Mean diel variability of surface energy fluxes over Manso Reservoir

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Abstract

Mean diel cycle of latent ($E$), sensible ($H$), net longwave ($Lw_{net}$), net shortwave ($Sw$), and net surface heat flux balance ($S$) were estimated from hourly meteorological and subsurface water temperature time series acquired for ~1 month during mid-austral autumn by a buoy system in a large tropical reservoir in Brazil. $E$ and $H$ were in phase and had their maximum ($E = 163 \text{ Wm}^{-2}; \ H = 39 \text{ Wm}^{-2}$) at early morning and minimum ($E = 112 \text{ Wm}^{-2}; \ H = 6 \text{ Wm}^{-2}$) midafternoon, resulting in Bowen ratios of 0.24 and 0.06, respectively. Heat loss by evaporation therefore dominates over sensible heat used to warm surface atmosphere. Atmospheric instability was present almost all the time, increasing latent and sensible heat flux exchange coefficients by ~50% over their neutral values (from $1.4 \times 10^{-3}$ to $2.2 \times 10^{-3}$). Mean $Lw_{net}$ varied from 76 at late afternoon to 89 $\text{ Wm}^{-2}$ at early morning, indicating its importance in the overall surface heat flux balance. All 3 fluxes ($E$, $H$, and $Lw_{net}$) were positive (reservoir losing energy) throughout the day. The integrated daily average net energy budget $S$ (net short wave radiation minus $E+H+Lw_{net}$) was $\sim -60 \text{ Wm}^{-2}$; nighttime energy loss exceeded daytime gain, with consequent cooling of the reservoir. A mean temperature drop of about $-0.1 \degree \text{ C d}^{-1}$ was obtained by fitting a linear trend line to observed daily mean surface temperatures. In a qualitative way, diel time variations of surface water temperature were consistent to the time variability of $S$, indicating the dominant role of the surface heat budget in modulating surface layer temperatures of the reservoir.

Key words: ABL instability effect, diel variability, Manso reservoir, surface heat budget, turbulent and radiational surface heat fluxes

Introduction

Several multi-institutional and multi-disciplinary research projects have been conducted in Brazil in the last few years as part of an initiative to assess the contribution of the country’s large hydroelectric reservoirs to global emissions of greenhouse gases (GHG; Ometto et al. 2011, 2013). One of these monitored and studied reservoirs is Manso Reservoir, located in central-western Brazil in a typical tropical climate and savannah biome. With a surface area of about 400 km², it can be classified as a large inland waterbody that, according to the global statistics of lakes and reservoirs of Downing et al. (2006), fits the relative frequency of only 3 waterbodies of such size per 1 million km². The reservoir is formed by 2 major branches and has a dendritic shape (Fig. 1). Most of the precipitation across its drainage basin is concentrated during the rainy season that extends from September to March; a dry season is observed from April to August (Valério et al. 2009, Assireu et al. 2011; see Table 1 for physical characteristics of Manso Reservoir).

Explaining spatial and temporal changes of GHG emissions and of water quality of these waterbodies requires a better understanding of the physical and dynamic processes controlling reservoir surface and subsurface water temperature as well as stratification and horizontal and vertical water circulation (Ishikawa and Tanaka 1993, MacIntyre 1993, Engle and Melack 2000, Joyce and Jewell 2003). The surface heat flux balance is among the most important processes controlling water temperature and vertical stratification (Henderson-Sellers 1986). A net heat input results in a positive buoyancy flux, making the surface layer warmer and more stable, while a net surface heat loss cools the surface layer and promotes...
vertical mixing. In addition to heat-induced buoyancy, mechanical vertical mixing is produced by the wind. So, apart from the contribution of inflow and outflow conditions, a combination of heat and wind mixing strongly controls water column stratification, thickness, and temperature of the surface mixed layer of a reservoir (Imberger and Hamblin 1982). An accurate incorporation of surface heat fluxes is also needed for dynamic numerical lake and reservoir modelling (Lofgren and Zhu 2000).

