Modeling Lake Titicaca Daily and Monthly Evaporation

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Abstract

Lake Titicaca is a crucial water resource in the central part of the Andean Mountain range, which is one of the lakes most affected by climate warming. Since surface evaporation explains most of the lake´s water losses, reliable estimates are paramount for the prediction of global warming impacts on Lake Titicaca and for the region´s water resources planning and adaptation to climate change. Evaporation estimates were done in the past at monthly time steps and using the four methods, as follows: water balance, heat balance, and mass transfer and Penman´s equation. The obtained annual evaporation values showed significant dispersion. This study used new, daily frequency hydro-meteorological measurements. Evaporation losses were calculated following the mentioned methods using both, daily records and their monthly averages, to assess the impact of higher temporal resolution data in the evaporation estimates. Changes in the lake heat storage needed for the heat balance method were estimated based on the morning water surface temperature, because convection during nights results in a well-mixed top layer every morning over a constant temperature depth. We found that the most reliable method for determining the annual lake evaporation was the heat balance approach, although the Penman equation allows an easier implementation based on generally available meteorological parameters. The mean annual lake evaporation was found to be 1700 mm year^{-1}. This value is considered an upper limit of the annual evaporation since the main study period was abnormally warm. The obtained upper limit lowers by 200 mm year^{-1} the highest evaporation estimation obtained previously, thus reducing the uncertainty in the actual value. Regarding the evaporation estimates using daily and monthly averages, these resulted in minor differences for all methodologies.

Key words: Lake Titicaca, heat balance, water balance, lake evaporation.

1 Introduction

Lake Titicaca, the largest freshwater lake in South America, is located in the endorheic Andean mountain range plateau – Altiplano, straddling the border between Perú and Bolivia (Fig. 1). The lake plays an essential role in shaping the semiarid Altiplano climate, feeds the downstream Desaguadero River and Lake Poopó (Pillco & Bengtsson, 2006; Abarca-del-Río et al., 2012) and supplies the inhabitants with water resources for domestic, agricultural, and industrial use (Revollo, 2001). Anthropogenic pressure on the Altiplano water resources has increased during the last decades due to population growth and increased evapotranspiration losses (FAO, 2011; Canedo et al., 2016; Satgé et al., 2017), as well as due to industrial pollution (UNEP, 1996; CMLT, 2014). The challenge of managing water resources in the Altiplano Basin is further exacerbated by climate conditions: annual rainfall is highly variable (Garreaud et al., 2003), while warming in this region exceeds the average global trend (López-Moreno et al., 2015) which is expected to intensify the evaporation from the lake surface and the evapotranspiration losses from the whole basin. The combine impact of these pressures becomes evident at the downstream end of the system, where Poopó Lake is situated. In recent years this lake suffered extreme water shortages, including its complete drying out in December 2015 (Satgé et al., 2017).

Lake Titicaca has a large surface area of about 8500 km^2 on average. Over a certain water surface level, the lake spills at the South East end and feeds the Desaguadero River. However, the major water output from Lake Titicaca is due to evaporation, which accounts for approximately 90% of the losses (Roche et al., 1992; Pouyaud, 1993; Talbi et al., 1999; Delclaux et al., 2007). In recent years, Lake Titicaca's level dropped below the outlet threshold for some periods. Thus, a small increase in evaporation or decrease in precipitation may turn the lake into a closed system with no outflow.

Since evaporation dominates the water balance in Lake Titicaca, it is essential to improve the knowledge of the lake evaporation. This is especially important in light of anthropogenic pressure and due to evident strong global warming that experiencing this region. Previous studies of Lake Titicaca evaporation have all been based on monthly meteorological observations. Due to the importance of lake evaporation, detailed calculations using daily as well as monthly observations may be necessary. Consequently, this paper investigates different methods for calculating evaporation using both daily and monthly data, in addition we discussed the possibility for the appropriate evaporation models at both time scales to be used on the study the climatic functioning and sensitivity. The main problem with the Lake Titicaca evaporation estimation is the lack of high resolution temporal data. Taking into account only the mass transfer models for different time scale, Singh et al. (1997) calculated monthly evaporation. However, doing the same calculations on a daily basis could give radically different results. For both time scales, the evaporation estimation could be more sensitive to vapor pressure. On the other hand, random errors in input data could have a significant effect on evaporation estimation at a monthly scale rather than at daily scale (Singh et al., 1997).

The Lake Titicaca surface water is cold with a temperature that remains 12-17 ^{o}C throughout the year, and below 40 m depth the temperature is almost constant (Richerson et al., 1977). The water is usually warmer than the air during day time, which means that the air immediately above the lake is unstable. The air temperature shows large diurnal variations, often exceeding

15 °C in summer. At an average terrain elevation above 4000 m.a.s.l. the solar radiation is strong and the atmospheric pressure is low, which means that the ratio between sensible and latent heat flux (Bowen ratio) is lower than at sea level. To determine the evaporation rate using the aerodynamic mass transfer approach, the atmospheric vapor pressure and surface temperature must be known. Furthermore, a wind function must be used because the atmosphere over Lake Titicaca is unstable most of the time. This means that the wind function may be different from the function used for most other lakes.

It can generally be assumed that during a year the lake water temperature returns to the value at the beginning of the year. Thus, for the heat balance, it is sufficient to know the annual net radiation, provided that the sensible heat flux can be estimated from the constant Bowen ratio. When using the method for shorter time periods, the time variation of the lake water temperature profile must be known. The heat balance approach and the aerodynamic method can be combined. The Penman method is such a combined approach. A wind function must be included also in this approach.

