of the land surface to evaporation from another one with empirical coefficients (or factors), which are regionally specific (Finch and Calver, 2008). The estimates of the evaporation are also available from atmospheric reanalyses which share results of simulations carried out applying numerical weather prediction models. Also in the most recent global atmospheric reanalysis, the ERA5 of the European Centre for Medium-Range Weather Forecasts (Hersbach et al., 2020), the evaporation is estimated based on short-term weather forecasts applying the bulk aerodynamic method.

The eddy covariance (EC) method is recognized as the most accurate method in estimationing evaporation. This method has been introduced more than 30 years ago (Stannard and Rosenberry, 1991; Blanken et al., 2000; Aubinet et al., 2012), but it is rarely used in remote regions. The turbulence measurements require special instruments and sensors which are difficult to maintain and operate in places such as Antarctica.

This study addresses summertime evaporation over the ice-free water surface of a glacial lake evaluated by applying various methods, namely, the eddy covariance, the bulk-aerodynamic and combination equations. The EC measurements are used as a reference to evaluate the uncertainties in the estimates based on the bulk aerodynamic method and the combination equations. This information is beneficial, as EC measurements over glacial lakes are rarely available, and other estimates have to be used. The field experiment was carried out on the shore of the largeLake Zub/Priyadarshini located in the Schirmacher oasis, East Antarctica, from 1 of January to 8 of February 2018.

2 The study area, weather and lakes

The Schirmacher oasis (70° 45′ 30″ S, 11° 38′ 40″ E) is located approximately 80 km from the coast of the Lazarev Sea, Queen Maud Land, East Antarctica (Fig. 1a). The oasis is the ice-free area elongated in a narrow strip around 17 km long and 3 km wide from west–north-west to east–north-east, and its total area is 21 km^2 (Konovalov, 1962). The relief is hillocks with absolute heights up to 228 m above sea level. The oasis separates the continental ice sheet from the ice shelf, and the region allows studies on deglaciation processes and continental ice sheet mass balance components including melting and liquid water runoff (Klokov, 1979; Srivastava et al., 2012).

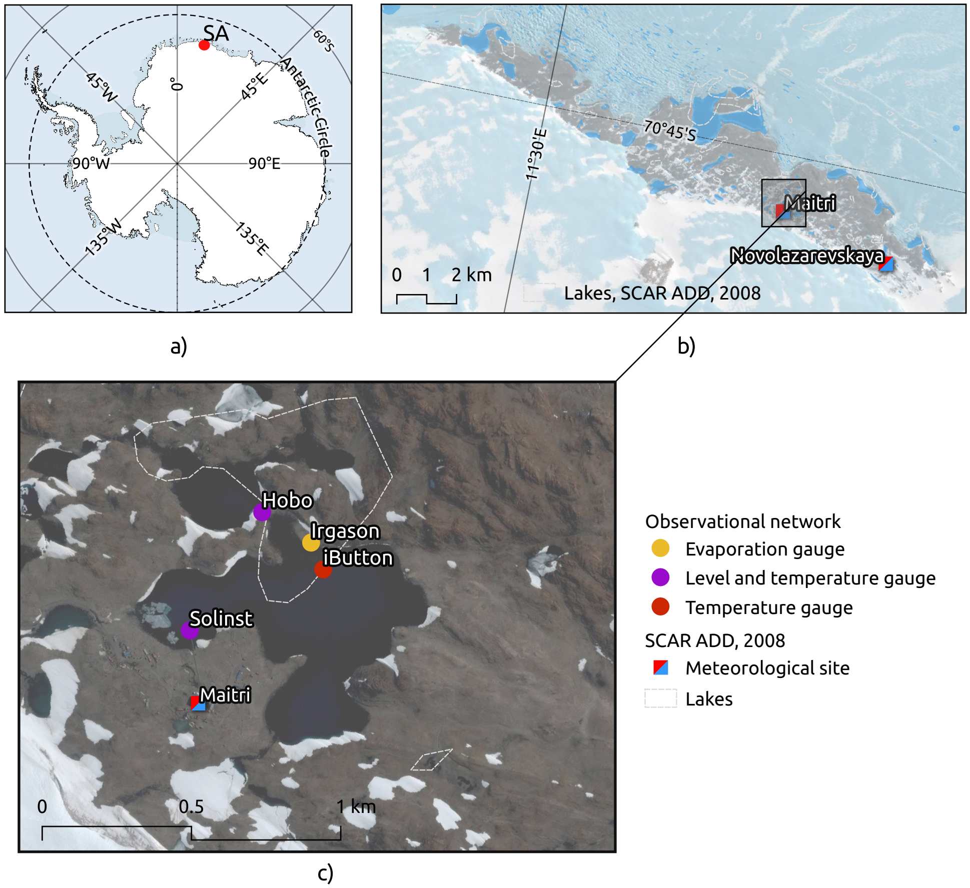

Figure 1. The lakes in the study region: (a) Location of the Schirmacher oasis (SA) in Antarctica; (b) the lakes in SA (SCAR ADD) with Landsat Image Mosaic of Antarctica, LIMA (https://lima.usgs.gov/) given as the background; (c) the observational network in the catchment of Lake Zub/Priyadarshini with a Google Earth image given as the background.

The location and boundary of the lakes are given in the Scientific Committee on Antarctic Research (SCAR) Antarctic Digital Database (ADD): https://www.add.scar.org/, last access February 8, 2022. In this dataset the location of the lakes in the Schirmacher oasis were systematically shifted to the LIMA composite (Fig. 1 b and the red lines in Fig. 1c). The climate of the oasis is characterized by low air humidity and temperature, and persistent (katabatic) wind blowing most of the year. This easterly-south-easterly wind blows from the continental ice sheet, and advects cold continental air masses to the oasis (Bormann and Fritzsche, 1995). There are two meteorological sites operating in the Schirmacher oasis: the observations were started in 1961 at the Novolazarevskaya (Novo) meteorological site (70°46′36″S, 11°49′21″ E, 119 m asl, World Meteorological Organization (WMO) number 89512). The Maitri meteorological site (70°46′00″S, 11°43′53″ E, 137 m asl, WMO number 89514) opened in 1989, and is located 5.5 km from the Novo site. Both meteorological sites are included in a long-term monitoring network, and their measurements are done according to the WMO's standards (Turner and Pendlebury

2004). The meteorological data gathered at these two stations are available from the British Antarctic Survey Dataset (https://www.bas.ac.uk, last accessed 14.12.2018). Table 1 shows weather conditions during the austral summer 2017–2018 and averaged over the period of 1961–2010 according to the observations at the Novo site (the data given by the Arctic and Antarctic Research Institute at http://www.aari.aq/default_ru.html, last access December 7, 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), and their monthly average calculated for the austral summer 2017 – 2018. The values are evaluated from the observations at the Novo site.

Meteorological parameter	1961 – 2010			2017	2018	2018
	December	January	February	December	January	February
Air temperature, °C	–3.9 / –1.0 /1.5	–2.5 / –0.4 / 1.4	–4.7 / –3.3 / –1.0	–0.1	–1.3	–3.0
Relative Humidity,%	47 / 56 / 69	49 / 56 /66	41 / 49 / 59	50	57	49
Atmospheric pressure, Pa	965 / 975 / 991	964 / 976 / 986	964/ 973 / 987	970	970	967
Wind speed, ms^{-1}	4.3 / 7.4 / 10.3	3.1 / 7.0 / 10.4	5.8 / 9.4 / 13.1	7.0	6.2	9.4
Soil surface temperature, ºC	3.0 / 6.7 / 10.0	3.0 / 6.7 /11.0	–2.0 / 0.2 / 4.0	5.0	3.0	0.0
Precipitation, mm	0.0 / 5.3 / 54.8	0.0 / 2.6 / 38.0	0.0 / 2.9 / 25.9	1.9	10.9	4.6

The field experiment lasted 38 days in January–February 2018. Generally, the weather during the experiment was colder and less windy, while the relative humidity and amount of the precipitation were close to the monthly means estimated for the period 1961–2010 (Table 1). According to data from the Novo meteorological site, during the period of the campaign the daily air temperatures ranged from –8.3 to 2.8 ºC, and the wind speed from 1.5 to 14.3 ms^{-1}, with an average of 6.2 ms^{-1}. The observations at the Maitri site were very similar to those at the Novo site, with the correlation coefficient between the daily series of air temperature, relative humidity and wind speed varying from 0.95 to 0.98. According to the Maitri meteo site, the wind speed varied from 1.6 to 14.4 ms^{-1}, with an average of 6.7 ms^{-1}. The air temperature ranged from –8.3 to 2.1 ºC, with an average of 1.5 ºC. The average relative humidity during the summer was 54 %.

More than 300 lakes are mapped in the Schirmacher oasis (Fig. 1 b), and many of the lakes stay free of ice in the summertime for almost two months (Simonov, 1971; Richter and Borman, 1995; Kaup and Haendel, 1995; Kaup, 2005; Phartiyal et al., 2011). The hydrological cycle and changes of the lakes' volume are modulated by the seasonal weather cycle (Sokratova, 2011; Asthana et al., 2019). The physiography of the lakes is available from bathymetric surveys for only the largest lakes (Simonov and Fedotov, 1964; Loopman et al., 1988; Khare et al., 2008; Dhote et al., 2021). This study focuses on Lake Zub (also known as Lake Priyadarshini), and hereafter we will use both names of the lake, which is among the largest and warmest water bodies of the Schirmacher oasis. The lake's surface area is 35 $x10^3$ m^2, its volume is over 10 $x10^3$ m^3, and the maximum depth is 6 m (Khare et al., 2008; Dhote et al., 2021). Lake Zub/Priyadarshini occupies a local depression and is fed by two inflow streams present in warm seasons. The outflow from the lake occurs via a single stream. The lake stays free of ice for almost two summer months from mid-December to mid-February (Sinha and Chatterjee, 2000). The water level (and volume) of Lake Zub/Priyadarshini has been reducing continuously, and in 2018 the lake water level lowered by approximately 0.4 m (Dhote et al., 2021). The lake is used as the water supply for the year-round Indian scientific base Maitri.