The stratification of mid to high latitude lakes and reservoirs shows a strong seasonal cycle. They stratify in the summer when the net heat flux is toward the reservoir and is normally high. By the end of autumn and beginning of winter, the net flux reverses and high outgoing fluxes of latent and sensible heat flux force strong cooling of surface waters, producing convective mixing (Lofgren and Zhu 2000). In tropical lakes and reservoirs, the seasonal signal is still present but is much weaker. For low latitude waterbodies, the upper water layer normally stratifies during daytime and becomes near isothermal at the end of nighttime if the water column is not too deep (MacIntyre and Melack 1982, 1984, 1988, Hare and Carter 1984). Studies conducted in other tropical lakes show that stratification and mixing in a diel time scale is strongly controlled by the surface energy fluxes (MacIntyre et al. 2002).

Heat flux exchanges between reservoirs and the atmosphere and the effects of these fluxes on stratification and water circulation is not well understood for Brazilian reservoirs. To help fill this knowledge gap, we carried out an in situ data collection campaign in Manso Reservoir during part of the austral autumn 2007. The main goal of this effort was to collect the necessary data to calculate the surface heat flux components (latent, sensible, and longwave) and the net shortwave radiation and to determine the net surface heat flux balance. In this study we examine the magnitude and relative role played by each flux component in the overall net surface heat balance and their average diel variability. We also discuss the influence of the net heat flux on the observed diel variability of the surface layer temperature. Particular attention was given to the role of the atmospheric boundary layer (ABL) instability in strengthening the turbulent fluxes.

**Dataset and methods**

**Dataset**

The in situ data acquisition of lake-wide meteorological and limnological parameters used a buoy system developed in-house, the Integrated System for Environmental Monitoring (SIMA, a Portuguese acronym), which was moored by a cable near the reservoir dam at about 60 m water depth (Fig. 1). The SIMA buoy can function autonomously with acquisition, on-board storage, and satellite transmission of all collected variables (Lorenzzetti et al. 2005).

The meteorological data included (a) shortwave solar radiation (Novalynx 240-2101; spectral sensitivity 0.3–3 µm; linearity <0.5% in the range 0.5–1330 W m⁻²); (b) air temperature and relative humidity (Rotronic MP103A; accuracies of ±0.3 °C and ±1.5%, respectively) sensors mounted within a housing for solar radiation shielding; (c) atmospheric pressure (Vaisala PTB100A; accuracy of ±0.3 hPa at 20 °C); and (d) wind magnitude and direction (R.M. Young 05106; accuracies of ±0.3 ms⁻¹ and ±3°, respectively). All SIMA-acquired meteorological data were collected at a height of ~3 m above the water surface. Water temperatures were acquired at 2, 5, 20, and 40 m water depths using a thermistor chain, and at 1 m depth using a YSI 6560 sensor deployed in a YSI 6600 V2 multiparameter sonde with a factory accuracy of ±0.15 °C. In this report, water surface temperature (Tₛ) is represented by our temperature closest to the surface at 1.0 m depth.
All data were acquired once per hour, centered at the “full hour.” The wind data were sampled every 30 s, 10 times before and 10 times after the full hour; an average value was calculated from the 21 samples for direction and intensity. For all other variables, a burst sampling (a sequence of short-lived, high frequency sampling interleaved by a waiting period) during 0.3 ms at a rate of one sample per 10 µs was used at the full hour. This procedure resulted in 30 samples used to calculate an average value and to reduce electronic noise.

The available time series of all mentioned variables began 22 April 2007 at 00:00 h and ended 20 May 2007 at 23:00 h (local time), resulting in 696 hourly data points per time series. Additionally, we used air temperature data collected at a meteorological surface station installed on-land ~5 km from the reservoir margin and ~10 km from the SIMA position (Fig. 1).