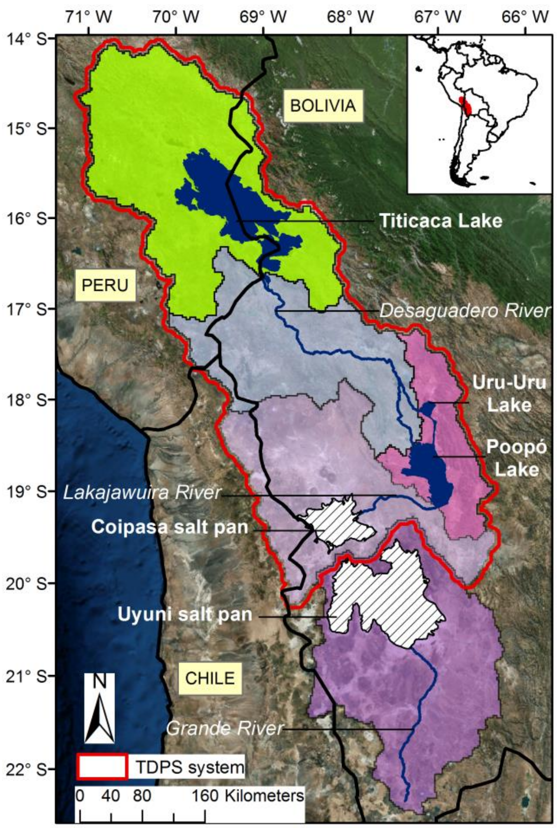

Figure 1: Lake Titicaca and the TDPS system within the Altiplano

1.1 Previous investigations

One of the first evaporation studies for Lake Titicaca applied the water balance method using measurements for the period 1956-1973 (Carmouze et al., 1977) and estimated a mean annual lake evaporation of 1550 mm y^{-1}. Taylor and Aquize (1984) applied a bulk transfer approach for a shorter period and determined the lake evaporation to be 1350 mm $year^{-1}$. The largest reported annual evaporation is from using the energy balance approach. Richerson et al. (1977) found the lake evaporation equal to 1900 mm $year^{-1}$. Later, Carmouze (1992) used the same approach and found the lake evaporation to be 1720 mm $year^{-1}$. Using observations for the period 1965-1983 and the water balance method, Pouyaud et al. (1993) found the mean annual evaporation equal to about 1600 mm $year^{-1}$. Thus, the mean annual evaporation has been estimated in the range 1350-1900 mm $year^{-1}$. While the precipitation can vary much from year to year, the large range of calculated annual evaporation, 1350 to 1900 mm $year^{-1}$, is likely to be the result of uncertainties in the evaporation estimations and in the temporal resolution of the measurements. Recently, Delclaux et al. (2007) studied the evaporation from Lake Titicaca using in-situ pan evaporation measurements, energy balance, mass transfer, and the Penman methods. They concluded that the mean annual evaporation may be about 1650 mm $year^{-1}$, with low seasonal variation between 135 mm in July (winter) and 165 mm in November (summer).

This study applies the above mentioned methodologies using the frequent and accurate hydro-meteorological measurements acquired at Lake Titicaca in 2015 and 2016, with the aim to reduce the uncertainty in evaporation estimates and evaluate the effect of using daily records instead of monthly ones.

2 Study area

Lake Titicaca is a unique biosphere due to its large depth and volume, high elevation and tropical latitude. It is located in the northern part of the Perú-Bolivian Altiplano, between latitude of 15º 25’ and 16º 35’ south. It is surrounded by the eastern and western Andean mountains. The total Lake Titicaca basin area including the lake itself is close to 57000 km^2 with a mean elevation higher than 4000 m.a.s.l. The outlet sill is at 3807 m.a.s.l. The lake volume is about 903 km^3 with a corresponding mean depth of 105 m (Boulange and Aquize, 1981; Wirrmann, 1992). The only outlet is the Desaguadero River that ends in the shallow Poopó Lake. The modern Lake Titicaca consists of the major Titicaca (Lake Chuquito), which is 284 m at the deepest point, and the smaller Titicaca (Lake Huiñamarca). The latter lake represents 1200 km^2 with a maximum depth of 35 m below spill-level. The threshold between the two lake basins is at 19 m below spill-level (see Fig. 2). The lake is described by Dejoux and Iltis (1992) and in the Encyclopedia of Lakes and Reservoirs edited by Bengtsson et al. (2012).

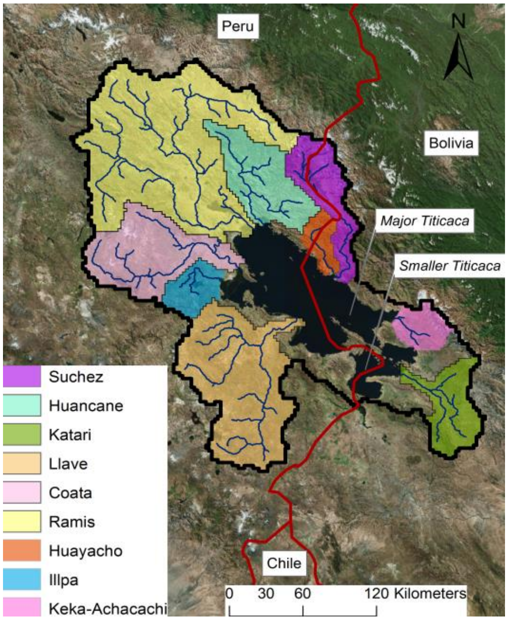

Figure 2: Lake Titicaca Basin with sub-basins and major and smaller lakes.

2.1 Hydrology

The Lake Titicaca watershed is a part of the TDPS system (Titicaca, Desaguadero, Poopó, and Salares) within the Altiplano (Revollo, 2001). The Poopó Lake is considered a terminal lake with only one discharge event into the downstream Coipasa salt pan occurring in the last century (Pillco & Bengtsson, 2006). The Lake Titicaca Basin itself includes the sub-basins Katari, Coata, Huancane, Huaycho, Ilave, Illpa, Keka-Achacachi, Ramis, and Suchez. The largest is the Ramis River Basin with an area of 15000 km^2, representing 30% of the total basin (Fig. 2). The mean flow of the Ramis River for the period 1965-2011 was 72 m^3 s^{-1}. The mean outflow from Lake Titicaca through the Desaguadero River, for the same period was 35 m^3 s^{-1}. During these 50 years, Lake Titicaca experienced large changes in water level, with a mean close to 3808.1 m.a.s.l., which is about 1 m above the outlet threshold (Pillco & Bengtsson, 2006). From low to high water level, the Lake Titicaca water surface area might change from a minimum of 7000 to a maximum of 9000 km^2.