Gopinath et al. (2020) used water samples collected from 12 lakes (including Lake Zub/Priyadarshini) located in the Schirmacher Oasis to recognize major sources of water in the lakes. The samples were analysed with the isotope method (Ellehoj et al., 2013), and the isotopic concentrations show that Lake Zub/Priyadarshini is mostly sourced by the melting of the adjacent glaciers. For landlocked lakes, the major source of water is melting seasonal snow cover (Shevnina and Kourzeneva, 2017; Hodgson, 2012). It allows us to suppose that Lake Zub/Priyadarshini is the glacial type, as it is not the landlocked type as given in Phartiyal et al. (2011). Lake Zub/Priyadarshini is the lowest in the chain of the glacial lakes sourced by the ice/snow melting in the lowermost zone of the glaciers, and we estimated that more than 60 % of its catchment area is covered by rocks. This allows for the specific thermal regime and water balance of this glacial lake, which is among the warmest in the oasis: its water temperature rises up to 8 – 10 ºC in January (Ingole and Parulekar, 1990). Such water temperatures are typical for the landlocked lakes (Simonov, 1971).

Lake Zub/Priyadarshini presents ideal conditions to study evaporation over a glacial lake, and to plan the field experiment, we accounted for the location to set up the EC measuring systems. Selection of the exact site for EC measurements requires, among others, data on the prevailing winds and their fetch over the lake, and naturally also the accessibility for regular maintenance. To evaluate the prevailing wind direction, we used 6-hourly synoptic observations at the Novo site available from the British Antarctic Survey Dataset (https://www.bas.ac.uk, last accessed December14, 2018) covering the period 1998–2016. We calculated the number of cases when wind was blowing from 36 sectors each 10 degrees wide, and then defined the prevailing wind directions (marked with the black arrows in Fig. 2). The prevailing wind directions range from 110 to 140º.

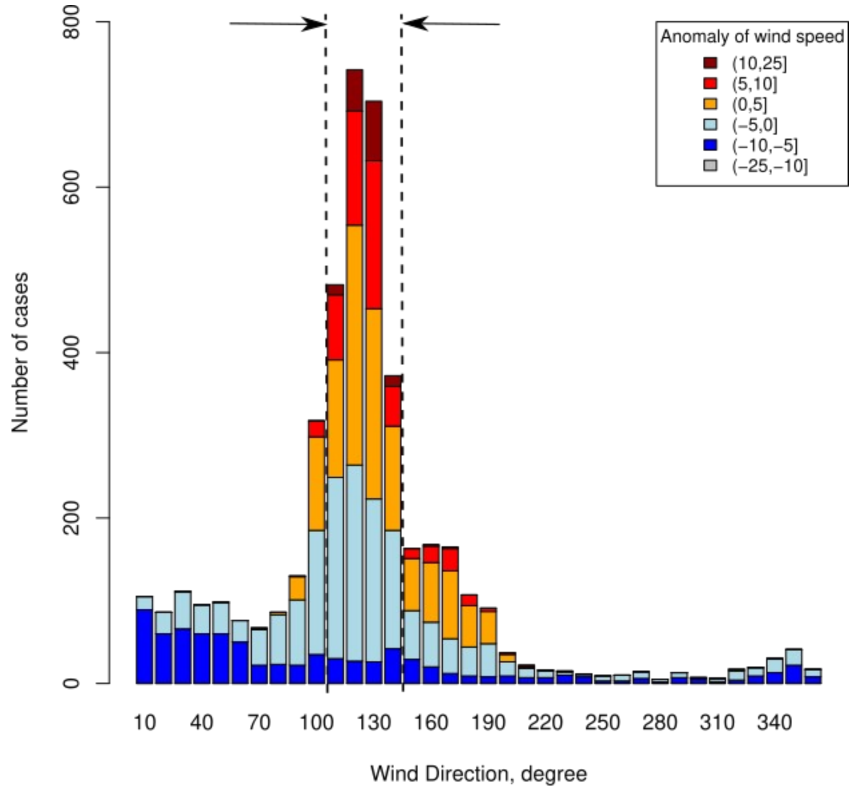

Figure 2: Wind direction and wind speed anomalies for two austral summer months (December and January).

We also calculated the wind speed anomalies of each 10-degree sectorsgiven in colour codes in 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. The region of the study is featured by persistent katabatic winds blowing from the continental interior. Fig. 2 shows that almost all 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.

3 Data and Methods

3.1 Data

In the field experiment in 2017–2018, we collected the hydrological and meteorological observations needed to evaluate the water balance terms of Lake Zub/Priyadarshini. The hydrological network included water level/temperature gauges, water discharge/level gauges and an evaporation gauge (Fig. 2 c). In this study we used only those data required in the evaluation of only one term of the water budget of a lake, namely evaporation. The evaporation gauge was a flux tower equipped with an Irgason device by Campbell Scientific. The Irgason consists of a 3D sonic anemometer and two gas analysers measuring CO_2/H_2O concentrations, and it is one of theEC devices taking this kind of measurements (in this study, we named our EC station as Irgason). The Irgason was deployed on the shore of the lake to collect high-frequency data on wind speed/direction and water vapor concentration needed to evaluate evaporation with the EC method (Fig. 3 a). The flux tower was placed 5–6 m inland of the shoreline of Lake Zub/Priyadarshini for the period of 1 January to 7 February 2018 (Shevnina, 2019). The meteorological parameters (air temperature, wind speed and relative humidity) were measured simultaneously at the Maitri meteorological site and at the evaporation gauge located on the lake shore (Irgason in Fig. 1 c). The data gathered by the sensors cover various observational periods (Table 2). The shortest 14-day period with the measurements is available for the iButton sensor, and this period lasted from January 27 to February 9, 2008.

Table 2. The hydrological and meteorological data collected during the field experiment in the summer 2017–2018: "–" no information available.

Site / Sensor (Fig. 1 c)	Elevation, m	Measured variables	Accuracy / (Resolution)	Time series used in the analysis	Period
Irgason site	124	air temperature, ºC; H_2O concentration, g/m^3; wind speed, ms^{-1}	±0.15 / (0.025) ±0.037 / (0.00350) –	30 minute	01.01.2018 – 07.02.2018
Hobo	122	water temperature, ºC barometric pressure, Pa	±0.44 / (0.10) –	daily average (not used in this study)	30.12.2017 – 09.02.2018
iButton	122	water temperature, ºC	±0.5 / (0.5)	daily average	27.01.2018 – 09.02.2018
Maitri site	137.5	air temperature, ºC; relative humidity, %; wind speed, ms^{-1}	±0.2 / (–) ±1/ (–) ±0.5 / (–)	daily average	01.12.2017 – 28.02.2018

Table 2 shows the information on the accuracy and resolution of the sensors according to the technical specifications given by the manufacturers. Ramesh and Soni (2018) give the information for the sensors installed at the Maitri site. The elevation of the lake water level was measured by the geodetic instrument Leica CS10 during the installation of the Solinst water level logger on 30 December 2017 (Fig. 1 c). The same instrument was used to measure the elevation of the Maitri site in January 2018 (Dhote et al., 2021). The elevation of the lake water level was 122.3 m (WGS84 ellipsoid vertical datum), we further used this elevation while calculating the elevation for the Hobo, iButton and Irgason temperature sensors.

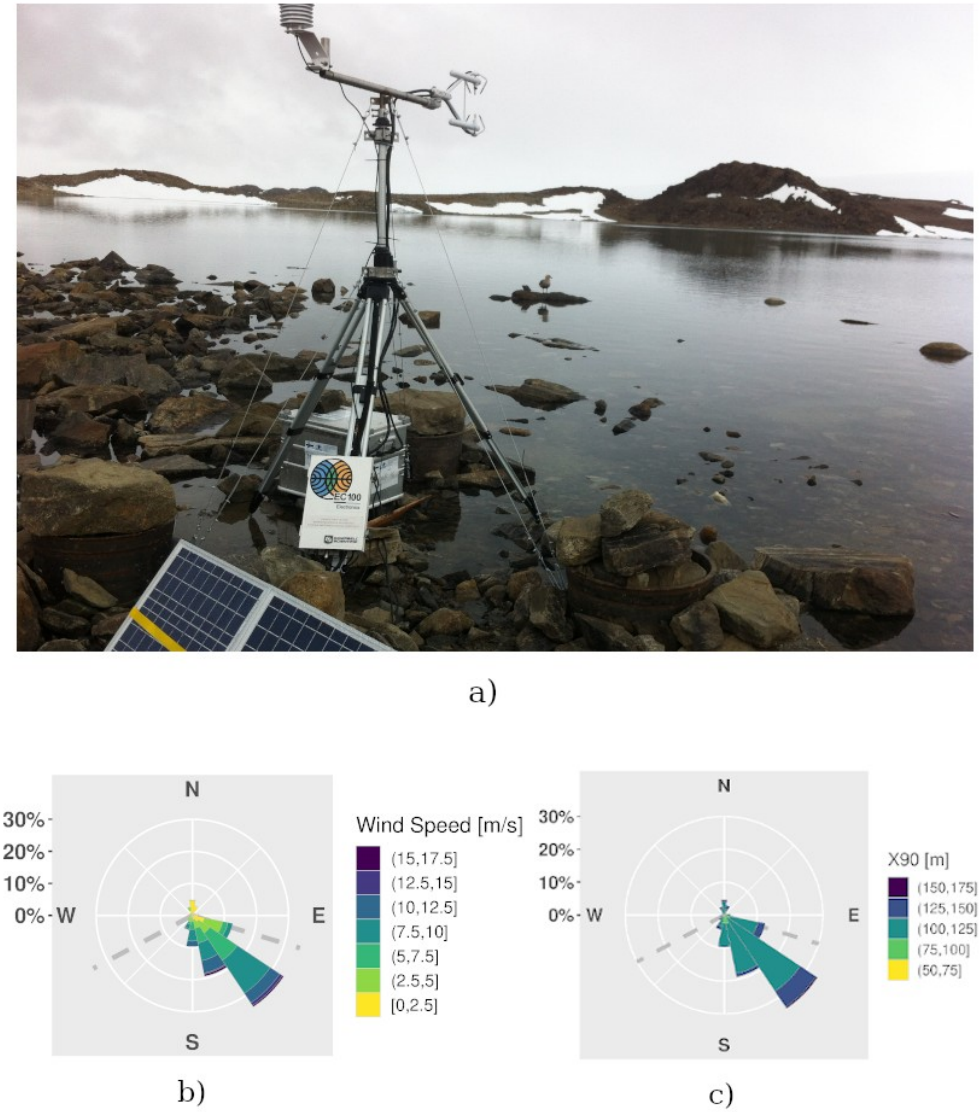

Figure 3: The experiment on the coast of Lake Zub/Priyadarshini: (a) Irgason deployed on the lake shore (06.01.2018); (b) wind speed and direction measured at the Irason site, dashed line indicates the footprint wind sector; (c) the footprint length estimate (X90).