**Methodology**

The turbulent fluxes of latent and sensible heat were estimated using the Bulk Aerodynamic Transfer Method, taking into account the dependence of the transfer coefficients on the ABL stability (Amorocho and DeVries 1980, Imberger and Patterson 1990, Verburg and Antenucci 2010). The SIMA-measured variables used to calculate the fluxes were air temperature ($T_a, ^\circ C$), surface water temperature ($T_w, ^\circ C$), relative humidity ($R_h, \%$), atmospheric pressure ($p, hPa$), and surface wind speed ($U_z, \text{ms}^{-1}$). All meteorological variables were acquired at about 3 m height above the water surface. Latent ($E$) and sensible ($H$) heat fluxes (Wm$^{-2}$) were calculated by:

$$E = \rho_a L_v C_E U_z (q_s - q_z)$$
$$H = \rho_a C_a C_H U_z (T_w - T_a),$$

where $\rho_a$ is the air density (kg m$^{-3}$), $L_v$ is the latent heat of vaporization (J kg$^{-1}$), $C_a$ is the specific heat of air (J kg$^{-1}$ °C$^{-1}$), $q_s$ (kg kg$^{-1}$) is the specific humidity at saturation pressure at $T_w$, and $q_z$ (kg kg$^{-1}$) is the specific humidity of air at height $z$. The exchange coefficients $C_E$ and $C_H$ are assumed equal (Zeng et al. 1998).

The correction of the turbulent fluxes for variations of the atmospheric stability is made via the Obukhov stability length $L$ (m):

$$L = \frac{-\rho_a U_z^2 T_v}{\kappa g \frac{H}{C_a} + 0.61 \frac{(T_{a} + 273.16) E}{L_v}},$$

where $\rho_a$ is the air density (kg m$^{-3}$), $L_v$ is the latent heat of vaporization (J kg$^{-1}$), $C_a$ is the specific heat of air (J kg$^{-1}$ °C$^{-1}$), $q_s$ (kg kg$^{-1}$) is the specific humidity at saturation pressure at $T_w$, and $q_z$ (kg kg$^{-1}$) is the specific humidity of air at height $z$. The exchange coefficients $C_E$ and $C_H$ are assumed equal (Zeng et al. 1998).

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**Fig. 1.** (a) Manso Reservoir regional location; (b) settings; and (c) SIMA autonomous data collection buoy system.
where \( u_* = C_D^{1/2} U_f \) is the friction velocity (m s\(^{-1}\)), \( C_D \) is the drag coefficient (Smith 1988), and \( T'_w \) is the virtual air temperature. If \( L < 0 \), the atmospheric boundary layer is unstable, increasing the transfer coefficients; if \( L > 0 \), air above water is stable, and \( H \) and \( E \) are reduced. The case of a neutral atmosphere corresponds to \( L \) tending toward \( \pm \infty \), and the stability parameter \( \zeta = \frac{z}{L} \to 0 \). Because \( E \) and \( H \) depend on \( C_w \) and \( C_{w*} \) and \( u_* \) depends on \( C_u \), which are corrected by stability using \( L \), which in turn depends on \( E \), \( H \), and \( u_* \), the fluxes were calculated using an iterative procedure suggested by Hicks (1975). The procedure begins by defining the neutral atmospheric stability transfer coefficients \( C_{sw*}, C_{en*}, \) and \( C_{hn*} \), which are used to derive initial values for \( E, H \), and \( u_* \). From these initial values, a first \( L \) is derived. A loop calculation is now implemented in which the transfer coefficients are modified using the atmospheric stability functions \( \phi \), which depend on \( \zeta \), a function of \( L \) (Brutsaert 1982). With the adjusted coefficients, new values of \( u_* \), \( E \), and \( H \) are calculated and used to update \( L \) and the \( \phi \) functions; this process is repeated until the new \( L \) converges to the previous one to within 0.001\%, as suggested by Verburg and Antenucci (2010).

To avoid division by zero in the calculation of \( L \), we set the minimum wind speed to 0.2 m s\(^{-1}\). In addition, as suggested by Imberger and Patterson (1990), we limited our algorithm to \( |z/L| \leq 15 \). A minimum of 3 and a maximum of 10 iterations were needed for convergence. Water density and \( T_w \) were calculated as functions of temperature; air density was calculated as a function of air temperature, humidity, and atmospheric pressure.