2.2 Climate

The northern part of the Altiplano is semiarid, while the southern part including the biggest salt pans in the world is arid (TDPS, 1993). The climate is further characterized by a short wet season (Dec-Mar) and a long dry season (Apr-Nov) (Garreaud et al., 2003). The average precipitation over the Lake Titicaca Basin is about 800 mm year^{-1}, out of which more than 70% fall during the wet season (Garreaud et al., 2003). Over the lake, annual precipitation is assumed to vary from 1200 mm year^{-1} in the central part to 800 mm year^{-1} along the shores (TDPS, 1993). January is the wettest (about 180 mm month^{-1}) and

July the driest (less than 10 mm) month. In the central and southern parts of the Altiplano, the total annual precipitation is about 350 mm (Roche et al. 1992; Pillco & Bengtsson, 2006) and less than 200 mm over the Salares in the southernmost areas (Satgé et al., 2016). The seasonal variability of precipitation in the basin is related to changes in the upper troposphere circulation. During the Austral summer, an upper-level cyclone is established southeast of the central Andes. The Bolivian High brings easterly winds and allows influx of moisture from the central continent over the plateau during periods intensifying the precipitation (Garreaud, 1999; Vuille et al., 2000).

The daily air temperature over Lake Titicaca is rather constant throughout the year, usually varying between 7 and 12 °C but sometimes up to 20 °C in summer. The Titicaca region is more humid than the more southern parts of the Altiplano. The relative humidity varies between 52 to 68% as monthly average with diurnal variation between 33 and 80%. According to Carmouze (1992), the dominant wind on the lake is north-west to south-east direction, with mean monthly wind velocity close to 2 m s^{-1}, rarely reaching 5 m s^{-1} at daily time step. The general climate and hydrology are summarized in Table 1. The total river inflow was estimated through a representative area approach based on the Ramis River discharge.

Table 1. Climatological and hydrological components of Titicaca Lake for 1966-2011

Months	Ramis flow (m^3/s)	Inflow (m^3/s)	Outflow (m^3/s)	Lake depth (m)	Precipitation (mm/month)	Air temperature (ºC)	Relative humidity (%)	Wind velocity (m/s)
Jan	173.5	549.0	31.3	283.9	177.0	10.2	67.5	1.9
Feb	198.8	629.4	39.4	284.1	141.6	10.1	67.4	1.8
Mar	190.4	602.8	49.8	284.4	126.8	10.0	67.3	1.8
Apr	104.9	332.1	50.9	284.5	50.6	9.8	61.9	1.7
May	38.7	122.6	46.5	284.5	13.2	8.8	54.9	1.7
Jun	20.7	65.5	41.8	284.4	7.3	7.8	52.9	1.9
July	14.5	46.0	37.0	284.3	6.6	7.7	52.8	1.9
Aug	11.0	34.9	32.1	284.2	13.2	8.3	53.7	2.0
Sep	9.9	31.3	28.1	284.1	29.9	9.2	55.1	2.1
Oct	14.0	44.2	24.1	284.0	45.5	10.2	55.4	2.1
Nov	24.0	76.0	21.7	283.9	56.7	10.8	56.4	2.1
Dec	55.2	174.7	22.1	283.9	102.5	10.6	62.4	2.0

3 Methods

3.1 Lake evaporation models

Four evaporation estimation conventional methods were applied in this study, water balance, energy balance, mass transfer, and the Penman method. These approaches have previously been used by other researchers to estimate Lake Titicaca evaporation at a monthly time step (Carmouze, 1992; Pouyaud, 1993; Delclaux et al., 2007). The methods are briefly described as follows.

Energy balance approach

The energy balance approach (Maidment, 1993) which comes from energy balance equation in term of evaporation and multiply by the coefficient of latent heat to convert it into units of energy is:

$$\lambda E = \frac{R_n - Q_{heat}}{1+\beta} \tag{1}$$

Where λ is the latent heat vaporization (J Kg^{-1}), E is evaporation rate (mm day^{-1}), R_n is net radiation (w m^{-2}), Q_{heat} is heat storage within the water (W m^{-2}), and β is the Bowen ratio (Bowen, 1926):

$$\beta = \gamma \frac{T_w - T_a}{e_w - e_a} \tag{2}$$

$$\gamma = \frac{c_p p_a}{0.622\lambda} \quad (3)$$

where γ is the psychrometric constant (mbar °C^{-1}), T_w and T_a are the surface water and air temperature (°C), respectively, e_w and e_a are the water surface and air vapor pressures (Pasc), respectively, c_p is the specific heat capacity (J Kg^{-1} °C^{-1}), and Pa is the atmospheric pressure (Kg Pa). The psychrometric constant and thus, also the Bowen ratio, are lower at this high altitude than at sea level. The net radiation is the sum of net short-wave and net long-wave radiation. The net short wave radiation is R_s (1- albedo), where R_s is the solar radiation reaching the lake (W m^{-2}). The atmospheric long-wave radiation as well as the back radiation are computed from the Stefan law. The emissivity of the water is well known, it was set to 0.98 and the emissivity atmosphere must be known. The emissivity of the atmosphere depends on humidity, temperature, and cloudiness. Atmospheric emissivity accounting for clouds was proposed by Crawford and Duchon (1999):

$$\varepsilon_e = (1-s) + s\varepsilon_o(T_a, e_a) \quad (4)$$

$$s = \frac{R_s}{R_{s,o}} \quad (5)$$

$$R_{s,0} = R_a e^{\left(\frac{-0.0018 P_a}{K_t \sin\phi_{24}}\right)} \quad (6)$$

$$\varepsilon_o = 1.18\left(\frac{e_a}{T_a}\right)^{\frac{1}{7}} \quad (7)$$

Where s is the proxy cloudiness defined as the ratio of measured incoming solar radiation R_s (W m^{-2}) to the solar radiation received for clear sky conditions $R_{s,o}$ (W m^{-2}), ε_0 is the emissivity in clear-sky condition, which is determined from the vapor pressure e_a expressed in hPa and T_a temperature in Kelvin. The Φ_{24} is mean daily sun angle. The constant 1.18 describes the attenuation defined for the region according to Lhome *et al*. (2007). The extraterrestrial solar radiation R_a (W m^2) is determined as function of local latitude and altitude, and time of year using the turbidity coefficient K_t=0.85.

The energy equation is fairly easy to use over a full year, since the lake water usually returns to its initial state when computations were started, or when Q_{heat} equals zero (W m^{-2}). When using the approach over shorter periods the variation of the water temperature in the lake must be accounted for. In Eq. (1), the change of heat storage is included. From temperature water profiling observations, it was assumed that the water temperature below 40 m does not change from month to month. The temperature, T_w, above this mixing depth, h_{mix} (m), changes but remains almost homothermal after convective mixing during night (Richerson et al., 1977), which also is corroborated by our own field investigations. Thus, the change of heat content can be estimated from measured surface temperature:

$$Q_{heat} = \rho c_p \frac{V_{mix}}{A_{lake}} \frac{\partial T_w}{\partial t} \quad (8)$$

Where ρ is density of water (kg m^{-3}), c_p is the specific heat capacity of water, and V_{mix} is the volume above the mixing depth (km^{-3}). Carmouze et al. (1992) introduced the concept of exchange of heat between surface and deep water using the energy balance concept. The results of Carmouze *et al*. (1992) were compared to the calculation results in the present study.

