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ORIGINAL PAPER

Evaluation of ten methods for estimating evaporation in a small high-elevation lake on the Tibetan Plateau

Binbin Wang[1,2,3,4] · Yaoming Ma[1,2,4] · Weiqiang Ma[1,2] · Bob Su[3] · Xiaohua Dong[5]

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Abstract

To quantify lake evaporation and its variations in time, ten methods for estimating evaporation at a temporal resolution of 10 days over a small high-elevation lake in the Nam Co lake basin of the Tibetan Plateau (TP) were evaluated by using eddy covariance (EC) observation-based reference datasets. After examination of the consistency of the parameters used in the different methods, the ranking of the methods under different conditions are shown to be inconsistent. The Bowen ratios derived from meteorological data and EC observations are consistent, and it supports a ranking of energy-budget-based methods (including the Bowen ratio energy budget, Penman, Priestley-Taylor, Brutsaert-Stricker and DeBruin-Keijman methods) as the best when heat storage in the water can be estimated accurately. The elevation-dependent psychometric constant can explain the differences between the Priestley-Taylor and DeBruin-Keijman methods. The Dalton-type methods (Dalton and Ryan-Harleman methods) and radiation-based method (Jensen-Haise) all improve significantly after parameter optimization, with better performance by the former than the latter. The deBruin method yields the largest error due to the poor relationship between evaporation and the drying power of the air. The good performance of the Makkink method, with no significant differences before and after optimization, indicates the importance of solar radiation and air temperature in estimation of lake evaporation. The Makkink method was used for long-term evaporation estimation due to lack of water temperature observations in lakes on the TP. Lastly, long-term evaporation during the open-water period (April 6 to November 15 from 1979 to 2015) were obtained; the mean bias was only 6%. A decreasing-increasing trend in lake evaporation with a turning point in 2004 was noted, and this trend corresponds to the published decreasing-increasing trend in reference evapotranspiration on the Tibetan Plateau and can be explained by variations in related meteorological variables.

1 Introduction

Lakes and reservoirs occupy more than 3% of the Earth's terrestrial surface (Downing et al. 2006) and approximately 2% of the land surface on the Tibetan Plateau (TP) (Zhang et al. 2014). Due to the significance of lake evaporation in catchment-scale hydrologic processes, water resource management and regional climate modelling, numerous methods for deriving evaporation over wet surfaces have been proposed (Bowen 1926; Brutsaert and Stricker 1979; Dalton 1802; Finch and Calver 2008; Priestley and Taylor 1972; Winter et al. 1995; Yao 2009). The TP hosts a total lake area of approximately 43,000 km^2 at an average elevation of 4000 m above sea level. Due to the harsh natural environment and difficulties in the setup and maintenance of instruments, evaporation over the enormous high-elevation lakes is estimated mostly from pan observations of surrounding areas or by model simulations based on a combination of traditional meteorological data and remote sensing products (Biskop et al. 2016; Haginoya et al. 2009; Ma et al. 2016; Xu et al. 2009; Yu et al. 2011; Zhou et al. 2013). With global climate change, trends of increasing numbers of lakes, growing lakes and rising lake levels are widely observed in the central TP due to increased precipitation, accelerated glacial

✉ Binbin Wang
wangbinbin@itpcas.ac.cn

1 Key Laboratory of Tibetan Environment Changes and Land Surface Processes, Institute of Tibetan Plateau Research, Chinese Academy of Sciences, No. 16, Lincui Road, Chaoyang District, Beijing City 100101, China

2 CAS Center for Excellence in Tibetan Plateau Earth Sciences, Chinese Academy of Sciences, Beijing 100101, China

3 Faculty of Geo-Information Science and Earth Observation, University of Twente, Enschede 7500 AE, The Netherlands

4 University of Chinese Academy of Sciences, Beijing 100049, China

5 College of Hydraulic and Environmental Engineering, China Three Gorges University, Yichang 443002, Hubei, China

melt and permafrost ablation (Lei et al. 2013; Yang et al. 2014; Yao et al. 2004; Zhang et al. 2011). Furthermore, the climate has shown warming and moistening of the air, solar dimming and wind stilling since the 1980s (Yang et al. 2014) and a generally decreasing trend in pan evaporation (Liu et al. 2004), the turning point of which was observed on the TP in 1993 (Wang et al. 2013; Xing et al. 2016). Evaporation from Nam Co Lake has shown an increase in simulations based on the Flake (fresh water lake) model (Lazhu et al. 2016), but a slight decrease in estimations with the Complementary Relationship Lake Evaporation method (Ma et al. 2016). Thus, it becomes paramount that the reliability is established of different methods for estimation of evaporation for long-term evaporation analysis based on direct observational evidence.

Various methods for estimation of evaporation have been developed and applied to wet surfaces worldwide (Dalton 1802; Rosenberry et al. 2007; Singh and Xu 1997; Winter et al. 1995; Yao 2009). These methods mainly include empirical methods, bulk transfer methods, water budget methods, energy budget methods, combination methods and direct observations (Finch and Calver 2008). The principles and disadvantages of these methods are briefly summarised as follows. Simple empirical methods, such as pan evaporation corrected with pan coefficients, are inadequate for widespread application due to the large uncertainties originating from pan types, lake properties and different in environmental and meteorological settings. Other empirical methods relate evaporation to meteorological variables, for example solar radiation, air temperature and wind speed (McGuinness and Bordne 1972). Bulk transfer methods originate from Dalton (1802) and relate evaporation to water vapour differences between the water surface and atmosphere, and evaporation is regulated by a wind function and a bulk transfer coefficient, which can be influenced by lake size and depth and local climatic and environmental conditions (Assouline et al. 2008; Oswald and Rouse 2004; Panin et al. 2006). Water budget methods determine evaporation by conservation of inflow, outflow, precipitation and change in the water level. However, errors inherent in measurements of water budget components, including the difficulty in observing subsurface water exchange, will cause large uncertainties (Finch and Calver 2008). Energy budget methods treat the latent heat flux (LE; with L as the latent heat of vaporisation and E as evaporation) as the residual heat of all other energy components, and assume that incoming solar radiation is the principal source of energy for evaporation (McGuinness and Bordne 1972). The Bowen ratio ($Bo = H/LE$, Bo = H/LE (Bowen 1926)) has been widely used for allocating the ratio of heat loss by conduction to that by evaporation. Combination methods usually combine energy budget methods and bulk transfer methods by eliminating requirement of water surface temperature. However, the key observations of water temperature for Bo and heat storage in the water (Q_x) are always limited worldwide, especially on the TP. Studies of long-term lake evaporation have been conducted over all types of lakes in different climatic environments all over the world (see Table 1 of Yao (2009)), and the Bowen ratio energy budget (BREB) method has been widely chosen as the standard reference method for evaluation of other evaporation methods (Rosenberry et al. 2007; Winter et al. 1995; Yao 2009). In the past several years, direct observation of lake evaporation using an eddy covariance technique has been widely applied, mostly on large lakes of the TP, for analysis of lake-atmosphere interaction processes and evaluation of model simulations (Biermann et al. 2013; Li et al. 2015; Liu et al. 2014; Wang et al. 2015; Wang et al. 2017; Wen et al. 2016). However, due to the numerical dominance of small lakes in this area (Zhang et al. 2014), the adequacy and reliability of the evaporation methods over small high-elevation lakes urgently needs evaluation.