The footprint is an important concept for evaluating fluxes correctly with the EC method. The footprint is defined as the area upwind of the EC station where the fluxes observed at the station originate from Hence, the footprint area depends on the location of the EC station, the height of its sensors, the roughness of the upwind surface, and the stratification of the upwind atmospheric surface layer (Kljun et al., 2004; Burba, 2013). The Irgason was settled at the height of 2 metres above the ground, which yields footprint lengths of less than 200 metres, which in this study was defined as X90 and represented 90 %

of the cumulative contribution to the fluxes (Fig. 3 c). This distance is less than twice of that between the Irgason and the shore of Lake Zub/Priyadarshini in an east-southeast direction (Fig. 1 c), and it ensures that the measured data is representative only for the lake and free of contamination from the upwind shore. The tower height of 2 m generates a blind zone near the tower, so that the stones on the downwind shore do not affect the fluxes. The location of the EC tower accounted for the prevailing wind directions (Fig. 2) meaning that the footprint area is mainly represented by the lake surface. We filtered out data outside the footprint (Fig. 3 b). Gaps in the wind direction were replaced with the average values of the neighboring 30-minute blocks. The Irgason's raw data consisted of values measured at a frequency of 10 Hz. We used these raw data to calculate a 30-minute time series of evaporation, turbulent fluxes of momentum, sensible heat and latent heat, as well as air temperature, wind speed, and wind direction. The daily evaporations were calculated as a sum of the 30-minute time series. The low observation height of 2 m guarantees that the vertical divergence of the water vapour flux is negligible, and therefore the water vapour flux observed at the height of 2 m represents the surface evaporation.

To allow the estimation of evaporation by the combination equations, measurements of the water temperature are needed. We measured the water temperature of the lake's surface with two sensors: the iButton temperature sensor was installed in Lake Zub/Priyadarshini in the depth of 0.2 metres and was placed ahead of the EC station (Irgason) toward the prevailing wind directions. The Hobo temperature sensor was deployed in the depth of 0.2 metres in the end of the stream inletting the neighbouring lake (Fig. 1 c). This stream is an outlet of Lake Zub/Priyadarshini, and we assumed that the observations collected by the Hobo were representative for the stream more than for the neighbouring lake itself. The accuracy of both temperature sensors is similar, and the resolution of the Hobo temperature sensor is better than the iButton's resolution (Table 2). The lake water temperature was measured every 10 minutes, and we further calculated the daily average time series of the water temperature in the lake. The Hobo consists of two sensors measuring temperature and barometric pressure allowing us to evaluate the water level/stage, and we used only the temperature measurements in this study. Sinha and Chatterjee (2000) reported that Lake Zub/Priyadarshini was thermally homogeneous down to the bottom almost from mid-January 1996 to mid-February 1997, and we assumed that the lake had no thermal stratification during the whole field experiment in 2018.

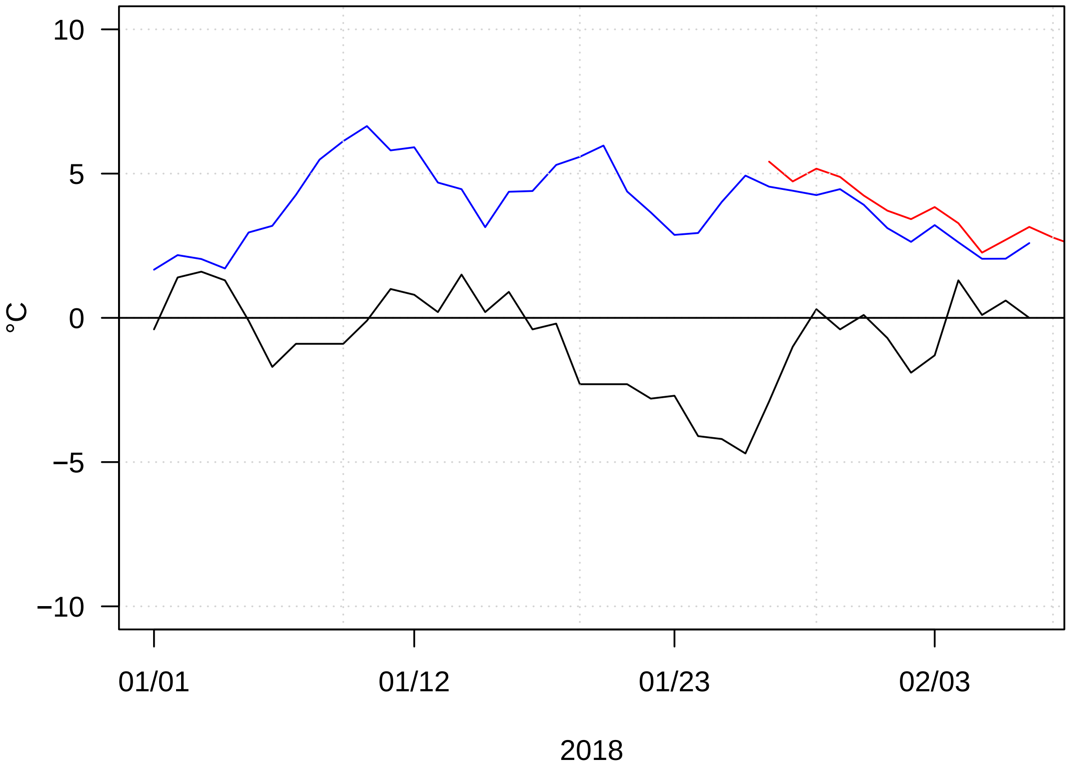

Figure 4: Daily time series of the lake surface water temperature measured by the Hobo (blue), by the iButton (red), and the air temperature measured at the Maitri site (black).

Figure 4 shows the daily time series of the lake water temperature, air temperature and wind speed calculated from the measurements done by the sensors during the period of the experiment. The correlation coefficient equals to 0.89 for the series of the water temperature measured by two temperature sensors (Hobo and iButton). We further used the measurements collected by the Hobo temperature sensor to estimate the evaporation over Lake Zub/Priyadarshini in January 2018. In our calculations based on the combination equations we applied the data collected by the meteorological sensors installed both at Maitri and Irgason sites, and the meteorological sensors are deployed at the different elevations (Table 2). The elevation of the temperature sensor and gas analyser of the Irgason is lower than the sensors at Maitri site, and therefore we used the logarithmic approximation of the wind profile to correct the wind speed data measured at the Maitri site, for which we estimated a constant aerodynamic roughness length of 0.002 m (Stull, 2017). We did not use any height correction for the data on the relative humidity and air temperature since their changes with elevation are negligible in our case (Tomasi et al., 2004).

3.2 Methods

To evaluate the evaporation with the direct EC method, we used the data collected by the Irgason installed on the shore of Lake Zub/Priyadarshini. The Irgason raw data were measured with a frequency of 10 Hz, which were further analysed in the following steps. First, the bad data with less than 50 % of total 10 Hz measurements were excluded. Second, we excluded all data automatically flagged for low quality, and the data with a gas signal strength less than 0.7 (or 70 % of the strength of a perfect signal). The gas signal strength is usually lower than 0.7 during rainfalls, which were not observed in January, 2018 in the Schirmacher oasis. Generally, rainfalls are rare along the East Antarctic coast where rainfall occurs 22 days per year at most (Vignon et al., 2021). In the third step, the spiks were removed applying the method by Vickers and Mahrt (1997), fixing the threshold window of 3.5 standard deviation for horizontal wind speed, CO_2 and H_2O and 5.0 for vertical wind speed. This procedure was repeated up to 20 times or until no more spikes were found. Finally, we obtained, among others, the 30-minute fluxes of momentum, sensible heat and latent heat (evaporation), as well as the water vapor and carbon dioxide concentrations (see the Supplement). The evaporation over the lake was calculated only by those values collected within the footprint of the ice-free surface of the lake. Therefore, we filtered the data outside the footprint which covered the wind directions within the range of 105 – 240º (Fig. 3 b). We excluded 18 % of the total data from further consideration after the three step filtering. To fill these gaps we replaced the excluded values by the mean value, which was estimated from the time series of 30-minute values. We also evaluated the relative humidity from the water vapor concentration as given by Hoeltgebaum et al. (2020).

Uncertainties in the estimation of evaporation by any method include instrumental errors associated with the specific instrument. Aubinet et al., (2012) suggest three methods allowing the quantification of the uncertainty of the EC method. In this study, we applied the paired tower method to evaluate the instrumental uncertainties of the EC method taking advantage of an intercomparison campaign in Alqueva reservoir, Portugal, in October 2018. The instrumental error does not depend on the region where the instrument will be used, and therefore the intercomparison may be done elsewhere. The relative instrumental error estimated in this intercomparison campaign was 7 % (see the Annex). The uncertainties of the EC method also include the errors due to the filtering of measurements within the footprint area. The large number of filters and corrections that we applied to the EC data allowed us to reduce the errors and uncertainties. Even the EC method itself has some errors and uncertainties but it is the most versatile and accurate method to measure evaporation.