Mass transfer approach

The mass-transfer aerodynamic approach is used in various models based on Dalton's law (Dalton, 1802). The latent heat transfer is related to the vapor pressure deficit. Most often a linear wind function is used (e.g., Carmouze et al., 1992)

$$E = (a + bU)(e_w - e_a) \quad (9)$$

Where E is evaporation rate, U is wind velocity (m s^{-1}), (e_w - e_a) is vapor pressure deficit (mbar). The parameter a accounts for unstable atmospheric conditions. Carmouze *et al*. (1992) used a=0.17 (mm mbar^{-1} day^{-1}) and b=0.30 (mm mbar^{-1} s m^{-1}).

Penman approach

The Penman equation is a combination of energy balance and mass transfer for evaluating open water evaporation (Penman, 1963):

$$E = \frac{\Delta}{\Delta + \gamma}\frac{(R_n - Q_{heat})}{\lambda\rho} + \frac{\gamma}{\Delta + \gamma}\mathrm{c}(a' + b'U)(e_s - e_a) \quad (10)$$

Where E is open water evaporation. The slope of the water pressure-temperature curve is denoted by Δ (K Pa s^{-1}), e_s-e_a is the saturation deficit of the air (K Pa), here e_a is dependent on the relative humidity (%). Delclaux et al. (2007) applied the Penman equation to Lake Titicaca using a' = 0.26 and b' = 0.14 and c=1 after optimizing (mm day^{-1} mbar).

Lake water balance model

The water balance approach was applied to calculate water levels in Lake Titicaca in a previous study by Pillco & Bengtsson (2007). The water balance is:

$$A_{lake}\frac{\partial h}{\partial t} = (P - E)A_{lake} + Q_{in} - Q_{out} \quad (11)$$

Where $\partial h/\partial t$ represents change in water depth, P is precipitation on the lake (mm), E is evaporation from open water (mm day^{-1}), A_{lake} is water surface of the lake (km^{-2}), which is a function of depth, Q_{in} is inflow to the lake, and Q_{out} represents the outflow from the lake (m^3 s^{-1}). Computations were carried out at a monthly time scale for two periods, one for 1966-2011 also for 2015-2016. As already pointed out, the most reliable method of computing evaporation over long periods is probably the water balance method. However, the computation only can be general, since the inflow to Lake Titicaca is not measured in all rivers.

Possible errors when using monthly averages

The evaporation during individual days is not important for the water balance but only over longer periods as months. However, since the equations for calculating evaporation are not linear, the monthly evaporation computed from monthly mean meteorological data may differ from what is found when data with higher time resolution are used. In the aerodynamic approach the wind speed is multiplied by the vapor deficit. The energy balance approach includes the Bowen ratio, which may differ from day to day and can even be negative for certain periods. If high atmospheric vapor pressure is related to strong winds, the aerodynamic equation using monthly means can yield lower evaporation estimates than when daily values are used. This is further discussed below. The Bowen ratio changes during a month. When the net radiation is large, the air temperature is likely to be rather high but not necessarily related to high vapor pressure. For this situation, the Bowen ratio is relatively high and the computed evaporation higher than it would have been using a constant monthly Bowen ratio. This means that using monthly averages, the computed evaporation will tend to be low.

4 Instrumentation and data

For this study, hydro-meteorological parameters and water surface temperature were measured near Lake Titicaca's bank for 24 consecutive months (2015-2016). Vertical lake temperature profiles were also acquired periodically. Observations were taken at 15 min intervals. These records were averaged to daily and monthly values. A Campbell Scientific research-grade automatic weather station (AWS) was installed at the Isla de Luna (latitude 16°01'59'', longitude 69°04'01''), near the Titicaca Lake shore (Fig. 3). The AWS was equipped with a rain gauge sensor, a CS215 probe for measuring relative humidity and air temperature, an A100R vector anemometer and W200P wind vane to measure wind speed, and a pyranometer Skye SP1100 for solar radiation measurement. The surface water temperature was taken from Juli-Puno (latitude 16°12'58'', longitude 69° 27'31'') at a distance of 42 km from Isla de Luna. A hand-held thermometer was used to measure water surface temperature at 8 hour intervals at approximately 60 m from the shoreline. Increased day surface recorded temperature is representative of heat storage changes.

Hydrological data, such as inflow to the lake were observed at the outlet of Ramis River. The outflow through Desaguadero River was observed at Aguallamaya. This is 40 km downstream of the lake outlet. However, there are only few tributaries between the lake and this point that may contribute to the data uncertainty. The water level was observed at Huatajata at daily time step, shown as depth in Table 1. Additional lake water temperature sounding were carried out close to Isla de Luna (latitude 16°30'00", longitude 69°15'10") for specific days during summer, spring, and winter of the study period, using multiparameter data sond – Hydrolab DS5. The sounding reached a maximum lake depth of 95 m, with water temperature recording each 5 cm at the surface and each 0.5 m below of 1 m depth.

Long-term monthly temperature and wind observations from 1960 onwards were available from the Copacabana weather station mentioned above (Fig. 3; Table 1). The El-Alto station observations, 50 km from Copacabana, were used to fill in 2.5% missing wind data for the period. The monthly precipitation on the lake was determined using the rain gauge at Copacabana and Puno on the lake shore. The total inflow from all rivers was estimated from a representative area approach assuming the specific runoff to be the same for all rivers entering into Lake Titicaca. The long-term outflow from the lake was measured at the outlet of the lake and treated by Molina et al. (2015).