In this study, eddy covariance observations were used at a small lake in the Nam Co lake basin. The lake-air transfer parameters in a bulk aerodynamic transfer model were calibrated for 2012 and then validated independently for 2013 (Wang et al. 2015; Wang et al. 2017). The constructed data series provides a unique reference for the evaluation of evaporation estimation methods (including two radiation-based methods, two D_a-type methods and six energy-budget-based methods). The objectives of this study were (1) to evaluate the performance of these traditional evaporation methods at a temporal resolution of 10 days over a small high-elevation lake and (2) to obtain inter-annual variations in lake evaporation during the open-water period of the small lake during 1979–2015. The data and methods are briefly introduced in Sects. 2 and 3, respectively. The results are given in Sect. 4. A discussion and conclusions are presented in Sects. 5 and 6, respectively.

2 Site and materials

Nam Co lake (Fig. 1a) is in a transitional climate from semi-humid to semi-arid and is influenced by both the Asian summer monsoon and westerlies. The multi-year average air temperature and precipitation over 2007–2011 are approximately 0 °C and 500 mm, respectively (Zhou et al. 2013). The target lake (small Nam Co lake; Fig. 1b) is located 500 m southeast of Nam Co lake. Its area and mean depth are approximately 1.4 km^2 and 7 m, respectively, and the maximum depth is 14 m. Three terraces exist from the level of the observation site in the lake (white circle in Fig. 1b) to the Nam Co station (Nam Co Monitoring and Research Station for the Alpine Environment; black

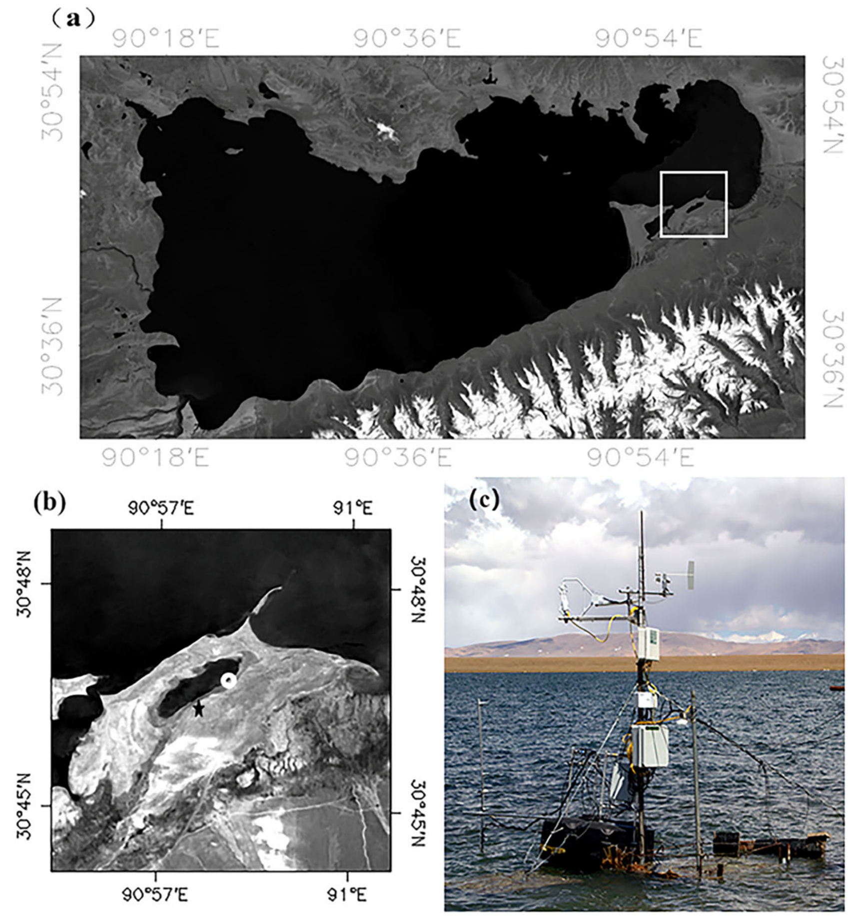

Fig. 1 **a** Nam Co lake and small Nam Co lake (white box). **b** Locations of Nam Co station (black pentagram) and energy budget observation (white circle) in the area of small Nam Co lake. **c** Instruments in small Nam Co lake

pentagram in Fig. 1b), spanning a distance of 900 m and an average slope of 8° (Biermann et al. 2013). The observations were obtained mainly from the instruments of the energy budget observation system in the lake (Fig. 1c). Three components of wind speed, air temperature, air humidity, air pressure and CO_2 were observed at a frequency of 10 Hz using a three-dimensional sonic anemometer (CSAT3, Campbell Scientific, Inc.) and an open-path CO_2/H_2O infrared gas analyser (LI 7550, LI-COR, Inc.). A radiation sensor (CNR4, Kipp & Zonen) at a height of 1.5 m provided measurements of downward shortwave radiation (Q_s), downward longwave radiation (Q_a), upward shortwave radiation (Q_r) and upward longwave radiation (Q_w). Observations of eddy covariance and the radiation budget were recorded by a data logger (CR5000) every half-hour. Six water temperature sensors were set from the water surface to a depth of 60 cm, and the one close to the water surface was used as the water surface temperature (T_s) considering the influences of waves and lake-level variations. Lake-level variations were observed at a 10-min resolution using a water-level gauge in the small Nam Co lake, whilst variations in the water level of Nam Co lake were measured manually each day during the open-water periods of 2012 and 2013. Precipitation and additional meteorological data were obtained from the Nam Co station (Ma et al. 2009). Additionally, a forcing dataset was produced by the Institute of Tibetan Plateau research (hereafter ITP forcing data) by merging a variety of data sources (He and Yang 2011). The ITP forcing data (including air temperature, specific humidity, downward shortwave radiation and downward longwave radiation) were corrected using site observations from the small Nam Co lake first and were then processed to yield the inter-annual variations in evaporation at a temporal resolution of 10 days during the open-water period of each year from 1979 to 2015.