In the bulk-aerodynamic approach, evaporation is defined as the vertical surface flux of water vapor due to atmospheric turbulent transport. It is calculated from the difference in specific humidity of the surface (i.e., ice or water for which the specific humidity equals the saturation specific humidity that depends on the surface temperature), and the air, as well as the factors that affect the intensity of the turbulent mixing: wind speed, surface roughness, and thermal stratification (Boisvert et al., 2020; Brutsaert, 1985).

The evaporation based on the bulk-aerodynamic method is calculated as follows:

$$E = \rho \quad C_{Ez}\, w_z\, (q_s - q_{az}) \tag{1}$$

where E is the evaporation (in kg m^{-2} s^{-1}, which we in the following convert to mm day^{-1}), ρ is the air density, (kg m^{3}); C_{Ez} is the turbulent transfer coefficient for moisture unitless), q_s is the saturation specific humidity at water surface of the lake(kg/kg), q_s is the air sauturation specific humidity (kg/kg), and w is the wind speed (m). The subscript z refers to the observation height (here 2 m). The turbulent transfer coefficient for moisture depends on the atmospheric stratification: for C_{Ez} under neutral stratification (C_{EzN}) we applied the value of 0.00107 based on previous measurements over a boreal lake (Heikinheimo et al., 1999; Venäläinen et al., 1998). It allows us to better take into account the different regime of turbulent mixing over a small lake compared to the sea (Sahlée et al., 2014).

Since the stratification of the atmosphere is not always neutral, we took into account its effects on the turbulent transfer coefficient as follows:

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

where, C_{DzN} is the neutral drag coefficient for the lake surface, k is the von Karman constant (0.4), ψ_m and ψ_q are empirical stability functions ; and L is Obukhov length (in meters):

$$L=-\frac{\rho c_p u_*^3 \theta_z}{k g H} \tag{3}$$

where ρ is the air density, c_p is the specific heat, u_* is the friction velocity, θ_z is the air potential temperature, g is the acceleration due to gravity, and H is the surface sensible heat flux. The Obukhov length (Obukhov, 1946) is the key element of the Monin-Obukhov similarity theory (Monin and Obukhov, 1954; Foken, 2006), and needed to adjust the bulk transfer coefficients to the actual stratification in the atmospheric surface layer. In our calculations, the neutral drag coefficient equals to 0.00181 as suggested by Heikinheimo et al. (1999). For ψ_m and ψ_q, 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 values by Heikinheimo et al. (1999) were given for z = 3 meters, and converted to our observation height of 2 meters using Launiainen and Vihma (1990), and the same algorithm was applied to iteratively solve the interdependency of the turbulent fluxes and L. The latent heat flux is obtained by multiplying the evaporation rate by the latent heat of vaporization.

Most of the empirical equations are based on a simple mass transfer relation between the evaporation rate and the water deficit and wind conditions. The general form of the relation reads as $E = K\, w_z\, (e_s - e_z)$, where K is empirical function approximated with a small number of coefficients. Among others, Shuttleworth (1993) suggests two mass transfer equations for the estimation of evaporation from the surface of lakes and ponds depending on their surface area. In this study, we used his formula for water bodies in the range of 50 m < $A^{0.5}$ < 100 km located in regions with a relatively arid climate. The equation reads as $E = 2.909\, A^{-0.05}\, w_2(e_s - e_2)$, where E is the evaporation in mm day^{-1}; A is the surface area in m^2; w_2 is 2-metre wind speed in ms^{-1}; and e_s and e_2 are the surface water and air vapor saturation pressure in kPa. In this study, we used this formula to estimate the daily evaporation from Lake Zub/Priyadarshini, whose surface area is estimated as 350 000 m^2

in 2016 (Dhote et al., 2021). The method by Shuttleworth (1993) has been used to evaluate evaporation over small lakes located in Antarctica (Boghini et al., 2013), however the scope of uncertainties inherent in the method is not known.

Penman (1948) first suggested taking the elements of the mass transfer and energy budget approaches into the estimation of evaporation from open water, and his formula is one of the combination equations (Shuttleworth, 1993; Finch and Calver, 2008). In this study, we applied three combination equations to calculate daily evaporation: $E = 0.26\ (1 + 0.54\ w_2)(e_s - e_2)$ and $E = 0.26\ (1 + 0.86\ w_2)(e_s - e_2)$ adopted from Tanny et al. (2008), where these formulas are referred to Penman (1948) and Doorenbos and Pruitt (1975) respectively. These equations are among those most often used in hydrological practice (Finch and Calver, 2008), and therefore we have chosen them in this study. We also used the formula $E = 0.14\ (1 + 0.72 w_2)(e_s - e_2)$, which has been applied to evaluate evaporation from lakes located in northern Russia (Odrova 1979). In these equations, e_s and e_2 are the surface water and air vapor saturation pressure (millibars), and we calculated them according to Tetens's formula given in Stull (2017). The method by Odrova (1979) has been used in estimations of evaporation over glacial lakes located in Antarctica (Shevnina and Kourzeneva, 2017), but the method's uncertainties have not been estimated. We calculated daily evaporation separately using the meteorological observations collected at the Maitri site and at the lake shore (Irgason site).

The empirical coefficients in the combination equations usually limit their applicability to the region where such coefficients are obtained (Finch and Hall, 2005). The empirical coefficients in four selected equations are evaluated from data gathered in regions with different climates, and therefore they probably will not be applicable for lakes located in Antarctica. In this study, we suggested two regional empirical relationships based on the daily series of evaporation estimated by the direct EC method and the meteorological observations at the Maitri site, which is the nearest meteorological site to the lake. In the first relationship, evaporation (E, mm day^{-1}) was evaluated with the linear model $(a + b\ w_2)\ (e_s - e_2)$, where a and b are fitted with empirical coefficients, and $(e_s - e_2)$ is expressed in mbar. The second relationship reads as $E = a\ w_2^{\,b}\ (e_s - e_2)$. The efficiency of fitting the coefficients were performed on the same data for the experiment (lasting 38 days).

Evaporation by the indirect methods were compared to the direct (EC) method in order to find the method with the lowest scope of the uncertainties, and, therefore, the method of the highest efficiency. We applied the Pearson correlation coefficient (*PR*), the root square standard error ($RMSE = \sqrt{\sum_{1}^{n} (E_{EC} - E_{mod})^2}$) and the s/σ criteria (*SSC*) to evaluate the scope of the uncertainties inherent in the indirect methods. The *SSC* reads as follows (Popov, 1979): $s = \sqrt{\sum_{i=1}^{n} (E_{EC}^{i} - E_{m}^{i})^2 / (n-m)}$, and $\sigma = \sqrt{\sum_{i=1}^{n} (E_{EC}^{i} - \bar{E}_{EC})^2 / n}$. In these formulas, $\bar{E}$ is the mean evaporation, (mm); n is the length of the series (38), and m is the number of empirical coefficients in the relationships (equal to 2). Overall, a new method is acceptable for further use in hydrological practice if the SSC value is less than 0.8 (Popov, 1979).

4 Results

4.1 Evaporation

We considered the direct EC method as the most accurate, providing the reference estimates for the evaporation over the lake surface (Finch and Hall, 2005; Tanny et al., 2008; Rodrigues et al., 2020). According to the EC method, the daily evaporation varied from 1.5 to 5.0 mm day^{-1} with the average being equal to 3.0 mm day^{-1}, and the standard deviation was ± 1.1 mm day^{-1}. The average was calculated by dividing 114 mm of evaporated water (which is the sum of the 30-minute series of evaporation) by the number of days in the observational period (which is 38). The sum of the evaporation over the period of the field experiment is 94 mm, if we simply excluded the gaps in the 30-minute series.

We estimated the uncertainties inherent in the indirect methods by comparing their results with those based on the EC method. The average daily evaporation was 2.0 mm day^{-1} calculated by the bulk aerodynamic method with the mass transfer coefficients after Heikinheimo et al. (1999), and this value is approximately 30 % less than those estimated by the EC method, and it is the best estimate among the indirect methods (bold notation in Table 3). All combination equations underestimated the evaporation over the lake surface by over 30 – 75 %, and the method by Odrova (1979) yielded the greates underestimation of the mean daily evaporation over the lake surface. The uncertainties in the estimates by indirect methods are approximately the same for both cases of the input data (Maitri and Irgason).

Table 3. The daily evaporation (mm day^{-1}) over the surface of Lake Zub/Priyadarshini for the period of 01.01.2018 – 07.02.2018): *SD* is the standard deviation; *r* is ratio the sum E_{EC} divided by the sum E_m.

Method	Input data: Irgason site				Input data: Maitri site			
	Min/Max	Mean ± *SD*	Sum	*r*	Min/Max	Mean ± *SD*	Sum	*r*
Bulk aerodynamic (Heikinheimo et al., 1999)	**0.6 / 3.5**	**2.0 ± 0.8**	**78**	**1.5**	**0.7 / 2.9**	**1.9 ± 0.6**	**72**	**1.6**
Shuttleworth, 1993	0.2 / 1.8	1.0 ± 0.4	38	3.0	0.1 / 1.9	0.9 ± 0.4	36	3.2
Penman, 1948	0.0 / 2.0	1.3 ± 0.5	2.4	1.9	0.1 / 2.2	1.2 ± 0.5	46	2.5
Doorenbos and Pruitt, 1975	0.0 / 2.9	1.8 ± 0.8	68	1.7	0.2 / 3.2	1.7 ± 0.7	66	1.4
Odrova, 1979	0.1 / 1.3	0.8 ± 0.3	32	3.6	0.1 / 1.6	0.8 ± 0.3	32	3.6

Figure 5 shows the daily evaporation estimated by the direct EC against those estimated by the four indirect methods calculated based on the meteorological observations collected at two measurement sites: Maitri and Irgason. There was not a large difference in the results, and therefore we can recommend using the meteorological observations gathered by the nearest site in further estimation of evaporation. Table 5 gives a summary of the scope of the uncertainties and efficiency of the indirect methods to model the day-by-day series of the evaporation with the selected criteria.