The net longwave flux (\( L_{w\text{net}} \); W m\(^{-2}\)) was estimated as the residual flux between the incoming (after correction for longwave albedo) and emitted longwave radiation. The emitted longwave flux (W m\(^{-2}\)) is given by:

\[
L_{w\text{em}} = \varepsilon_w \sigma T_w^4, \tag{4}
\]

where \( \varepsilon_w \) is the emissivity of water, assumed equal to 0.972 (Davies et al. 1971); \( \sigma \) is the Stefan-Boltzmann constant; and \( T_w \) is water temperature in K.

The net incident longwave flux \( L_{w\text{inc}} \) (W m\(^{-2}\)) was calculated as in MacIntyre et al. (2002):

\[
L_{w\text{inc}} = \varepsilon_w \sigma T_w^4 (1 + 0.17C^2) (1 - \alpha_{sw}) \tag{5}
\]

where \( \varepsilon_w = 0.642 (e_u/T_w)^{1/7} \) (Brutsaert 1982) is the air emissivity; \( e_u \) is the vapor pressure; \( \alpha_{sw} \) is the longwave albedo, here assumed equal to 0.03 (Henderson-Sellers 1986); and \( C \) is the fraction of cloud cover.

For short-term data acquisitions (one or a few days), \( C \) can be estimated by visual observations of the sky. For this investigation, which involved an autonomous data acquisition and a relatively long period (one month), this was not feasible. One possibility is to estimate a daily average value for \( C \) using the methodology presented by Reed and Stabeno (2002). In this case \( C \) is estimated based on the ratio of measured shortwave radiation to the insolation under clear skies (\( Sw_{\text{tot}} \); W m\(^{-2}\)) and noon-time solar altitude \( \alpha_s \) using the relation \( Sw_{\text{net}}/Sw_{\text{tot}} = 1 - 0.62C + 0.0019 \alpha_s \), proposed by Reed (1977). As indicated by Simpson and Paulson (1979), for mean daily insolation, results indicate that daily predictions can be made using this formulation with an average accuracy of \(<20\text{ Wm}^2\). The Reed relation can be applied from low to high latitudes; it was validated using an extensive set of observations of the eastern Pacific from the tropics to the high latitudes of Gulf of Alaska. Reed (1977) states that this relation is better suited for cloud covers of 0.3 or larger, and that for \( C \leq 0.2 \) the cloud factor can be practically neglected.

Considering our interest in analyzing the hourly values of surface flux components, we implemented the approach of Crawford and Duchon (1999), by which the hourly values of daytime cloud fraction are estimated as \( C = 1 - Sw/Sw_{\text{tot}} \). Here, \( Sw_{\text{net}} \) were estimated from top of the atmosphere (TOA) irradiance (Iqbal 1983) and then propagated to the surface using the procedure presented by Martin and McCutcheon (1999). Duarte et al. (2006) estimated daytime \( L_{w\text{inc}} \) using the CD99 approach and compared the results with measurements made using a calibrated pyrgeometer, obtaining a percent mean relative error of 4.5\%. For nighttime (from 18:00 to 06:00 h), the \( C \) values used in our work were interpolated from the near-sunset value of the previous day to the near-sunrise \( C \) of following day, an approach used by Sridhar and Elliot (2002). Daily average values of \( C \), calculated from hourly values estimated for the whole observation period using the CD99 methodology, were expected to be consistent with the variations of shortwave solar radiation acquired by SIMA throughout the day (Fig. 2).

The amount of shortwave radiation that penetrates the water after reflection at the surface is given by \( Sw (1 - \alpha_{sw}) \), where \( Sw \) is the above-water shortwave flux and \( \alpha_{sw} \) is the shortwave albedo. We estimated \( \alpha_{sw} \) using the relation \( \alpha_{sw} = A\theta_{	ext{sol}}^{B} \), proposed by Anderson (1954), where \( \theta_{	ext{sol}} \) is the solar altitude (in radians) and \( A \) and \( B \) are empirically adjusted constants, dependent on cloud cover fraction \( C \). As suggested by Henderson-Sellers (1986), we used the \( A \) and \( B \) values dependent on \( C \) as given by US Army Corps of Engineers (USACE 1982). Typical values of \( \alpha_{sw} \) were 15\% at 07:00 h, 4.3\% at noon, and 39\% at 17:00 h, with an average \( Sw \) albedo of 10\%. The asymmetry from 07:00 to 17:00 h values, which were the full hour nearest to sunrise and sunset, respectively, was due to a combination of a smaller solar altitude at 17:00 h and a much higher cloud cover in the early morning hours.