The latent heat flux (LE) and sensible heat flux (H) were obtained through standard processing of eddy covariance data, and a bulk aerodynamic transfer model was optimised for interpolation of LE and H in case of an inadequate footprint and malfunctioning of the instruments (Wang et al. 2015; Wang et al. 2017). The meteorological variables and energy budget components were subject to strict selection. The heat storage in the water is assumed to be the residual of the energy budget components ($Q_x = R_n - LE - H$). The temperature and vapour pressure over the water surface were higher than those in the air (Fig. 2a, b), and the average

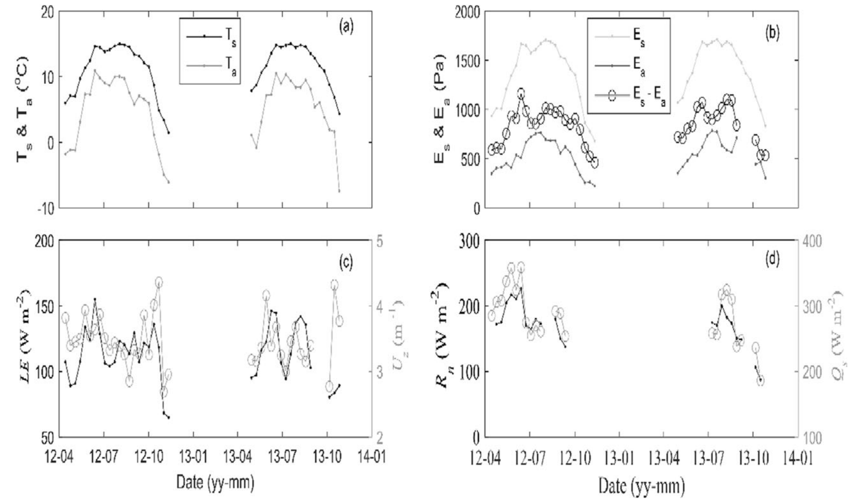

Fig. 2 Variations in **a** temperature at the water surface (T_s) and in the air (T_a). **b** Vapour pressure at the water surface (E_s) and in the air (E_a), and vapour pressure difference ($E_s - E_a$). **c** *LE* and wind speed (U_z). **d** Net radiation (R_n) and downward shortwave radiation (Q_s) at a temporal resolution of 10 days

wind speed was 3.46 m s^{-1} (Fig. 2c). An unstable and neutral atmosphere, which has been reported at several high-elevation lakes (Li et al. 2015; Liu et al. 2014; Wen et al. 2016), dominates the small Nam Co lake. The solar radiation in May and June of 2013 was disregarded because of stringent quality criteria (Fig. 2d). The evaporation throughout the ice-free period was approximately 812 mm, and the energy budget was generally closed with a closure ratio of 0.97 (Wang et al. 2017).

Precipitation, evaporation and water-level changes are the dominant factors of the water budget in small Nam Co lake (Fig. 3), ignoring the unobserved subsurface inflow and outflow. The lake level dropped from early April (ice-melt period) to the end of June in both years, at which time the precipitation minus evaporation was generally smaller than 0 (Fig. 3a, b). Afterwards, the lake level rose approximately 80 mm in 2012 and approximately 200 mm in 2013 due to monsoon precipitation extending into October. The total amounts of precipitation were 436.7 and 487.9 mm in 2012 and in 2013, respectively (Fig. 3c). The changes in the lake level closely tracked the precipitation minus evaporation ($Prec_i - E_i$) as shown in Fig. 3b; thus, precipitation is the main source of water supply in small Nam Co lake. The level of Nam Co lake increases every year until the end of September and then decreases afterwards. Because the rate of lake-level change is much lower in the small lake than in Nam Co lake (Fig. 3a), no direct water exchange exists between the two water bodies.

3 Methods

3.1 Methods for estimation of evaporation

Ten evaporation methods were evaluated and ranked using observations at a temporal resolution of 10 days (Table 1). The Penman (1948) method (PE) has been widely used for estimation of evaporation and combines equilibrium evaporation (a lower limit evaporation from moist surfaces) and a measure of the departure from equilibrium in the atmosphere (in other words, the drying power of the air). When the air over a moist surface is vapour-saturated and the variations in temperature and vapour pressure with height and time are small, the Priestley and Taylor (1972) method (PT) is used to take the equilibrium evaporation for estimating potential evaporation under a condition of minimal advection; the Priestley-Taylor constant is $\alpha = 1.26$. The Brutsaert and Stricker (1979) method (BS) is based on Bouchet's hypothesis ($E = 2E_w - E_p$), with E_w estimated by the PT method and E_p estimated by the PE method. Similarly, the De Bruin (1978) method (de) eliminates the energy budget part in the PT and PE methods and relates evaporation to the drying power of the air. The Bowen-ratio-energy-budget method (BREB) was first presented by Bowen (1926) and has been widely used for estimation and validation of evaporation over wet surfaces (Brutsaert 1982; Drexler et al. 2004; Rosenberry et al. 2007; Winter et al. 1995). The Bowen ratio (*Bo*) is expressed as $Bo = \gamma \frac{(T_s - T_a)}{(E_s - E_a)}$ through observations of temperature and vapour pressure at two heights. Note that the energy due to subsurface inflow and outflow and heat transfer in the sediment are ignored in our study due to limitations in the observational data. The Bruin and Keijman (1979) method (DK) introduces an observational relationship between $\frac{\gamma}{s}$ and *Bo* from Hicks and Hess (1977), where γ is a psychrometric constant and s is the slope of the saturated vapour pressure-temperature curve at mean air temperature (Allen et al. 1998). Many other formulas exist for estimation of evaporation based on only air and water parameters (McGuinness and Bordne 1972; Singh and Xu 1997). These methods have various forms and constants that are site-dependent and require calibration. Because model complexity cannot improve simulation