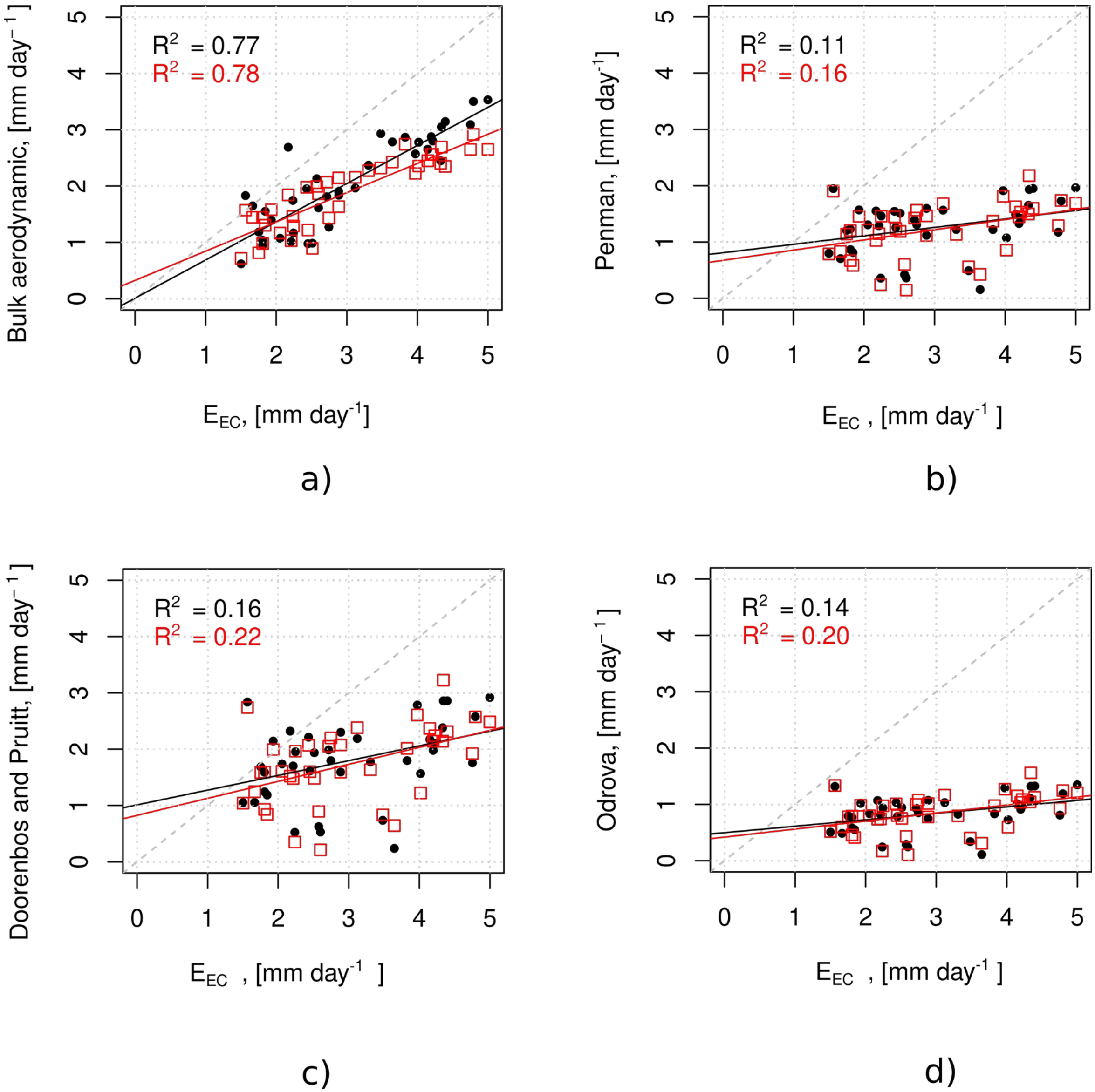

Figure 5: Scatter plots of the daily evaporation estimated with the indirect methods (Y-axis) against the direct EC method (X-axis): (a) the bulk-aerodynamic; (b) Penman; (c) Doorenbos and Pruitt; (d) Odrova. R^2 refers to the determination coefficient. The red dots indicate the estimates of the evaporation with the meteorological parameters measured at the WMO synoptic site Maitri, which is the nearest site to Lake Zub/Priyadarshini. The black dots indicate the estimates of the evaporation done with the meteorological parameters measured at the lake shore (Irgason site).

The bulk aerodynamic method gave the best fit to the EC method according to all criteria (bold notation in Table 4). As one can expect, the efficiency of the empirical equations is poor: the correlation coefficient varied from 0.33 to 0.55, and both the *RMSE* and *SSC* criteria indicate the low ability of the methods to estimate daily evaporation.

Table 4. The efficiency of the indirect methods with the Pearson correlation coefficient (*PR*), the root square standard error (*RMSE*) and the s/σ criteria (*SSC*).

Method	Input data: Irgason site			Input data: Maitri site		
	PR	*RMSE*	*SSC*	*PR*	*RMSE*	*SSC*
Bulk aerodynamic (Heikinheimo et al., 1999)	**0.87**	**1.0**	**1.1**	**0.88**	**1.1**	**1.2**
Shuttleworth, 1993	0.55	2.1	2.3	0.39	2.2	2.3
Penman, 1948	0.35	1.8	2.0	0.41	2.1	2.0
Doorenbos and Pruitt, 1975	0.43	1.3	1.6	0.46	1.6	1.6
Odrova 1979	0.35	2.2	2.4	0.45	2.4	2.4

The bulk-aerodynamic method also yields the best estimates for day-to-day time series of evaporation (Table 4). However, even this method cannot be suggested to match the daily evaluation of evaporation using the meteorological observations at the Maitri site (Fig. 6). The mean difference between the daily evaporation estimated by the EC and the bulk-aerodynamic method is 0.6 mm day^{-1}, and it is the greatest on those days when the wind speeds are 6 – 7 m s^{-1}. Therefore, the relationship between evaporation and 2-meter wind speed and saturation deficit was approximated by the formula reading as $E = a + bw_2 (e_s - e_2)$, and it's similar to the combination equations (given in form $E= a (1 + a\ b\ w_2)(e_s - e_2)$ in Table 3), where the saturation deficit $(e_s - e_2)$ is expressed in (kPa), and two empirical coefficients (a and b) were evaluated from the series of the evaporation (after the EC method) and the wind speed and air temperature observations done at Maitri site, which is nearest to Lake Zub/Priyadarshini. The daily series for the period lasting from 01.01.2018 to 07.02.2018 was used in the fitting procedure. Figure 6 shows the daily evaporation estimated by the EC method, by the bulk aerodynamic method and new combination equation with two empirical coefficients fitted from the observations.

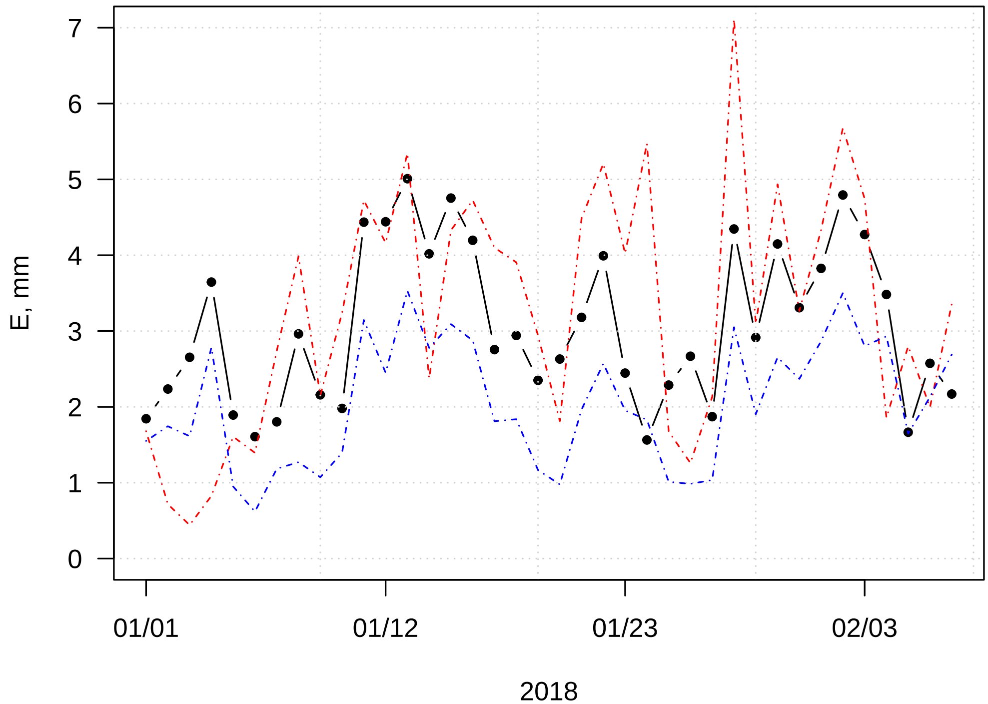

Figure 6. The daily time series of evaporation (mm day^{-1}) calculated by the EC method (black), by the bulk-aerodynamic method (blue) and by new combination equation (red) applying the meteorological measurements at the Maitri site.

The daily evaporation was estimated to be 3.3 ± 1.6 mm day^{-1} (where the numbers represent the mean and standard deviation, respectively) by the equation $E = -0.33 + 0.60w_2 (e_s - e_2)$; and sum of the evaporation for the period 38 days by this method differs for less than 10 % from those estimated by the EC method. It is the lowest difference for the indirect methods considered; the Peason correlation coefficient and the mean root square standard error are estimated to be 0.59 and 1.0, respectively. These scopes allow us to consider this equation the second best among the indirect methods (Table 3), the only bulk aerodynamic method showing the better scope. However, these estimates are done on the similar data as the empirical coefficients were fit, and the independent data are needed.

The efficiency of the empirical formula $E = -0.33 + 0.60w_2 (e_s - e_2)$ with the independent data was estimated from the wind speed and air temperature measured at Irgason site (Fig. 1 c). We also used the lake water surface temperature measured at iBuntton site for the period of 27.01.2018 – 07.02.2018 (or 12 days); the daily series of the evaporation were calculated with this formula and then they were compared with those estimated after the EC method. The Peason correlation coefficient and the mean root square standard error are estimated to be 0.68 and 1.3, respectively. The sum of the evaporation for the period

12 days by this method is over 30 % higher than those estimated by the EC method. It shows that new combination formula may tend to overestimate the evaporation.