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All flux calculations were done at each full hour of the day for the whole observation period (696 per each flux component). The mean diel cycle of the fluxes was derived by collecting at each full hour of the day (00:00–23:00 h local time) the corresponding values for the whole observation period, which were then averaged to generate a mean value for each hour.

**Results**

**Meteorological forcing**

A clear diel signal was observed in the hourly values of air temperature, shortwave radiation, and relative humidity; water temperature likewise showed a diel pattern, but of much smaller magnitude than that for air (Fig. 2). Most of the time, $T_w$ was higher than $T_a$, a condition of an unstable ABL. Briefly in midafternoon, $T_a$ can be a few degrees warmer than $T_w$. During cold front passages, such as observed days 27 April and 8–9 May, $T_a$ can be lower than $T_w$ throughout the day and for a few consecutive days. Strong frontal systems, such as occurred 8–9 May, can push air temperature down 10 °C or more lower than water surface temperature, substantially enhancing the sensible heat flux. Lower values of daily shortwave were associated with the frontal passages, a modulation clearly caused by an increase of cloud cover. Net shortwave radiation maximum varied from ~800 to 300–450 Wm$^{-2}$ during typical and cold front days, respectively. Relative humidity generally varied from ~40–50 to 90%, with minimum and maximum values, respectively, observed in midafternoon and near sunrise. The hourly wind speed values were much noisier, with a tendency of weaker winds around noon. In general, higher winds were observed associated with cold front passages. Wind speeds of ~9 ms$^{-1}$ were observed during the 8–9 May event.

To calculate the diel cycle, we averaged each hour of the day using the corresponding hourly values observed for the whole period, which reduced the high frequency and the synoptic weather signal variability of the original

![Fig. 2. Hourly values of air temperature and surface water temperature at 1 m, just below surface net shortwave radiation and daily average cloud fraction $C$, relative humidity, and wind speed. Data derived from SIMA system for 22 April–21 May 2007. Vertical gray bars indicate the passage of cold fronts (27 April, 08–09 May, and 19 May).](image-url)
data (Fig. 3). The mean diel cycle of near surface air temperature over the reservoir showed a temperature range of 6.5 °C, with a minimum of ~23 °C between 05:00 and 06:00 h just before sunrise, after having descended from its maximum value of ~29.4 °C observed at 15:00 h at midafternoon (Fig. 3a). The surface water layer showed a weaker but similar behavior, lagging behind \( T_s \) by ~1–2 h, with an average temperature range of 0.5 °C (Fig. 3a). Water was warmer than air most of the day; a slight inversion was observed only briefly at ~15:00 h, when air temperature was maximum and \( T_s > T_a \) by about 0.5 °C. At the end of the nighttime period, from 05:00 to 06:00 h, water was warmer than air at a maximum of 5.8 °C. This average diel pattern of air and water temperature reinforces the view present in the original data that the ABL above reservoir surface was unstable almost all day.

Average near surface wind speed \( U_z \) was relatively weak, but a significant modulation was observed during the course of the day. From about 10:00 to 14:00 h, wind was at its minimum of ~1.9 ms\(^{-1}\); from 14:00 to 19:00 h it grew rapidly to a maximum of 2.8 ms\(^{-1}\); and from 19:00 to 21:00 h, a minor decrease was followed by an increase to 2.6 ms\(^{-1}\) at 01:00 h, staying near this value until 05:00–06:00 h, when it began to decrease to minimum values at 09:00–10:00 h (Fig. 3b). The diel cycle of wind intensity was consistent with a breeze signal forced by a differential heating of the surface air above reservoir and land (Fig. 3c). At the end of the night period (06:00 h), the surface air over the reservoir was warmer than land air by 3 °C, and by 11:00 h cooler by 1.8 °C. During the night, winds blow predominantly from south-southwest from land toward the reservoir; during daytime, a lake breeze wind from the east, which is not present during the night, is frequent, blowing along the Manso Reservoir axis. Maximum hourly mean shortwave solar radiation was about 680 Wm\(^{-2}\), with sunrise and sunset at about 00:06 and 18:00 h, respectively (Fig. 3b).