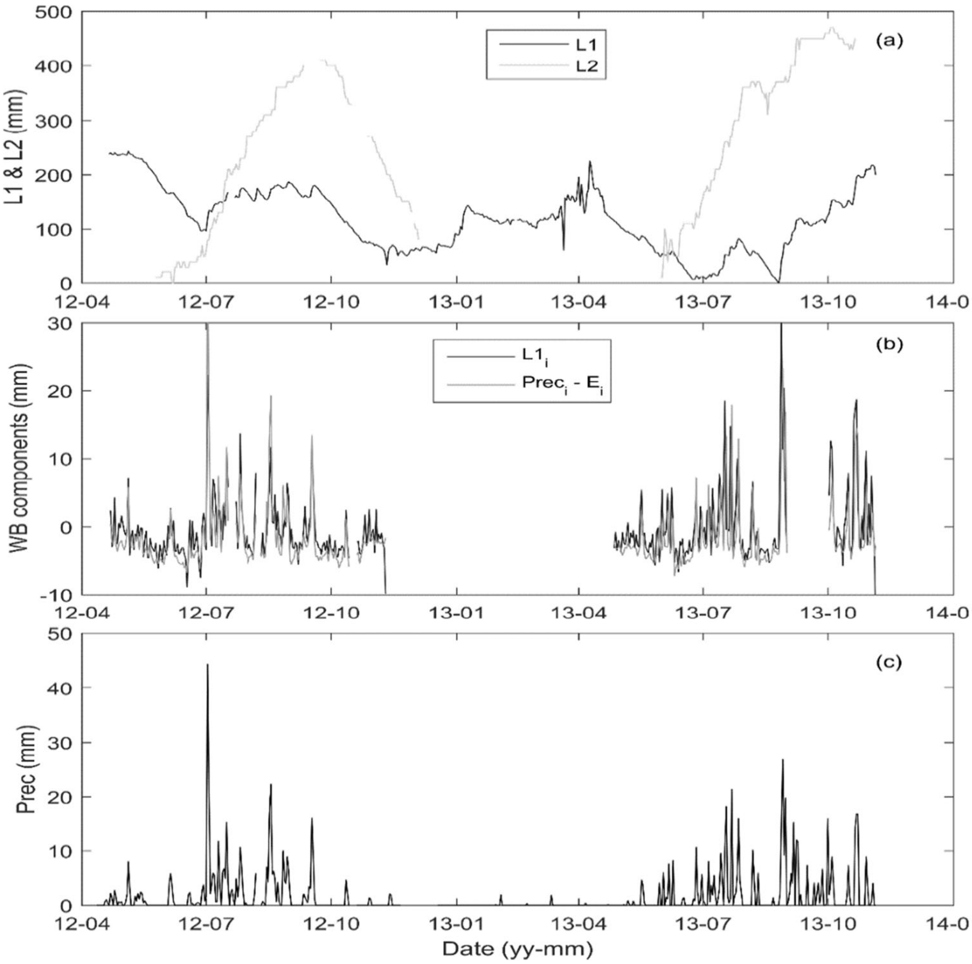

Fig. 3 **a** Variations in daily water-level changes in the small Nam Co lake ("L1") and Nam Co lake ("L2"); note that the reference water levels are different. **b** Variations in water balance ("WB") components in small Nam Co lake: daily water level ("$L1_i$") and difference between daily precipitation ("$Prec_i$") and daily evaporation ("E_i"). **c** Variations in daily precipitation ("Prec")

performance (Singh and Xu 1997; Xu and Singh 2001), two solar-radiation-based methods (the Jensen-Haise and Makkink methods, JH and Mak for short) and two Dalton-type methods (the Ryan-Harleman and Dalton methods, RH and DM for short) were chosen in our study. The input meteorological variables for the methods are average values over 10 days, and the outputs of the evaporation simulations were converted to mm d^{-1} using relevant constants. All the equations and variables are listed and explained in Table 1.

3.2 Methods for calibration and validation

The constants in several of the abovementioned methods can be calibrated using observations, including de_1 and de_2 in the de method, JH_1 and JH_2 in the JH method, Mak_1 in the Mak method, RH_1 in the RH method and N in the DM method (Table 1). Eddy-covariance-based reference datasets and meteorological observations are used to optimise these constants. The PE, PT and BS methods were developed gradually and are widely applied to all types of environments. In addition, the contribution to evaporation from the wind function in the PE and BS methods is small (approximately 21%). Thus, the constants in the PE and BS methods and the well-known Priestley-Taylor constant in the PT method remain unchanged. Similar situations exist for the BREB and DK methods. The performances of these methods can be tested and ranked based on the root-mean-square error (RMSE), Pearson correlation coefficient (R), relative error (RE) and mean bias (MB) in eddy covariance observations. Smaller values of RMSE, RE and MB and higher values of R indicate better performance of a method. The equations for obtaining these statistical measures are as follows:

$$\mathrm{RMSE} = \sqrt{\frac{\sum_{i=1}^{n} |S_i - O_i|}{n}} \tag{1}$$

$$\mathrm{RE} = \frac{1}{n}\sum_{i=1}^{n}\left|\frac{S_i - O_i}{O_i}\right| \tag{2}$$

$$\mathrm{MB} = \frac{1}{n}\sum_{i=1}^{n}(S_i - O_i) \tag{3}$$

where S_i are the simulated results, O_i are the observed results and n is the number of observations.

Table 1 Methods for estimating evaporation (E in mm d^{-1})

Method	Equation
PE	$E=\frac{s}{s+\gamma}\frac{s}{s+\gamma}\times 86.4+\frac{s}{s+\gamma}(0.26(0.5+0.54U_z)\ (E_{as}-E_a))$
PT	$E=\alpha\frac{s}{s+\gamma}\frac{s}{s+\gamma}\times 86.4$
BS	$E=(2\alpha-1)\left(\frac{s}{s+\gamma}\right)\left(\frac{s}{s+\gamma}\right)\times 86.4-\frac{s}{s+\gamma}0.26(0.5+0.54U_z)\ (E_{as}-E_a)$
De	$E=\left(\frac{\alpha}{\alpha-1}\right)\left(\frac{\alpha}{\alpha-1}\right)\frac{\alpha}{\alpha-1}\times 86.4$
BREB	$E=\frac{Q_n-Q_x}{(1+Bo)L_e\rho_w}\times 86.4$
DK	$E=\frac{s}{0.85s+0.63\gamma}\frac{s}{0.85s+0.63\gamma}\times 86.4$
JH	$E=(JH_1T_a+JH_2)(Q_s\times 3.523\times 10^{-2})$
Mak	$E=\left(\left(Mak_1\frac{s}{s+\gamma}\frac{s}{s+\gamma}\right)-0.12\right)$
RH	$E=RH_1\frac{\left(2.7(T_s-T_a)^{0.333}+3.1U_z\right)(E_s-E_a)}{L_e\rho_w}\times 86.4$
DM	$E=(NU_2(E_s-E_a))\times 10$

The multipliers of 86.4 and 10 that appear in equations are used to convert the output into mm d^{-1}

E evaporation (mm d^{-1}); $Bo=\gamma\frac{(T_s-T_a)}{(E_s-E_a)}$ Bowen ratio; γ psychrometric constant (depends on temperature and atmosphere pressure) (Pa$^\circ$C) and $\gamma=\frac{C_pP}{\varepsilon L_e}$, where C_p is specific heat at constant pressure (1.013 MJ kg^{-1} $^\circ$C^{-1}), P is the atmospheric pressure (kPa), $\varepsilon=0.622$ is the ratio of molecular weight of water vapour to that of dry air and L_e is the latent heat of vaporisation (2.45 MJ kg^{-1}); T_s water surface temperature ($^\circ$C); T_a air temperature ($^\circ$C; with unit of $^\circ$F in JH method); s slope of the saturated vapour pressure-temperature curve at mean air temperature (kPa$^\circ$C) and $s=\frac{4098\left[0.6108\exp\left(\frac{17.27T_a}{T_a+273.15}\right)\right]}{(T_a+273.15)}{}^2$; Q_n net radiation (equal to $Q_s-Q_r+Q_a-Q_w$; W m^{-2}); Q_s incoming solar shortwave radiation (W m^{-2}); Q_r reflected solar shortwave radiation (W m^{-2}); Q_a incoming atmospheric longwave radiation (W m^{-2}); Q_w longwave radiation emitted from the water surface (W m^{-2}); Q_x heat storage in the water body (W m^{-2}); ρ_w density of water (998 kg m^{-3} at 20°C); E_{as} *and* E_a saturated vapour pressure (hPa) and actual vapour pressure (hPa), respectively; α Priestley-Taylor empirically derived constant, dimensionless, $\alpha=1.26$; N mass-transfer coefficients, N is 0.0164 as in Rosenberry et al. (2007) and it is 0.0133 after calibration; $de_1=2.9$ and $de_2=2.1$ in the original equation, and they are 21 and -0.47, respectively, after calibration; $JH_1=0.014$ and $JH_2=-0.37$ in the original equation, and they are 0.0045 and 0.2, respectively, after calibration; $Mak_1=52.6$ in the original equation and is 56.5 after calibration; RH_1 are 1 in the original equation and 0.856 after calibration