4.2 Impact of katabatic winds on evaporation

The study region is dominated by winds from the south-easterly sector (Fig. 3 b). This corresponds to the katabatic winds, which the Coriolis force has turned left from the direct down-slope direction. To better understand the impact of katabatic winds, we carried out further analyses on the wind conditions in the study region. We calculated the geostrophic wind fields for each day of the study period from the mean sea level pressure fields estimated from the ERA5 reanalysis. The results demonstrated that the geostrophic (synoptic) wind was mostly from the east, i.e., some 45 degrees right from the mean direction of the observed near-surface wind. This deviation angle may partly result from Ekman turning in the atmospheric boundary layer, which over an ice sheet with a rather small aerodynamic roughness may contribute some 20 degrees, and from the katabatic forcing. In any case, in most cases the observed near-surface winds resulted from the combined effects of synoptic and katabatic forcing, which supported each other. Hence, it is very difficult to robustly distinguish the impact of katabatic forcing on the near-surface winds over the lake.

However, the geostrophic wind direction was distinctly different, 240 – 350°, in the following days: 6, 8 – 10, 19 and 25 – 27 January. These days were related to transient cyclones centred north-west of the lake or high-pressure centres north-east of the region under study. During the days, the wind speed over the lake was strongly reduced (Table 5), as the katabatic and synoptic forcing factors opposed each other. The lake surface temperature was higher than usual, but the air temperature was lower. The latter is partly because, during events when the geostrophic and katabatic forcing factors support each other (sector 60 – 130º), the strong wind effectively mixes the atmospheric boundary layer. In stably stratified conditions, which prevail over the ice sheet, vertical mixing results in higher near-surface air temperatures (Vihma et al., 2011). In addition, adiabatic warming during the downslope flow is a major factor contributing to higher air temperatures (Xu et al., 2021). The impact of adiabatic warming is also seen as lower relative humidity in cases when the geostrophic wind is from the sector 60 – 130°. Related to the compensating effects of air temperature and relative humidity, the specific humidity was not sensitive to the geostrophic wind direction. The effect of wind speed dominated the effect of the lake surface temperature (which controls q_s in Eq. (q)), and evaporation was strongly reduced when the geostrophic wind was from the sector 60 – 130° (Table 5).

Table 5. The mean values of evaporation (E_{EC}), wind speed (w_2), air specific humidity (Q_2), surface temperature (w_t), and air temperature (t_2) calculated over the days when the geostrophic wind direction was 60 – 130º and when it was 240 – 350º.

Geostrophic wind dir.	Evaporation (mm day^{-1})	w_2 (m s^{-1})	Q_2 (g kg^{-1})	t_w (ºC)	t_2 (ºC)
60 – 130º	3.1	6.9	2.0	3.6	–0.2
240 – 350º	1.3	2.3	2.0	4.8	–2.8

The katabatic wind was a quasi-persistent feature during the study period, and the major changes in the evaporation were driven by changes in the synoptic scale wind direction, which affected the local wind speed.

5 Discussion

Our study yielded estimates of evaporation over a glacial lake in the summer based on direct EC measurements during a field experiment lasting 38 days. These direct estimates of evaporation were considered as the reference when estimating the uncertainties inherent in the indirect methods including the bulk-aerodynamic method and four combination equations. The results based on the bulk-aerodynamic method reached the best skill scores based on the efficiency indexes, however, this method underestimated the daily evaporation by over 30 %. The efficiency of all selected combination equations was low: they underestimated the mean daily evaporation by up to 72 %. The empirical coefficients for the combination equation were fitted from the series of the evaporation (by EC method) and the meteorological observations at the station nearest to the lake site. This combination equation can be potentially used in estimations of the evaporation over the ice-free glacial lakes located in Schirmacher oasis. However, in this study the estimations of the daily evaporation and efficiency indexes were performed on the same data for the experiment (lasting 38 days). Also, we estimated the efficiency using the independent data on the air temperature, wind speed and lake surface temperature. The estimations of efficiency indexes were also done with the full independent data including the evaporation estimated by the EC method, therefore we would not suggest applying these coefficients as the regional references without further analysis. In this study, we did not estimate the evaporation using the energy balance method, but plan to further evaluate the uncertainties inherent also in this method while estimating the evaporation over the glacial lakes located in Antarctica.

At monitoring sites, evaporation over lakes is in practice measured with evaporation pans, which are not fully applicable in polar regions. The EC measurements require specific equipment not always possible to deploy and operate in the remote Antarctic continent. Hence, evaporation (or sublimation) over lakes is usually estimated only indirectly on the basis of regular or campaign observations or numerical model experiments. There are only a few studies of evaporation over lakes located in Antarctica. Borghini et al. (2013) propounded estimates of evaporation over a small endorheic lake located on the shore of Wood Bay, Victoria Land, East Antarctica (70º S). This lake is of 0.8 m depth, and by early 200s its surface area has decreased to half of the value in late 1980s. The lake is of the landlocked type, and Borghini et al. (2013) used the method by Shuttleworth (1993) to estimate the evaporation from the lake surface during a couple of weeks in December 2006. Thy estimated the mean daily evaporation as 4.7 ± 0.8 mm day^{-1}; and such an evaporation rate results in loss of over 40 ± 5 % of the total volume of the lake during the observation period. The lake studied by Borghini et al. (2013) differs from Lake Zub/Priyadarshini, but the daily evaporation rates are of the same order of magnitude, and although one caould expect a much larger evaporation from landlocked lakes than glacial lakes. Our results show that method by Shuttleworth (1993) underestimates the evaporation of lakes located in the Schirmacher oasis by over 60 %.

Shevnina and Kourzeneva (2017) used two indirect methods to evaluate daily evaporation for two lakes located in the Larsemann Hills oasis, East Antarctica (69º S). Lake Progress and Lake Nella/Scandrett are of the glacial type, however, they are much deeper and larger in volume than Lake Zub/Priyadarshini, and over 30–70 % of their catchments are covered by the glacier. The thermal regime of these glacial lakes is also different: Lake Nella/Scandrett and Lake Progress have partially lost their ice cover in austral summers when their surface water temperature is only 2–3 °C, which is lower than the water temperature over the surface of Lake Zub/Priyadarshini. The daily evaporation was estimated to be 1.8 mm day^{-1} and 1.4 mm day^{-1} on the basis of the energy budget method (Mironov et al., 2005) and by the equation of Odrova (1979), respectively. Shevnina and Kourzeneva (2017) concluded that daily evaporation over glacial lakes is underestimated by both of these indirect methods. Our results prove that the uncertainties inherent in the method by Odrova (1979) are the largest among other considered methods.

Faucher et al. (2019) evaluated the annual cycle of the terms of the water balance equation written for Lake Untersee, Dronning Maud Land, East Antarctica (71° S). Lake Undersee is perennially frozen year-round; it is the glacial type lake directly attached to the continental ice sheet, not being the landlocked.. The sublimation (evaporation) over the lake surface was estimated in terms of its water budget. These estimations were based on two years of in situ measurements using snow sticks. Faucher et al. (2019) estimated the water losses from the ice-covered surface of the lake due to sublimation to be from 400 to 750 mm $year^{-1}$. The daily evaporation from the lake surface was approximately 1.1–2.1 mm day^{-1}. Dhote et al. (2021) estimated the summertime evaporation over Lake Zub/Priyadarshini using the meteorological observations collected at the Maitri site. The sum of the evaporation over the lake surface was estimated to be 167 mm for January and February 2018 (or 2.8 mm day^{-1}), and this estimate is very close to those based on the EC method given in this study.

This study focused on the evaporation over a glacial lake located in the Schirmacher oasis, East Antarctica. Over 65 thousand glacial lakes have been detected in the coastal region via satellite remote sensing in austral summer 2017, and most of them spread over the ice shelf and the margins of the continental ice sheet (Stokes et al., 2019). The total area of glacial lakes in vicinity of the Schirmacher oasis was over 72 km^2 in January 2017 (Fig. 7), and the two largest glacial lakes being of a similar size as the Schirmacher oasis itself. During warm periods, a high number of glacial lakes (or melt ponds) are recognized over the margins of the Greenland ice sheet (How et al., 2021), and melt ponds are very common also on the surface of Arctic sea ice (Lu et al., 2018). The glacial lakes may exist over the snow/ice covered surface for 1 – 3 months, and their presence has changed land cover properties and affected the surface heat budget. A proper description of land cover is a crucial element of numerical weather prediction (NWP) and climate models, where the overall characteristics of land cover are represented by the surfaces covered by ground, whether vegetation, urban infrastructure, water (including lakes), bare soil or other. Various parameterization schemes (models) are applied to describe the surface-atmosphere moisture exchange and surface radiative budget (Viterbo, 2002). Lakes have been recently included in the surface parameterization schemes of many NWP models (Salgado and Le Moinge, 2010; Balsamo et al., 2012) with known external parameters (location, mean depth) available from the Global Lake Database, GLDB (Kourzeneva, 2010). The newest version of the

GLDB includes glacial lakes in Antarctica (Toptunova et al., 2019). In future studies, it is important to understand how glacial lakes affect the regional air moisture transport over the polar regions and local weather.