**Surface heat fluxes**

The hourly values of the surface heat loss components for the whole observation period (Fig. 4) indicate that the most important heat loss component was the latent heat flux, followed by the net longwave and sensible fluxes. During normal days (without meteorological disturbances), \( E \) varied typically from 100 to 200 Wm\(^{-2}\), \( L_{w_{\text{net}}} \) from 50 to 80 Wm\(^{-2}\), and \( H \) from ~0 to 25–50 Wm\(^{-2}\). During cold front disturbances, latent heat loss reached 350–500 Wm\(^{-2}\), \( L_{w_{\text{net}}} \sim 100 \) Wm\(^{-2}\), and \( H \sim 80–180 \) Wm\(^{-2}\), with \( H \) becoming eventually larger than \( L_{w_{\text{net}}} \). During normal days the total heat loss \((E + L_{w_{\text{net}}} + H)\) maximum was ~200–300 Wm\(^{-2}\), but during the strongest cold front it was near 800 Wm\(^{-2}\). This large heat loss was caused by a big increase in latent and sensible heat fluxes produced by strong winds and a large air–water temperature difference (Fig. 2). These hourly flux values are similar to those reported for April 1996 at Lake Victoria, east equatorial Africa (MacIntyre et al. 2002). Their strongest total \((E + H + L_{w_{\text{net}}})\), associated with stronger winds of ~10 ms\(^{-1}\) (not associated with frontal passage), was near 600 Wm\(^{-2}\), 200 Wm\(^{-2}\) weaker than that observed at Manso Reservoir. Our larger flux seems associated to a larger \( H \), produced by a stronger air–water temperature difference, and a higher \( E \) associated with a lower relative humidity at the time of maximum winds.

**Diel variability**

A summary of main characteristics of latent and sensible heat fluxes and their input variables (Table 2) indicated that the 2 most important components of the latent heat flux are wind speed and humidity deficit \((q_e - q_s)\); equation 1). The diel variability of \( q_e - q_s \) can be better analyzed considering constituent input physical parameters (Fig. 5). Specific humidity at saturation \((q_s)\) and the specific humidity \((q_e)\) are given by:

\[
q_s = 0.622 \frac{e_{\text{sat}}}{p} \quad \text{and} \quad q_e = 0.622 \frac{e_e}{p},
\]

where \( p \) (hPa) is the surface atmospheric pressure, \( e_{\text{sat}} \) (hPa) is the saturation vapor pressure at \( T_s \), and \( e_e \) (hPa) is the vapor pressure. These vapor pressures are given by:

\[
e_{\text{sat}} = 6.11 \exp \left[ \frac{17.27 T_s}{(237.3 + T_s)} \right] \quad \text{and} \quad e_e = \frac{R_h e_s}{100},
\]

where \( R_h \) is the relative humidity (%), and \( e_s \) (hPa) is the saturation vapor pressure at air temperature \( T_a \), which is given by:

\[
e_s = 6.11 \exp \left[ \frac{17.27 T_a}{(237.3 + T_a)} \right].
\]