4 Results

4.1 Evaluation of parameters

The Bowen ratio (Bo) has high significance amongst the energy budget methods for estimation of evaporation, and it has also been related to $\frac{\gamma}{s}$ using the equations (a) $Bo=0.79\gamma/s-0.21$ (PT and BREB) and (b) $Bo=0.63\gamma/s-0.15$ (DK and BREB) (Hicks and Hess 1977). Bo from eddy covariance observations (H/LE) and traditional meteorological observations ($\gamma\frac{(T_s-T_a)}{(E_s-E_a)}$) at a temporal resolution of 30 min is quite close to the 1:1 line (Fig. 4a), which supports the similarity of eddy transfer coefficients for heat and water in atmosphere boundary layer. Furthermore, it shows good consistency using reconstructed heat fluxes and meteorological variables at a temporal resolution of 10 days (Fig. 4b). Thus, energy budget methods based on Bo perform well, as shown in Table 2, whereas the DK method yields higher values of RMSE, RE and MB than does the PT method under condition S1 (condition S1 is explained in Sect. 3.2). The PT method (equation (a)) is more suitable for estimation of evaporation over high-elevation lakes than the DK method (equation (b)); the latter was developed from sea-level environments. When using meteorological observations at small Nam Co lake to develop the relationships between $Bo=\gamma\frac{(T_s-T_a)}{(E_s-E_a)}$ and $\frac{\gamma}{s}$ (Fig. 4c), due to the elevation-dependent variable γ, equation (a) is consistent at the actual air pressure above the lake, whereas equation (b) is more suitable for air pressure at sea level (1013 hPa). Thus, equation (a) in the PT method is more applicable over high-elevation lakes, whereas equation (b) in the DK method is elevation-dependent.

Due to a lack of long-term observations of temperature gradient in the high-elevation lakes on the TP, heat storage in the water (Q_x) needs an alternative expression based on traditional meteorological observations (i.e. net radiation, surface temperature, air temperature). First, Q_x could be related to net radiation (R_n): in 22 lakes described by Duan and Bastiaanssen (2015), the hysteresis effect was clearly present in large deep lakes but not small lakes. A high correlation coefficient of 0.82 between Q_x and R_n was obtained from our measurements (Fig. 5a). Second, increasing/decreasing variations in T_s could describe the accumulation/release of heat storage in the shallow water, and variations in daily surface temperatures over two consecutive days are positively correlated with daily Q_x (Wang et al. 2017). Thus, assuming that variations in mixing-layer

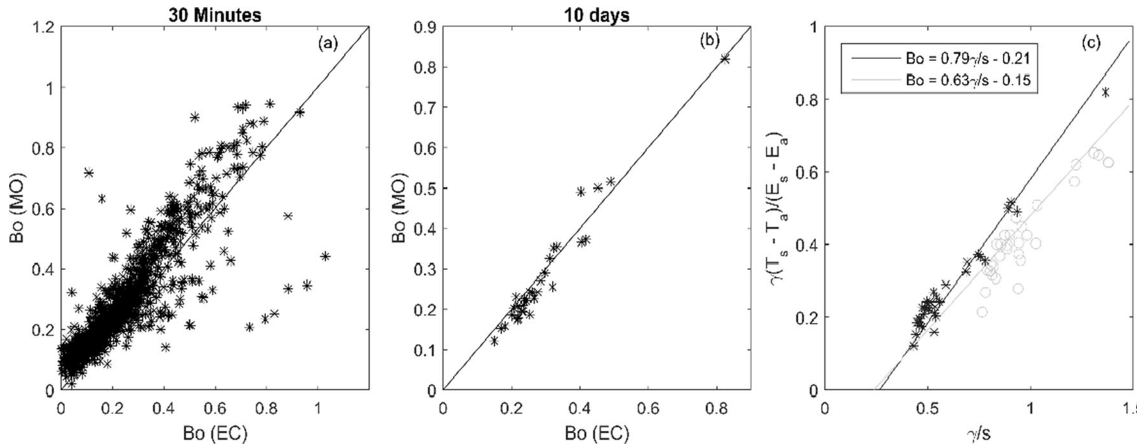

Fig. 4 Scatter of Bo derived from eddy covariance (EC) observations and meteorological observations (MO) at temporal resolutions of **a** 30 min and **b** 10 days. **c** Relationships between Bo and $\frac{\gamma}{s}$ based on actual air pressure (asterisks) and air pressure at sea level (circles)

temperatures (T_s) represent the mean temperature of the entire water column in the small Nam Co lake, an adequate correlation coefficient of 0.78 and a linear fitting line ($y = 15.91x + 29.22$) are obtained between Q_x and the variation in water surface temperature ($\Delta T_s = T_s^{i+1} - T_s^i$, with i and i + 1 denoting the start and end of a period, for example 10 days) (Fig. 5b). Third, considering the strong turbulent mixing between water and air, a correlation coefficient of 0.58 was found between Q_x and ΔT_a (similar to ΔT_s) (Fig. 5c), and this high correlation results from their similar seasonal variations. Overall, the RMSE values for Q_x with respect to R_n, ΔT_s and ΔT_a are 15.3, 16.7, and 21.2 W m^{-2}, respectively, which are smaller than the average RMSE values of 22 W m^{-2} from 22 lakes studied by Duan and Bastiaanssen (2015). Additionally, a high correlation coefficient ($R = 0.72$) exists between Q_s and the available energy ($R_n - Q_x$), which bridges between the Mak method and other energy-budget-based methods.

4.2 Evaluation of evaporation estimation methods

To rank the performances of the evaporation estimation methods, the evaluations were conducted under three conditions (Table 2). In condition S1, the original parameters are used without any optimization and a precisely estimated Q_x (heat storage in the water) is used. In condition S2, the parameters in the de, JH, Mak, RH and DM methods are optimised based on observations, in addition to a precisely estimated Q_x. In condition S3, Q_x is substituted based on its relationship with R_n, in addition to parameter optimization. The BREB method shows the best performance under condition S1 and S2 due to the observed fact that the Bowen ratios derived from the meteorological observations are consistent with those derived from the eddy covariance data (Fig. 4). Amongst all the methods, the de method shows the worst performance, which results from the observed poor correlation between evaporation and the

Table 2 Root-mean-square error (RMSE), correlation coefficient (R), relative error (RE) and mean bias (MB) between observed and simulated evaporation in conditions S1 (original parameters), S2 (calibrated parameters) and S3 (the same as condition S2 except that Q_x is replaced by R_n)

		BREB	PT	PM	BS	DK	Mak	RH	DM	JH	de
RMSE (mm)	S1	0.11	0.19	0.26	0.27	0.27	0.37	0.81	1.10	1.57	1.57
	S2	0.11	0.19	0.26	0.27	0.14	0.36	0.22	0.29	0.35	1.02
	S3	0.42	0.46	0.40	0.59	0.50	0.36	0.22	0.29	0.35	1.02
R ()	S1	0.995	0.98	0.98	0.94	0.99	0.88	0.96	0.94	0.82	0.84
	S2	0.995	0.98	0.98	0.94	0.99	0.88	0.96	0.94	0.87	0.84
	S3	0.80	0.78	0.85	0.57	0.78	0.88	0.96	0.94	0.87	0.84
RE (%)	S1	2.6	4.4	5.7	5.7	6.5	7.9	19.7	25.6	39.2	39.4
	S2	2.6	4.4	5.7	5.7	3.1	7.7	4.8	6.9	8.1	18.8
	S3	9.0	9.8	8.2	13.4	11.6	7.7	4.8	6.9	8.1	18.8
MB (mm)	S1	0.08	0.14	0.17	0.11	0.25	−0.16	0.77	1.02	−1.46	−1.51
	S2	0.08	0.14	0.17	0.11	0.08	−0.12	0.09	0.08	0.03	0.52
	S3	0.08	0.13	0.17	0.10	0.25	−0.12	0.09	0.08	0.03	0.52