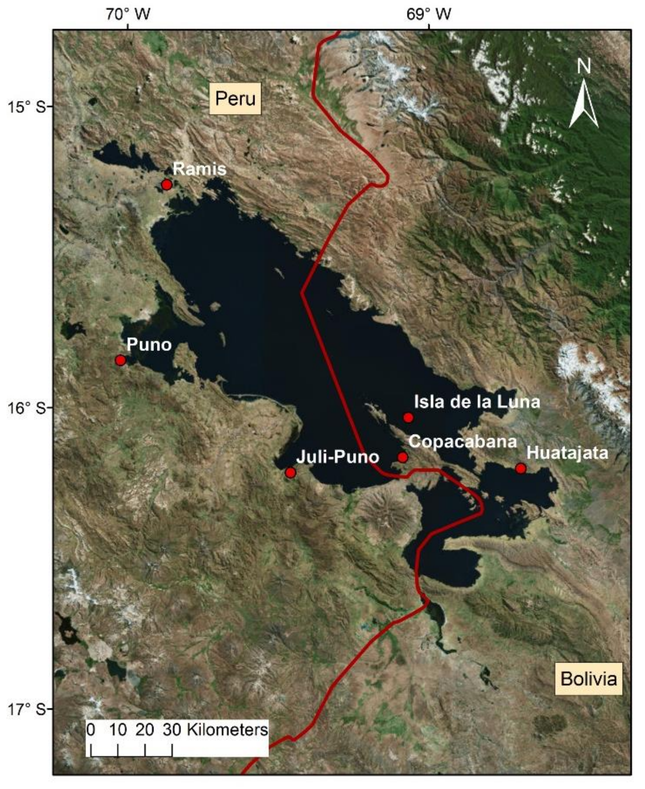

Figure 3. Location of observation points.

Table 2 and Table 3 summarize hydrological and meteorological measurements used in this study. Subscripts for vapor pressure are *w* for water, *s* for saturated air vapor pressure, and *a* is for actual air vapor pressure. The computed variables required for evaporation calculations are given in Table 4.

Table 2. Monthly mean of hydrological variables observed during the 2015-2016 period.

Month	Lake depth (m)	Inflow (m^3/s)	Outflow (m^3/s)	Precipitation (mm/month)
Jan	282,9	399,8	37,7	165,4
Feb	283,1	639,7	53,9	146,1
Mar	283,2	427,0	47,3	76,5
Apr	283,3	327,9	42,1	103,9
May	283,3	148,4	40,2	18,1

Jun	283,2	80,2	32,9	7,7
Jul	283,0	56,7	28,6	51,9
Aug	282,9	42,6	24,4	12,5
Sep	282,8	36,8	20,7	26,6
Oct	282,7	32,3	19,1	42,9
Nov	282,6	35,7	18,2	39,0
Dec	282,5	92,0	18,5	87,6

Table 3. Monthly averages of main climatic variables observed during the 2015-2016 period.

Month	Water surface temp. (°C)	Air temp. (°C)	Wind velocity (m/s)	Ralative humidity (%)	Vapour pressure (mbar)			Solar radiation, Rs (W/m^2)
					e_w	e_s	e_a	
Jan	17.2	11.1	1.60	68.3	19.8	13.9	9.5	273.3
Feb	17.3	11.4	1.61	70.3	19.9	14.2	10.0	292.4
Mar	17.5	12.0	1.50	66.1	20.2	14.7	9.7	273.4
Apr	16.5	10.8	1.52	66.7	19.0	13.6	9.1	229.2
May	15.4	10.7	1.33	53.4	17.5	13.5	7.2	234.9
Jun	14.3	10.1	1.30	51.8	16.7	13.0	6.7	236.7
Jul	13.7	9.7	1.41	50.6	16.1	12.8	6.5	237.9
Aug	14.0	9.7	154	54.4	16.4	12.8	7.0	265.8
Sep	14.7	10.5	1.50	57.6	17.1	13.6	7.8	304.7
Oct	15.5	10.9	1.63	58.2	17.6	13.9	8.1	318.7
Nov	16.4	11.7	1.61	57.1	18.8	14.8	8.4	332.1
Dec	16.9	11.6	1.70	64.9	19.5	14.5	9.4	315.9

Table 4. Monthly average parameters for evaporation calculations for the 2015-2016 period.

Month	Atmospheric Emissivity ε	Bowen ratio β	Slope of water vapour pressure Δ (mbar·100/°C)
Jan	0.81	0.38	87.40
Feb	0.80	0.38	89.10
Mar	0.79	0.31	91.90
Apr	0.80	0.35	86.20

May	0.74	0.22	85.20
Jun	0.71	0.19	82.50
Jul	0.71	0.18	80.70
Aug	0.72	0.21	80.60
Sep	0.74	0.21	84.60
Oct	0.75	0.23	86.60
Nov	0.76	0.21	90.70
Dec	0.78	0.30	90.00

Short-wave radiation was measured while long-wave radiation was computed as described above. The average for all components is shown in Fig. 4. The radiation budget is positive every day with a mean of about 150 Wm^{-2} varying from 100 in winter to 200 Wm^{-2} in summer.

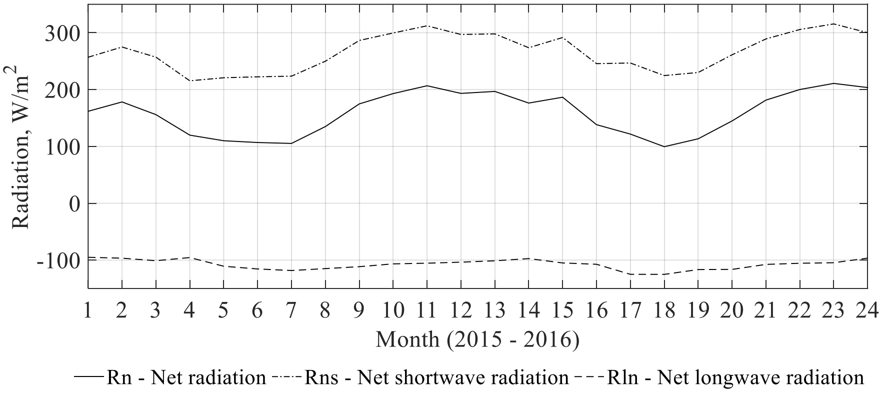

Figure 4. Monthly average radiative budget for Lake Titicaca 2015-2016.

5 Results and discussion

5.1 Monthly data

Detailed energy balance computations over the period 2015-2016 should give good estimates of the total lake evaporation for that period. After 24 months the lake surface temperature at Puno more or less returned to the temperature at the beginning of 2015. Applying this method over the two years of study the mean annual lake evaporation is 1700 mm $year^{-1}$. When computing the evaporation month by month the change of heat storage was considered in the way previously described. The mixing depth was set to 40 m. The change of the heat storage is shown in Fig. 5. The values suggested by Carmouze et al. (1992) are shown for comparison. The calculated monthly heat storage agrees well with the Carmouze estimates.