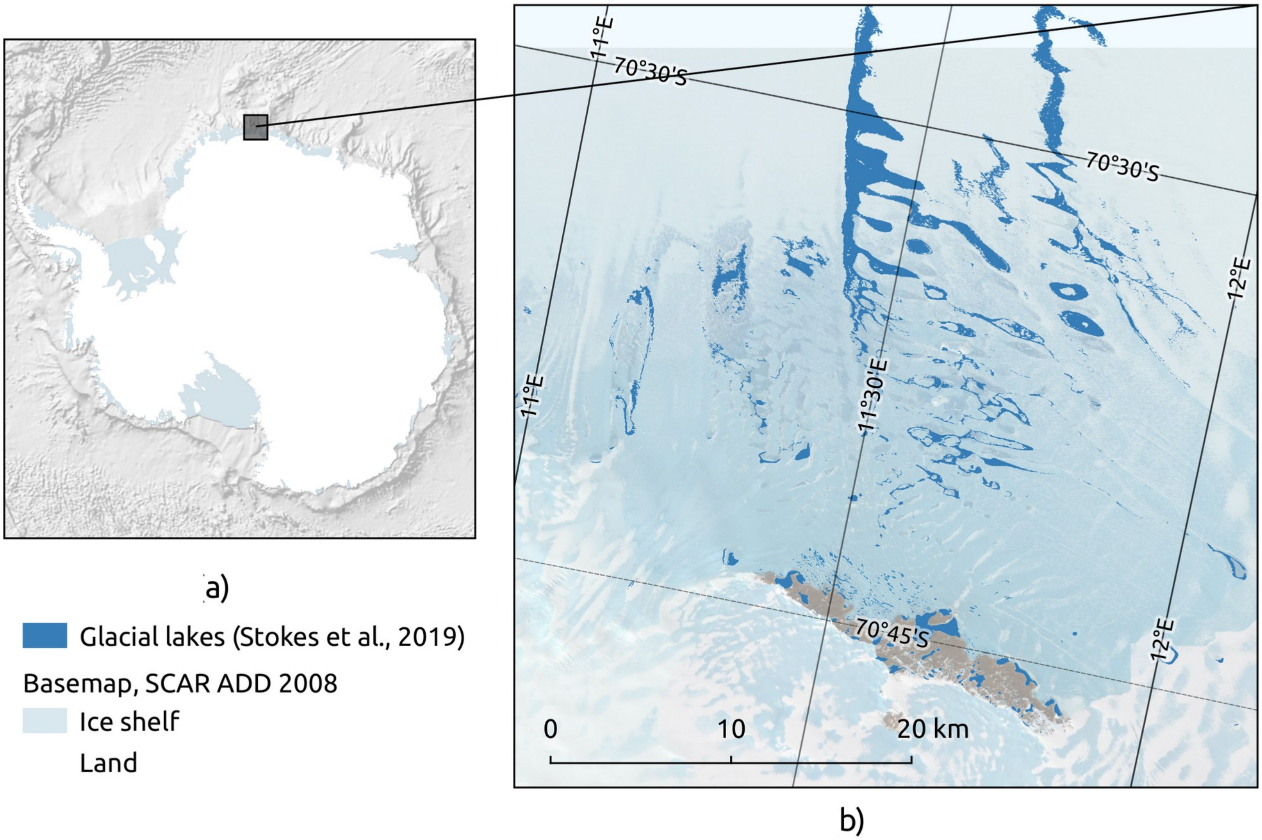

Figure 7. The glacial lakes over the surface of an ice shelf in the vicinity of the Scirmacher oasis, East Antarctica.

Estimates of evaporation are available from atmospheric reanalyses which share results of simulations done by NWP models. As for other reanalyses, ERA5 does not assimilate any evaporation observations, and the evaporation is based on 12-hour forecasts of an NWP model by applying the bulk-aerodynamic method. The results naturally depend on the presentation of the Earth’s surface in ERA5, and in Dronning Maud Land, the surface type is ice and snow with no lakes. Therefore, the estimate of evaporation does not include evaporation from liquid water surfaces. We estimated the daily evaporation also from ERA5, and the results suggest that the evaporation during summer (December – February) 2017 – 2018 was 0.6 mm day^{-1}. This is only one fifth of the evaporation estimated with the direct EC method.

Naakka et al. (2021) estimated the evaporation over the Antarctic region from the ERA5 reanalysis for five domains, including the East Antarctic slope where the Schirmacher oasis is located. There the average daily evaporation in summer is 0.3 mm day^{-1}, and this is reasonable for the ice/snow covered surface. In summertime, the presence of liquid water over ice/snow covered surface changes the fraction of lakes over the East Antarctic slope, and it is 6–8 % of the region in the vicinity of the Schirmacher oasis (Fig. 7). The increasing numbers of glacial lakes over the surface of the East Antarctic slope affects the surface-atmosphere moisture interactions, and it also changes the regional evaporation not accounted for by the numerical weather prediction systems and climate models. We assumed that the 0.3 mm of ERA5 is a fair value for the ice sheet in the East Antarctic slope and that 3 mm is a representative value for glacial lakes, and it may add up to 0.16–0.22 mm to the regional summertime evaporation over the margins of the East Antarctic slope. These numbers seem to be insignificant for the mass balance of the Antarctic ice sheet and ice shelves. However, we suggested more research to better understand the impact of glacial lakes on the surface heat budget and atmospheric moisture transport in the summer.

6 Conclusions

This study suggested the estimates of summertime evaporation over an ice-free surface of Lake Zub/Priyadarshini applying the direct EC method. Evaporation was also evaluated using six indirect methods only needing as input a few hydrometeorological parameters monitored at selected sites (e.g., WMO stations). The catchment Lake Zub/Priyadarshini has less than 30 % of its area covered by glaciers, which results in a specific thermal regime and water balance for the lake. We estimated the evaporation over ice free lake surface as 114 mm in the period from 1 January to 7 February 2018 on the basis ofthe EC method. The evaporation was estimated to be 3.0 mm per day in January 2018. The largest changes in daily evaporation were driven by synoptic-scale atmospheric processes rather than local katabatic winds.

This study gave the estimations of the uncertainties inherent in the indirect methods applied to evaluate summertime evaporation over a lake surface. The bulk aerodynamic method suggests the average daily evaporation to be 2.0 mm day^{-1}, which is 32 % less than the result based on the EC method. Four selected combination equations underestimated the evaporation over the lake surface by over 40–72 %. We suggested a new combination equation to evaluate the summertime evaporation of Lake Zub/Priyadarshini from meteorological observations from the nearest site. The performance of the new equation is better than the performance of the indirect methods considered. We stress the need for accurate measurements of the lake water surface temperature to allow better estimates of lake water budget and evaporation (sublimation).

The evaporation results were not sensitive to differences in the data collected at the meteorological site nearest to the lake and the site located on the lake shore. Hence, we suggest using the synoptic records at the meteorological site Maitri to evaluate the evaporation over the surface of Lake Zub/Priyadarshini. Field experiments are needed to make analogous comparisons of meteorological conditions between other glacial lakes and the permanent observation stations nearest to them. The water balance terms of glacial lakes (including evaporation) are closely connected to their thermal regime and coupled thermophysical and hydrological models are needed to predict the amount of water in these lakes. Our results also demonstrated the need to present glacial lakes in atmospheric reanalyses as well as NWPs and climate models. Ignoring them in a lake-rich region, such as the Schirmacher oasis, results in a large underestimation of regional evaporation in the summer.

Annex.

To evaluate the uncertainties of the EC method with the paired tower method : the intercalibration experiment at Alqueva reservoir, Portugal.

The eddy covariance method has some errors and uncertainties associated with the nature of the measurement and the instrument system. Therefore, the results need to be treated with special attention. Nevertheless, the complexity of the method, namely the filters and corrections that this method requires (see Section 3.3), make it possible to reduce the errors and uncertainties. According to Aubinet et al. (2012), there are three methods to quantify the total random uncertainty for the eddy covariance method: the paired tower, 24 h differencing, and the model residual. In our study we apply the paired tower method to evaluate the errors of the Irgason installed on the shore of Lake Zub/Priyadarsini. The intercalibration experiment lasted from 12 October to 25 October 2018, and during this period two Irgason instruments were deployed on a floating platform in Alqueva artificial lake located southeast of Portugal.

The floating platform (38.2° N; 7.4° W) has been operating continuously since April 2017, and in this experiment, two eddy covariance stations (Irgason) were installed on the height of 2.0 m next to each other facing the same footprint (Fig. A1). In this experiment, we compare the measurements of the Irgason of the Finnish Meteorological Institute (FMI) to those collected by the Irgason of the Institute of Earth Sciences (ICT), University of Évora. Taking advantage of the fact that both instruments are identical, the settings were set exactly the same. The standard gas zero and span calibration was performed before the experiment. The raw measurements from both instruments were post-processed applying the algorithm given in Potes et al. (2017). It allows precise estimates of random instrument uncertainty, rather than total random uncertainty which demands that both instruments are in the same area but with different footprints (Dragoni et al., 2006).

Figure A1: The instruments installed in Alqueva reservoir (Portugal) for the intercalibration. The left instrument belongs to the Institute of Earth Sciences, and the instrument on the right belongs to the Finnish Meteorological Institute.

Figure A2 shows a scatter plot between 30-minute evaporation evaluated from the measurements of two instruments during the intercomparison campaign that took place in Alqueva reservoir. The correlation coefficient between the evaporation calculated by two Irgasons is over 0.98, and it suggests strong agreement between the measurements. Figure A3 presents the frequency distribution of the 30-minute evaporation random instrument uncertainty (ε_F) during the intercomparison campaign (see the Eq. 9 from Dragoni et al., 2007). The random instrument error in 30-minute evaporation, estimated as the standard deviation of the evaporation random instrument uncertainty (ε_F), is 0.004324 mm. Thus, in relative terms, the intercomparison campaign allows obtaining an estimate of a random instrument error of 7.0 %. This value is below other studies presented by several authors, namely: Eugster et al. (1997), that used the same approach of the paired towers in Alaskan tundra, and obtained 9 % for latent heat flux; Finkelstein and Sims (2001), that present a value between 14 and 35 % for latent heat flux in forest and agricultural sites; and Salesky et al. (2012), that found typical errors of 10 % for heat flux.