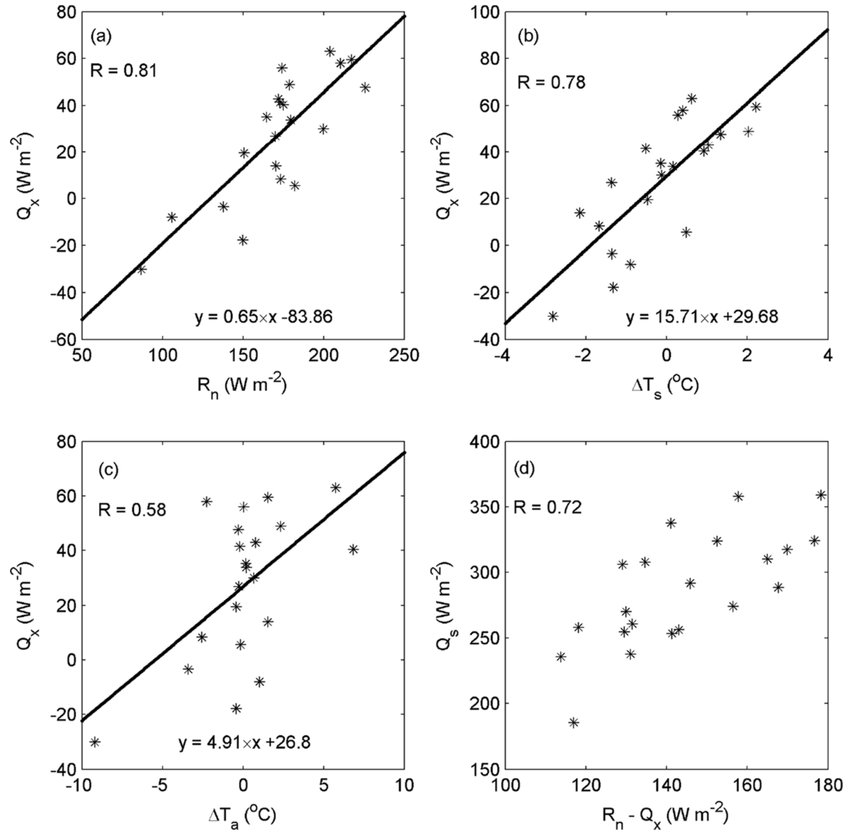

Fig. 5 Scatterplots of **a** Q_x and R_n. **b** Q_x and water surface temperature change (ΔT_s). **c** Q_x and air temperature change (ΔT_a). **d** Q_s and $R_n - Q_x$ at a temporal resolution of 10 days. Correlation coefficients (R) and linear fitting equations are noted

drying power of the air, indicating poor prediction of evaporation without observations of solar radiation and water surface temperature. Moreover, the drying power of the air is only a small modification (21%) to lake evaporation in the PE method. Generally, the energy-budget-based group (BREB, PT, PE, BS and DK methods) under condition S1 performs better compared with the radiation-based group (Mak and JH) and Dalton-type group (RH and DM). The RMSEs, REs and MBs of the energy-budget-based group are smaller than 0.3 mm, 6% and 0.3 mm, respectively. However, when the parameters in the de, JH, Mak, RH and DM methods are optimised based on observations under condition S2, the performances generally improved. The DK method with relatively acceptable results from the relationship in Hicks and Hess (1977) improved slightly with a calibrated relationship between the Bowen ratio and $\frac{\gamma}{s}$ due to the dependence of γ on elevation. The Mak method shows only a small improvement before and after calibration (with relative errors of 7.9 and 7.7%, respectively). The parameters in the JH, RH and DM methods are site-specific and need calibration before being applied in new environments. Overall, the variation in estimated evaporation after calibration matches the observations quite well, with the largest deviation being that of the de method (Fig. 6). An approximate ranking of all the methods under condition S2 from best to worst is as follows: energy-budget-based methods > Dalton-type methods > radiation-based methods.

The energy-budget-based methods and Dalton-type methods after calibration perform better than the radiation-based methods, and the latter show superiority in long-term evaporation analysis due to the difficulty of accessing T_s and Q_x. When Q_x is determined using empirical formulas (for example, through R_n under conditions S3), the energy-budget-based methods lose their superiority over the calibrated radiation-based methods (Table 2). Moreover, the similar performance of the Mak method under all conditions, together with the published good results for a small Canadian lake (Yao 2009) and a small mountain lake in the USA (Rosenberry et al. 2007), suggest good prediction of lake evaporation based on solar radiation and temperature parameters in small and shallow lakes. Thus, the Mak method together with ITP forcing data were used to analyse the variation in evaporation in small Nam Co lake during 1979–2015.

4.3 Inter-annual variations in evaporation

The simulated evaporation amounts (867.5 mm in 2012 and 846.7 mm in 2013) over the period of April 11 to November 6 are very close to the observed average evaporation over the 2 years (812 mm), with a slight overestimation of approximately 6%. The smallest evaporation occurred in 2005, and

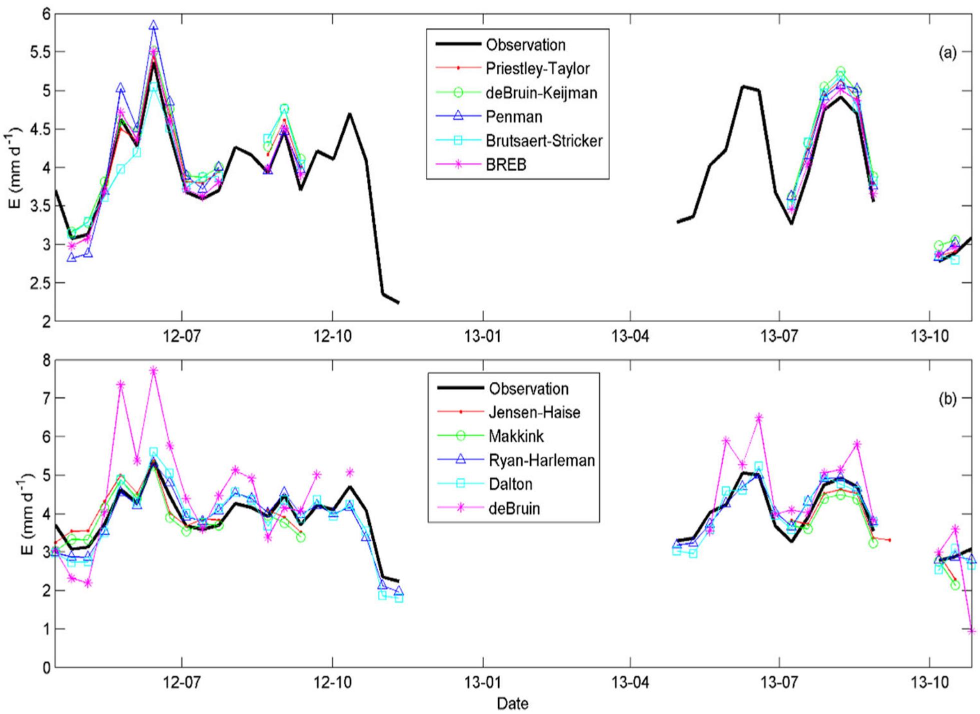

Fig. 6 Variations in observed and simulated evaporation obtained using the calibrated methods of **a** Priestley-Taylor, deBruin-Keijman, Penman, Brutsaert-Stricker and BREB. **b** Jensen-Haise, Makkink, Ryan-Harleman, Dalton and deBruin in 2012 and 2013