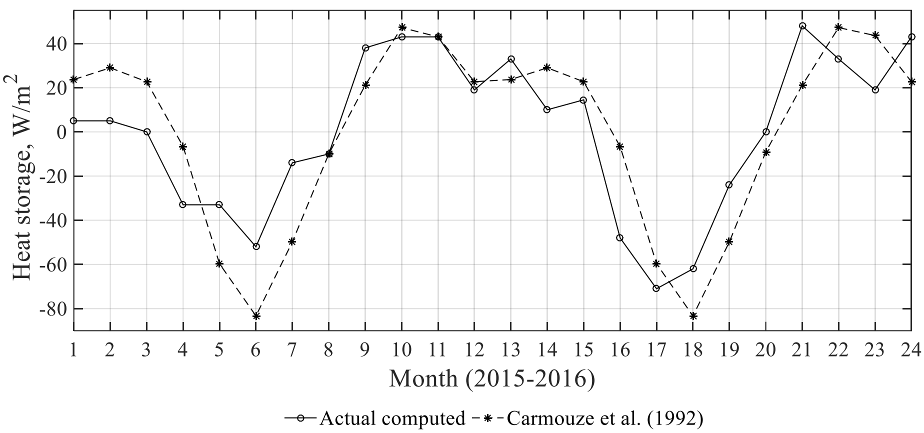

Figure 5. Change of heat storage during 2015-2016.

The computed monthly evaporation using monthly average data and the energy balance method was somewhat higher in 2016 than in 2015, 1725 mm year^{-1} as compared to 1680 mm year^{-1}. The small gap of evaporation between two consecutive years is mainly explained by the warmer season autumn of 2015 occurred; otherwise the evaporation was fairly evenly distributed over the year, being about 140 mm month^{-1}, with somewhat lower evaporation rates in July through September (see in Fig. 6).

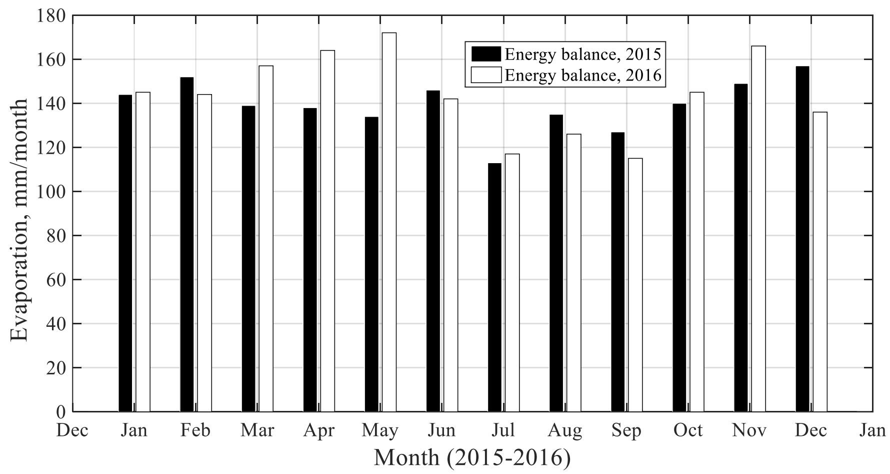

Figure 6. Monthly evaporation computed using energy balance approach.

From the energy balance and the water balance methods, the annual evaporation from Lake Titicaca was estimated in the range of about 1700 mm year^{-1}. The monthly variation depends on the change of heat storage and therefore the calculated evaporation may be high one month and low the following month. When using the mass transfer approach, similar annual evaporation as from the energy balance approach may be anticipated when applying the approaches over two full years. This may be the case even though there may be differences when comparing monthly calculations. However, when the coefficients suggested by

Carmouze et al. (1992) were used the evaporation was much higher than what was found from the energy balance method of 1720 mm year^{-1}. A good fit for the total evaporation was found using the coefficients a=0.17 mbar and b=0.155 mm mbar^{-1} s m^{-1}.

The mass transfer computed monthly evaporation over the two years is compared with the energy balance calculations in Fig. 7 for 2015 and in Fig. 8 for 2016. The computed annual evaporation by the last method was 1700 mm year^{-1} in 2015 and 1675 mm year^{-1} in 2016. Consequently, for individual years the two methods gave similar results and similar seasonal trends. This is expected since the coefficients in the mass transfer equation were chosen to fit over the two year period. For individual months, there are deviations. However, it is not possible to note any systematic differences related to different seasons of the year. The largest difference between the two methods for an individual month was about 30 mm.

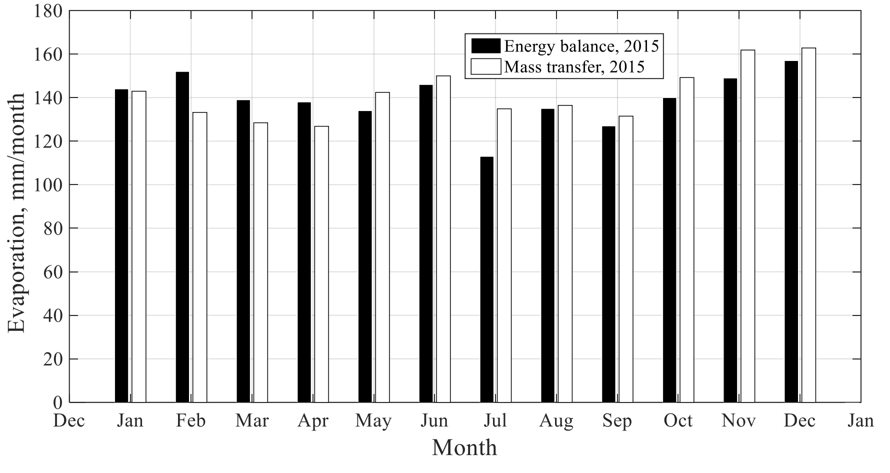

Figure 7. Comparison of monthly evaporation computed by energy balance and mass transfer method for 2015.