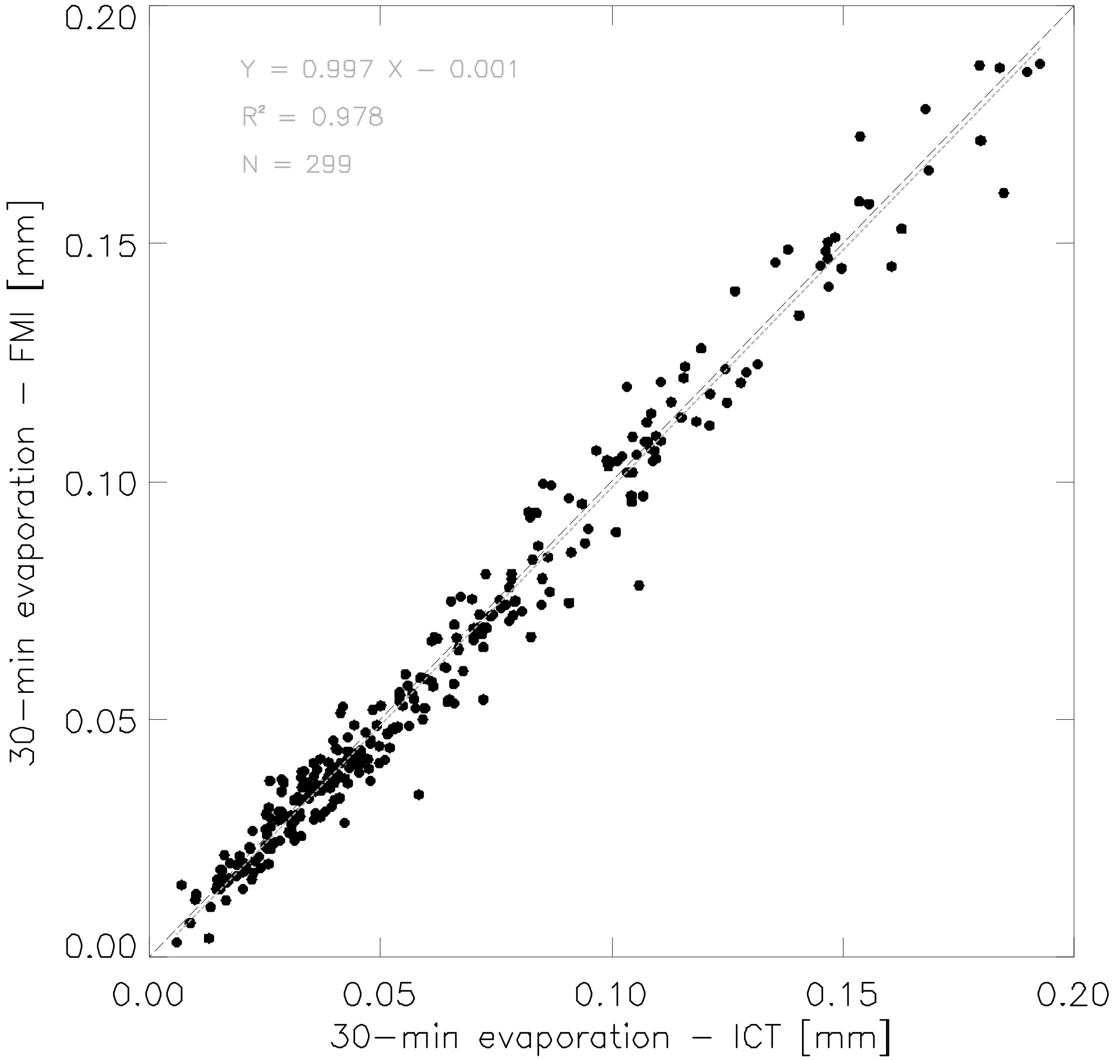

Figure A2: Scatter plot between 30-minute evaporation from both instruments: the Y-axis shows the values estimated after the measurements by the Irgason of the FMI, and the X-axis shows the values after the measurements of the Irgason of the ICT.

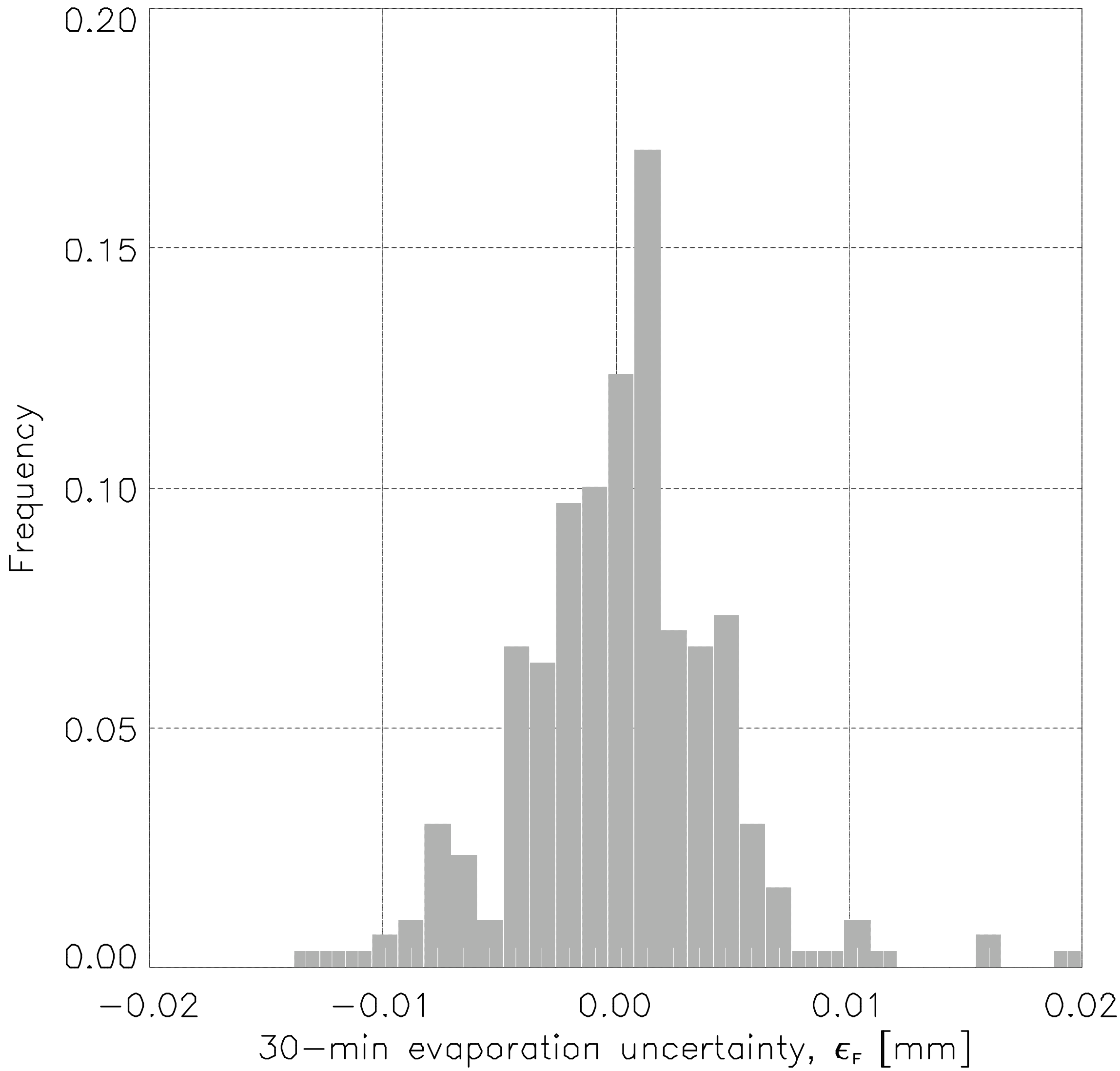

Figure A3: Frequency distribution of the 30-minute evaporation random instrument uncertainty (ε_F).

Acknowledgement. The study was funded by the Academy of Finland (contract number 304345), the H2020 project PolarRES (contract 101003590) and the COST Snow Action ES1404. The measurement campaigns were supported by the Finnish Antarctic Research Program, the Russian Antarctic Expedition, and the Indian Antarctic expedition. We thank Daniela Franz, Ekaterina Kourzeneva and Rui Salgado for the discussions during the 5th and 6th Workshops on Parameterization of Lakes in Numerical Weather Prediction and Climate Modelling (October 2017, Berlin, Germany, and

October 2019, Toulouse, France). We thank the participants of two scientific conferences held in St. Petersburg in 2020 and Lisbon (on-line) in 2021 for their questions and comments. Our special thanks to Alexander Piskun, who shared with us his private collection of Bulletins of the Soviet Antarctic Expedition (1960–1986). We are grateful to Indian Space Research Organisation and National Centre for Polar and Ocean Research, Goa, for constant support and encouragement during this work. We thank the editor and two anonymous referees for their fruitful comments and suggestions allowing the improvement of the presentation of our results.

Supplement. ES attached the calculation of the evaporation with the combination equations in Table 3 (combination_equations_results.csv) and the code (results_code.r). MP has attached the post processed by the EC method (20180101_20180207_EC_FLUX.txt). TV attached the calculations by the bulk aerodynamic method (Bulk_method_results_Irgason_input.txt and Bulk_method_results_Maitri_input.txt). PD attached the meteorological data measured at the Maitri site (Meteorological_Parameters_Summer_2017-18.xlsx). TN provided the series of the daily evaporation from the ERA5 reanalysis at the grid note nearest to the Novo meteorological site (Evaporation_Schirmacher_Oasis_from_ERA5.csv). The supplement related to this article is available online at:

Data and code availability. The data and code used in this study are available in the Supplement. We also used two datasets stored at zenodo: http://doi.org/10.5281/zenodo.3469570 and http://doi.org/10.5281/zenodo.3467126.

Author's contribution. ES collected the data in the field experiment 2017–2018, and calculated the evaporation applying the combinational equations, their uncertanties and efficiency indexes. MP supervised the EC measurements in the field, then he calculated the evaporation applying the EC method, and he analysed the data collected during the intercalibration campaign.TV contributed to the estimations of evaporation applying the bulk aerodynamic method. TN contributed with analyses of evaporation based on ERA5. TV and TN made the analysis of the impact of the katabatic winds. PD and PKT contributed with the analysis of the meteorological observations at the Maitri site. All authors contributed to writing of the manuscript.

Competing interests. The authors declare that they have no conflicts of interest.

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