Mean diel values of relative humidity reached a minimum of 54% at 14:00–15:00 h and a maximum of 83% at 04:00–05:00 h (Fig. 5b) in anticorrelation, as expected, to maximum and minimum air temperatures, respectively (Fig. 3a). Saturation vapor pressure at \( T_a \) \((e_s)\) (Fig. 5b) followed air temperature, with a maximum at 15:00 h, but vapor pressure \((e_e)\) (Fig. 5a), a product of \( e_e \) by \( R_h \), was minimum at this time, pushed down by the reduction of \( R_h \). Saturation vapor pressure at \( T_a \) \((e_{\text{sat}})\) followed water temperature, with a maximum at 16:00 h.
and minimum at 06:00–07:00 h. Although \( q_s \) and \( q_z \) also depend on atmospheric pressure as well as \( e_{sat} \) and \( e_a \), respectively, their variability is nearly equal to that of \( e_{sat} \) and \( e_a \) because a change of only 4 hPa (~0.4%) in surface atmospheric pressure is present during the day. The mean diel variability of \( q_s - q_z \) (Fig. 6c) shows that the maximum occurred at 15:00 h (Table 2) because the maximum of \( q_s \) is almost coincident with the minimum of \( q_z \) at this time of the day. Thus, the maximum \( q_s - q_z \) is highly dominated by the decrease in relative humidity associated with the peak air temperature.

Maximum values of \( E \) (~163 and 147 Wm\(^{-2}\)) were observed at 06:00 and 19:00 h, respectively (Fig. 6a; Table 2), when surface wind reached maximum values (Fig. 3b; Table 2). The decrease in \( E \) between 06:00 and 14:00 h was highly determined by the fall of the wind speed. Note, however, that \( E \) grew from midnight until 06:00 h, a period of a relatively constant wind. This increase seems to be associated with the growth of \( q_s - q_z \), which rises steadily until 15:00 h after reaching its minimum near midnight (Fig. 6c; Table 2). The minimum \( E \) (112 Wm\(^{-2}\)) was observed at 14:00 h (Table 2), although \( q_s - q_z \) was near its daily maximum by then. This was clearly caused by the minimum wind value at this time of the day, indicating the dominant role of the wind. This analysis shows that, while the wind dominates \( q_s - q_z \), part of the variability of \( E \) is still controlled by \( q_s - q_z \) (discussed later).

Variability in air density and latent heat of vaporization (Fig. 6b, 6d), which also affect \( E \), were in phase with latent heat flux changes, but their percent variations (Table 2) were too small to contribute significantly to the variability of \( E \).

The diel time variability of sensible heat flux (\( H \)) was in phase with latent heat flux (Fig. 6a). Note, however, that at 14:00 h, while minimum \( E \) was near 112 Wm\(^{-2}\), \( H \)

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**Fig. 3.** Mean diel variability of (a) air temperature \( T_a \) and surface water temperature \( T_w \); (b) wind speed and incoming shortwave radiation \( Sw \); and (c) lake–inland air temperature differences (in land–air temperature registered at a meteorological station located about 5 km inland).

**Fig. 4.** Hourly values of latent heat (\( E \)), sensible heat (\( H \)), net longwave (\( Lw_{net} \)), and total (\( E+H+Lw_{net} \)). Vertical gray bars indicate the passage of cold fronts (27 April, 08–09 May, and 19 May).
Table 2. Hourly mean turbulent heat fluxes: latent (E) and sensible (H) heat fluxes and input components. Units: H (Wm$^{-2}$); $\rho_{air}$ (kg m$^{-3}$); $L_r$ (Jkg$^{-1}$); $U_z$ (ms$^{-1}$); $q_z$ - $q_a$ (kg kg$^{-1}$); $T_{water} - T_{air}$ (°C).

<table>
<thead>
<tr>
<th></th>
<th>Max</th>
<th>Min</th>
<th>Δ (%)</th>
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</tr>
<tr>
<td>$10^3 (C_v = C_h)$</td>
<td>2.7 / 2.5</td>
<td>1.9</td>
<td>42.1/31.6</td>
<td>09:00/20:00</td>
<td>16:00</td>
</tr>
<tr>
<td>$U_z$</td>
<td>2.8</td>
<td>1.9</td>
<td>47.4</td>
<td>06:00 and 19:00</td>
<td>10:00 and 14:00</td>
</tr>
<tr>
<td>$10^3 (q_z - q_a)$</td>
<td>11.8</td>
<td>9.7</td>
<td>47.4</td>
<td>15:00</td>
<td>23:00</td>
</tr>
<tr>
<td>$T_{water} - T_{air}$</td>
<td>5.8</td>
<td>-0.1</td>
<td>—</td>
<td>05:00–06:00</td>
<td>15:00</td>
</tr>
<tr>
<td>$E$</td>
<td>163/147</td>
<td>112</td>
<td>45.5/31.3</td>
<td>06:00 and 19:00</td>
<td>14:00</td>
</tr>
<tr>
<td>$H$</td>
<td>39</td>
<td>7</td>
<td>—</td>
<td>06:00</td>
<td>14:00</td>
</tr>
</tbody>
</table>