the largest evaporation occurred in 2015 (Fig. 7a). The evaporation decreased slightly at a rate of − 0.19 mm yr^{-1} during the 1979–2004 period and increased significantly at a rate of 5.66 mm yr^{-1} during the 2004–2015 period; there was an overall increase of 0.42 mm yr^{-1} during the 1979–2015 period. The overall trend coincides with the trend of evaporation from nearby Nam Co lake based on simulations using the Flake model (Lazhu et al. 2016). Furthermore, our results support the decreasing-increasing (DI) pattern of reference evapotranspiration since the 1980s over the TP (Xing et al. 2016). Even if there was a rapid decrease from 1998 to 2008, which indicates that lake evaporation may have contributed to

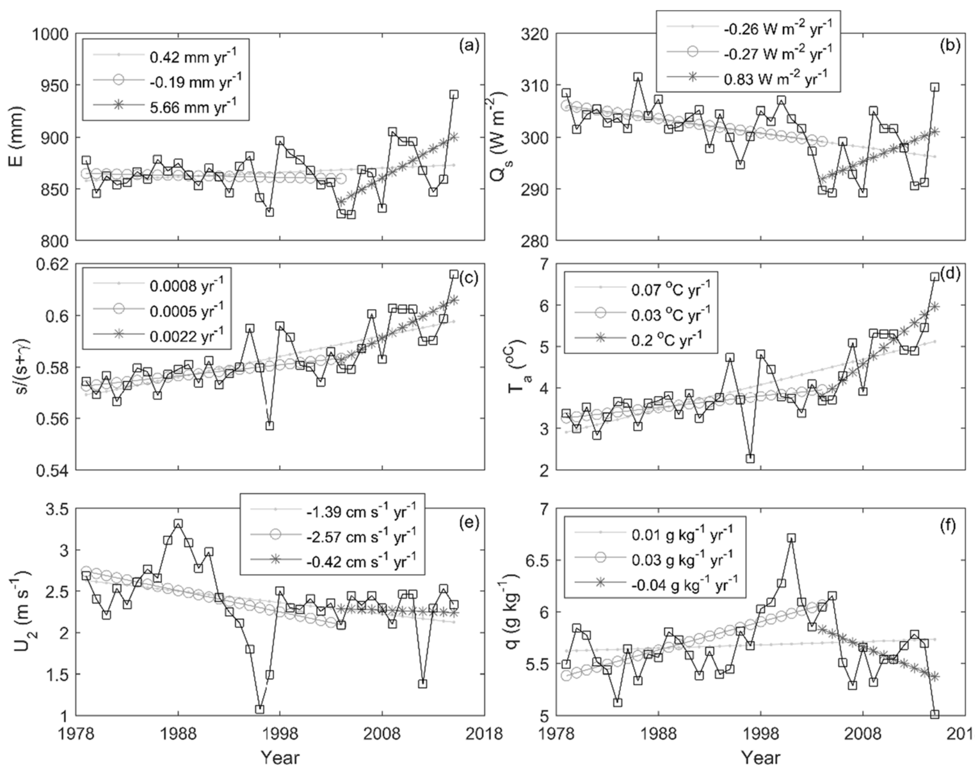

Fig. 7 Variations in **a** evaporation (E), **b** downward shortwave radiation (Q_s), **c** $s/(s+\gamma)$, **d** air temperature (T_a), **e** wind speed (U_2) and **f** specific humidity (q) at a height of 2 m during the ice-free season from 1979 to 2015. Linear trends of the dotted series, circle series and asterisk series for 1979–2015, 1979–2004 and 2005–2015, respectively, are noted in each panel legend

the rapid rises in lake water levels during this period (Lei et al. 2013; Ma et al. 2016), lake evaporation has a negative effect on lake expansion in the long run.

Lake evaporation is obtained from the product of a temperature-related parameter ($\frac{s}{s+\gamma}$) and solar radiation. And the overall trends of increasing evaporation and decreasing solar radiation contradict the intuitive expectation of a decisive influence of solar radiation on evaporation under conditions of unlimited water. Because a clear DI trend of evaporation is centred around 2004, the evaporation and climatic variables are separated into two periods: 1979–2004 (period 1) and 2004–2015 (period 2). This DI trend of evaporation is most likely related to the DI trend in solar radiation (Fig. 7a, b), whilst the monotonically increase in temperature (or temperature-related parameter) could have accelerated the evaporation (Fig. 7c, d). In period 1, the decreasing wind speed and increasing specific humidity (Fig. 7e, f) indicate a smaller advective budget (Hobbins et al. 2004) and could explain the decreasing lake evaporation. The wind speed during period 2 shows a much smaller decrease than that during period 1, and a decrease in specific humidity occurred, which will increase the drying power of the air and increase lake evaporation (Hobbins et al. 2004; Xing et al. 2016).

To quantitatively analyse the contributions of the meteorological variables (air temperature T_a and solar radiation Q_s) to the simulated evaporation obtained from the Mak method, the "trend removal method" of Xu et al. (2006) was used (Fig. 8). The decrease in Q_s and increase in T_a are removed separately, thereby yielding two recovered stationary series for Q_s (Fig. 8a) and T_a (Fig. 8b). The increasing rate (0.42 mm yr^{-1}) of simulated evaporation based on the original data series can be reduced to much closer to 0 (-0.17 mm yr^{-1}) based on the recovered data series with both trends removed. In addition, the increasing trend can even be enhanced (1.20 mm yr^{-1}) by performing simulations with the recovered Q_s and original T_a and be significantly reduced (-0.93 mm yr^{-1}) by performing simulations with the recovered T_a and original Q_s (Fig. 8c). Moreover, both the air temperature and solar radiation show positive correlations with the simulated evaporation. A 5% increase/decrease in Q_s and an 2°C increase/decrease in T_a will lead to a combined increase/decrease of approximately 10% of the simulated evaporation. Specifically, a ±5% variation in solar radiation has an effect on evaporation similar to that of a variation of ±2°C in air temperature, and both will separately influence the simulated evaporation by ±5%.