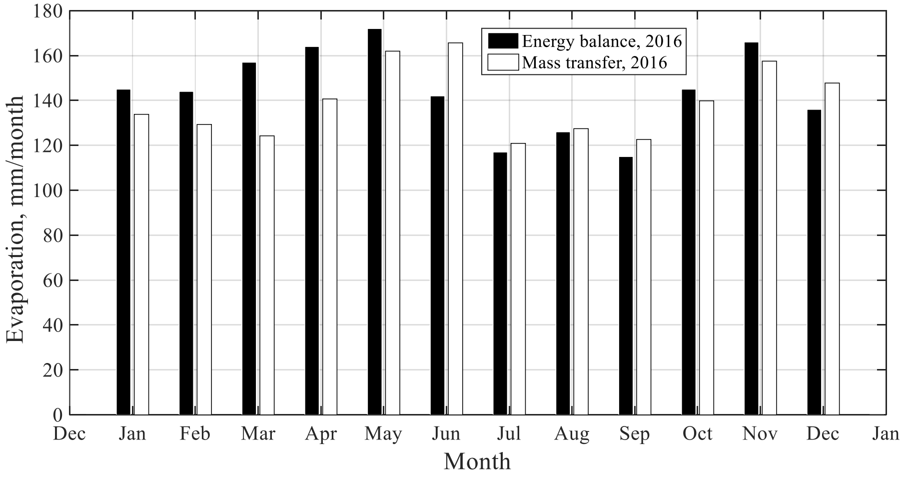

Figure 8. Comparison of monthly evaporation computed by energy balance and mass transfer method for 2016.

A summary and comparison of all investigated methods for the study period are shown in Fig. 9 and Table 5. As seen from these, the evaporation calculated from water balances differs from the three other methods. The water balance method yielded as average for tow year 1672 mm year^{-1}. As well, the same method gave the largest standard deviation. The variation in mean annual evaporation was 1633-1711 mm year^{-1}. Since the lake water level at best can be observed with cm-resolution, individual monthly evaporation becomes highly uncertain, like in February-2016 and May-2016. Thus, calculated annual evaporation is better performed using the three other methods. In general, the mass transfer, energy balance, and the Penman method gave similar monthly variation as described above, only the evaporation by Penman gave less rate from the rest of models to be 1621 mm year^{-1}; while energy balance gives the highest evaporation rate to be 1701 mm year^{-1}.

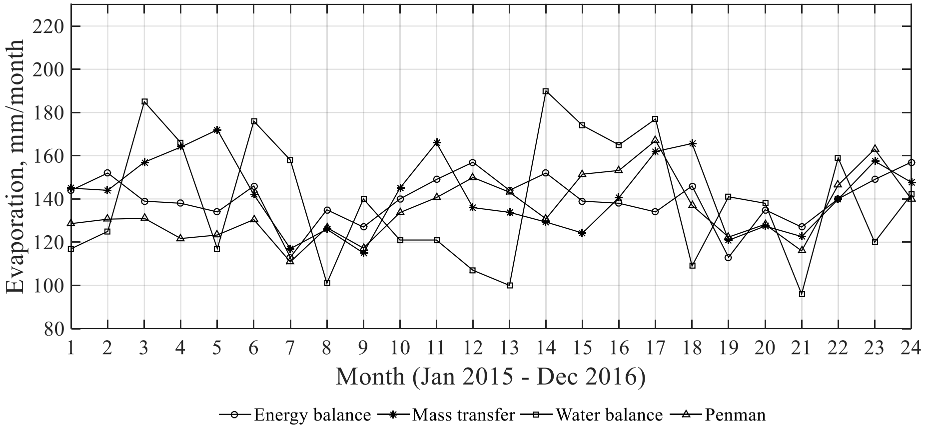

Figure 9. Monthly actual evaporation calculated by the four methods for the period January 2015 to December 2016.

Table 5. Descriptive statistics of monthly evaporation calculated by the four methods for the period January 2015 to December 2016.

Method	Mean annual (mm/year)	Mean monthly (mm/month)	Min (mm/month)	Max (mm/month)	Stand. Dev. (mm/month)	Stand. Error of mean (mm/month)
Energy balance	1701	141.8	113.0	172.0	15.4	3.1
Mass transfer	1686	140.5	120.9	165.6	13.8	2.8
Water balance	1672	139.4	96.3	189.0	29.3	4.2
Penman	1621	135.2	110.9	167.0	14.5	2.9

Water balances were computed as well for the long-term period and continuous available data 1966-2011. During the computation the A-lake parameter in the model was assumed constant equal to 8800 km^2. The resulting annual lake evaporation for the period 1966-2011 was about 1600 mm year^{-1}. The mean water balance components for the last period at month mean values are shown in Table 6. When we used the water balance approach the computed evaporation internally varied much from month to month and year to year. This is an indication that some hydrological input data are uncertain, like the precipitation which can be improved from one point observation at lake shore by the satellite observation. In any case, water balance computations over a long time period should give reasonable estimate of mean lake evaporation.

Table 6. Monthly mean water balances for Lake Titicaca for the period 1966-2011.

Month	Ramis flow (m^3/s)	Inflow (m^3/s)	Outflow (m^3/s)	Lake water depth (m)	Precipitation (mm/month)	Evaporation (mm/month)
Jan	173.5	549.3	31.3	283.9	177.0	143.7
Feb	198.8	629.4	39.4	284.1	141.6	123.2
Mar	190.4	602.8	49.8	284.4	126.8	124.6
Apr	104.9	332.1	50.9	284.5	50.6	137.2
May	38.7	122.6	46.5	284.5	13.2	139.6
Jun	20.7	65.5	41.8	284.4	7.3	136.6
Jul	14.5	46.0	37.0	284.3	6.6	133.1
Aug	11.0	34.9	32.1	284.2	13.2	121.7
Sep	9.9	31.3	28.1	284.1	29.9	118.3
Oct	14.0	44.2	24.1	284.0	45.5	134.9
Nov	24.0	76.0	21.7	283.9	56.7	134.3
Dec	55.2	174.7	22.1	283.9	102.5	127.9

5.2 Using daily data

When calculating evaporation using daily data, it was found that there are large differences between the methods and ignoring the water balance method. The maximum daily evaporation using the mass transfer method was 12 mm day^{-1}. Neither the energy balance nor the Penman method gave higher evaporation than 8 mm day^{-1}. There was a poor agreement between the mass transfer computed daily evaporation and the corresponding results using the other two methods.

When the mass transfer approach is used, it is straight-forward to determine the daily evaporation. Using the energy balance, the change of heat storage in the lake must be determined with high resolution. Detailed water temperature measurements are not available, however. Instead, it was assumed that the temperature changes at a steady rate through individual months. August as an example since the temperature for the whole month changed very little (0.2ºC). The computed daily evaporation is shown for August 2015 in Fig. 10. There were two days with average wind exceeding 6 ms^{-1}. Consequently, the evaporation was high during these days when the mass transfer approach was used.

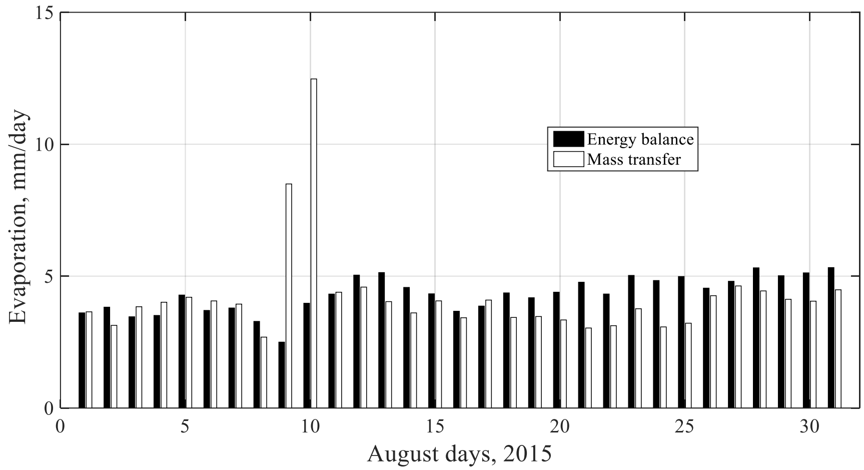

Figure 10. Daily evaporation computed by the mass-transfer (shaded staples) and energy balance (filled staples) method.

When annual evaporation was determined using daily data instead of monthly mean data, there was hardly any difference for the mass transfer method. As indicated above it is not possible to use the energy balance method with short time resolution when temperature changes from day to day have to be taken in to account. However, the evaporation can be computed neglecting the heat change keeping the Bowen ratio constant through-out a month and changing the Bowen ratio day by day. In this case, it was found that evaporation increased by about 2%. The conclusion, considering the many uncertainties involved in estimating evaporation, is that it is sufficient to use monthly means when estimating evaporation.

The evaporation computed with the Penman equation falls between what was found by the energy balance and the mass transfer approach being somewhat closer to the energy balance than to the mass transfer results. Since monthly means are sufficient for computing evaporation with the two above methods, mean values are sufficient also when using the Penman method.

It has to be noted that Lake Titicaca's near-bank surface temperatures have been observed to be warmer than the lake surface's average during daytime using satellite thermal imagery, as reported in other lakes (e.g. Marti-Cardona et al., 2008). According to this observation, the temperatures acquired for this study are likely to be an overestimation.

The spatial distribution of Lake Titicaca's surface temperature and its impact on the evaporation losses is currently under analysis. However, for the energy balance method daily changes rather than absolute temperatures were used, which are considered to be reasonable approximations of the heat storage changes. Over the larger period of air temperature observed at Copacabana weather station (1966-2016), the particular months in 2015-2016 have been characterized by the strongest El-Niño dry phenomena during the last 50 years (http://www.ciifen.org), in comparison to rest of years the air temperature were recorded higher than the average of 10 °C and close to 20 ˚C at daily time step. Then the rates of evaporation found might express in somehow the indicated warmer period.

6 Conclusions

Due to uncertainty of most observed data such as river inflow to Lake Titicaca, and mainly the discharge data, it might be no easy to improve the water balance results; then it is suggested that the most reliable method of determining the lake evaporation is using the heat balance approach. To estimate the lake evaporation using this method, heat storage changes must be known. Since convection from the surface layer is intense during nights resulting in a well-mixed top layer every morning, it is possible to determine the change of heat storage from the measured morning surface temperature. The lake evaporation is fairly uniformly distributed over the year with lows between July and September. The mean annual evaporation is about 1700 mm year^{-1}, and the mean monthly evaporation is 141.8 mm month^{-1}. When using the mass transfer approach, the required coefficients in the aerodynamic equation was set so that the mean annual evaporation agreed with that obtained from the heat balance calculations. These coefficients were found to be lower than coefficients used in previous studies. Also, when using the mass transfer approach, the evaporation was found to be lowest in July - September.

However, for the climate changes effect on Titicaca Lake evaporation assessment purposes the practical approach rather than the two empirical models might be the Penman equation, one due to available observed data for this lake, and two due to integral behavior of the equation. Also in comparison with the two models proposed in Delclaux et al. (2007) for modeling the lake evaporation, where the first model only depends on the solar radiation data, and the second one depends plus on the air temperature factor; thus both models cannot be applied broadly. The Penman model based on the adjusted wind coefficient, the mean annual evaporation is 1620 mm year^{-1} and the mean monthly is 135 mm month^{-1}. As so far, monthly evaporation computed using daily data and monthly means resulted in minor differences. The most practical model for using at daily scale might be the mass transfer approach and the Penman in comparison to energy balance approach for being high demand observed data. Particularly the Penman equation at daily temporal scale correctly might applied for the climate changes assessment at this altitude. Nonetheless, according to spatial available data from remote sensing, the evaporation equations used at daily and month scales could be applied from now for improving the spatial pattern of the lake evaporation. Since we had really extreme single warmer days during the period 2015-2016 due to El-Niño phenomenon, must be expected to have higher daily rates of evaporation; therefore the application of the models at both time scales for the study period we believe that was found the upper limits of yearly evaporation.

7 Acknowledgements

We would like to express our sincere appreciation to HASM, Research Programme – Hydrology of Altiplano from Space to Modeling, from Lab. GET-IRD and IHH-UMSA (Instituto de Hidráulica e Hidrología, UMSA, Bolivia), financed by TOSCA-CNES (Centre National d'Etudes Spatiales), especially to Franck Timouck from IRD/France, who lead the lake weather monitoring. We would like to thank Senamhi-Bolivia (Servicio Nacional de Hidrometeorología de Bolivia) for providing long-term climatic data. Thanks also to the IMARPE-Perú (Instituto para el Mar del Perú/Puno) for providing additional hydrological data and as well surface water temperature of Lake Titicaca. As well, our acknowledgment is directed to the project: Fortalecimiento de Planes Locales de Intervención y Adaptación al Cambio Climático en el Altiplano Boliviano at Agua Sustentable-Bolivia, for providing the Titicaca Lake discharge data. Finally, we thank the Programme Babel-Erasmus UE for providing economic assistance and completing this work in Sweden.

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