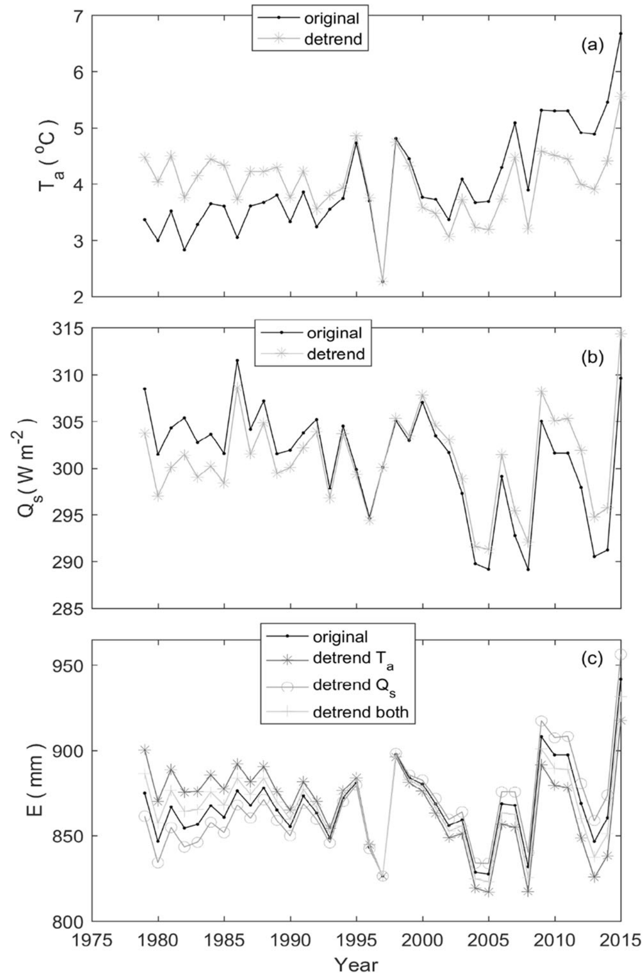

Fig. 8 Plots of the original series ("original") and the recovered series ("detrend") of **a** air temperature (T_a) and **b** downward shortwave radiation (Q_s). **c** Comparison of original total evaporation and recalculated total evaporations from April 11 to November 6. The original line is the evaporation estimated using original values of T_a and Q_s, the "detrend both" line is the evaporation estimated from detrend values of T_a and Q_s, the "detrend T_a" is the evaporation estimated from original values of Q_s and detrend values of T_a and the "detrend Q_s" is the evaporation estimated from original values of T_a and detrend values of Q_s

5 Discussion

The changes in the water level of Nam Co lake differ markedly from the changes in precipitation due to the large amount of water supplied through surface inflows by melting glaciers. Nevertheless, a significant influence of precipitation on the water levels of Nam Co lake could be observed in the period with low precipitation (July 27 to August 26 in 2013 in Fig. 3a), when a slight drop in the water level (20 mm) occurred in Nam Co lake whilst a significant drop (80 mm) in the water level was observed in the small Nam Co lake. Furthermore, the average evaporation (812 mm) (Wang et al. 2017) during the open-water period of the small lake in 2012 and 2013 was much larger than the average precipitation (approximately 462 mm), whilst the water-level changes were only approximately -190 and 20 mm during the 2 years. Strict data quality checking and cross-validation of the observations indicate that there exists groundwater supply from the surrounding water tables. Moreover, when considering the water budget over a complete year, the influence of groundwater supply is minor, and the summer water imbalance could be partly compensated for by winter snowfall. Evidence of a distinct lake-level rise (approximately 100 mm) during the winter of 2012 (Fig. 3a) was observed, although part of this rise was caused by volume increase through water-ice phase change. Therefore, the lake-level rise in winter suggests a higher closure ratio of the water budget over a complete year. In addition, the phenomenon in which the water level rose in Nam Co lake but fell in the small lake in June indicates a strong contribution of surface water inflow to Nam Co lake from glacial melt (Zhu et al. 2010).

The actual evaporation and potential evaporation, which have a complementary relationship, become identical when evaporation is not limited by ready access to water but rather by the available energy (Szilagyi 2008). Further, an oasis effect of possible sensible heat advection and mixing of dry air from its surroundings may affect small lakes. However, the evaporation amounts we obtained using the PE, BT and BS methods all nearly match, which indicates a radiation-limited and humid environment caused by the presence of Nam Co lake. When these three methods are applied to a lake in a more arid environment, their differences should be much larger. The poor performance of the de method may result from strong lake-land breeze, which will alter the drying power of the air. The water surface temperature and site-specific bulk transfer coefficients are key parameters in Dalton-type methods, whereas the thermal temperature distribution in the water plays a significant role in the energy-budget-based methods. The relatively good performance of the Mak method under condition S3 indicates that solar radiation and air temperature can provide good estimates of evaporation from small lakes.

Only the trends in wind speed and specific humidity were used in our analysis of evaporation change. The strange behaviours of wind speed in 1990s and specific humidity in the 2000s shown in Fig. 7 probably originate in the relevant data sources and do not influence our conclusions. Because small lakes have a fast thermal response to solar heating, the DI trend of Q_s corresponds well to that of evaporation in the small Nam Co lake. However, the overall trend of evaporation during the 1979–2015 period contrasts with that of Q_s, which may be attributed to the temperature-related parameter. When air warms, its ability to hold water vapour increases, and evaporation thereby increases. The short duration of the ice-cover period due to increased air temperature may increase the amount of absorbed solar radiation stored in the water, thus enhancing the increasing trend of annual evaporation and causing additional positive climate feedback. However, the increased downward longwave radiation indicates an increase in cloud covers, thereby causing a decrease in shortwave radiation, which will cause negative feedback. Pan evaporation and reference evapotranspiration over the TP decreased significantly during the few decades prior to 2000 (Hobbins et al. 2004; Xing et al. 2016). In contrast, an increasing rate of lake evaporation in the recent 10 years is indicated by our findings.

6 Conclusions

Based on unique eddy covariance measurements from a high-elevation small lake during the open water periods of 2012 and 2013, ten methods widely used for evaporation estimation were evaluated. The difference between the PT method and DK method could be attributed to the elevation-dependent variable γ. Generally, when heat storage in the water can be precisely obtained, the energy-budget-based methods perform better than the radiation-based methods and the Dalton-type methods. When the parameters used in all the methods can be optimised, the Dalton-type methods approach the performance of the energy-budget-based methods, with both types of methods performing better than the radiation-based methods. However, the radiation-based methods perform well under conditions of inadequate measurements of water temperature and heat storage in the water; for example, the Mak method yield quite similar results before and after optimization. The results reveal the significance of solar radiation and the temperature-related parameter (or *Bo*) in lake evaporation prediction. Evaporation obtained from the Mak method shows a decreasing-increasing variation with a turning point in 2004. This decreasing-increasing variation mainly results from the decreasing-increasing variation in solar radiation and is further controlled by the variations in air temperature. However, considering the overall trends from 1979 to 2015, trend removal and sensitivity analysis suggest that air warming plays the dominant role in the increasing rate of lake evaporation. Heat storage in the water body and water surface temperatures are key parameters in estimation of lake evaporation. Otherwise, a combination of solar radiation and air-temperature-related parameters could provide a good alternative.

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