was small (~7 Wm$^{-2}$), a consequence of a small air–water temperature difference (Fig. 6c; Table 2) and a weak wind (Fig. 3b) at this time of day. $H$ attained its maximum of 39 Wm$^{-2}$ at 06:00 h when $T_w - T_a$ was maximum (~6 °C) and the wind was also relatively high. In a mean diel sense, $H$ was positive throughout the day. We stress, however, that on an hourly basis (Fig. 2) during midafternoon hours and under normal conditions (no frontal passages), air temperatures briefly can exceed water surface temperatures, leading to negative sensible heat flux (reservoir gaining sensible heat). $H$ is also a function of wind speed, however, and during this time of the day winds are normally weak. The average wind speed corresponding to these negative sensible heat fluxes was 1.34 ms$^{-1}$, resulting in small sensible heat flux gains in Manso Reservoir (average value of ~2.82 Wm$^{-2}$). These negative fluxes are almost negligible (Fig. 4, dashed line). We also must consider that during cold front passages $H$ is positive (reservoir losing sensible heat) throughout the day, with values larger than during normal conditions; therefore, in an average sense, we can consider the diel values of $H$ positive throughout the day.

Latent and sensible heat fluxes are also affected by changes in the exchange coefficients ($C_v$ and $C_h$), which increase from their neutral values when the ABL is unstable. In general, cooler air over warmer water is associated with unstable atmospheric conditions, a prevailing situation in Manso Reservoir. For hot and humid conditions such as occur in the reservoir, however, the atmosphere can be unstable, even when $T_w - T_a = 0$ due to water vapor-induced buoyancy (Liu 1990). During the observation period, the instability parameter $\zeta$ was almost always negative. The calculated heat transfer coefficients (Fig. 6d) were consistently above their neutral values, as expected for $T_w - T_a > 0$ and low winds ($U_z \leq 3–4$ ms$^{-1}$; Verburg and Antenucci 2010). $C_v$ and $C_h$, which were high and relatively constant from midnight until 05:00 h, grew rapidly until ~09:00 h, an increase associated with the decrease of the wind during this period (Fig. 3b) accompanied by $T_w - T_a \approx 3.5$ °C; a maximum negative $\zeta \sim -4.8$ was reached at this time. Between 16:00 and 18:00 h, the ABL was still unstable but at its minimum negative $\zeta \sim -1$ when the air–water temperature difference is only 0.5 °C and wind is near its maximum; $C_v$ and $C_h$ were closest to their neutral conditions at this time of the day (Fig. 6d). Even when water and air temperatures were

![Fig. 5. Diel variability of (a) $e_z$, the vapor pressure at $T_z$ and $e_{sat}$, the saturation vapor pressure at $T_z$; (b) $e_s$, the saturation vapor pressure at $T_z$ and $R_h$, the relative humidity.](image-url)
close \((T_w - T_a \approx 0)\) near 15:00 h (Fig. 6c), \(\zeta \approx -2.5\), and \(C_E\) and \(C_H\) were well above \(C_{EN}\) and \(C_{HN}\), the corresponding coefficients under neutral condition, an anticipated result considering the light wind, high humidity, and temperatures present.

For our observation period, the average \(C_{EN}\) was \(1.46 \times 10^{-3}\). Typical values of \(C_{EN}\) are \(1.4–1.5 \times 10^{-3}\) (Strub and Powell 1987); in Lake Tanganyika, also a large tropical waterbody (\(\sim 6^\circ S\)), \(C_{EN}\) is \(\sim 1.5 \times 10^{-3}\) (Verburg and Antenucci 2010). Under the unstable conditions observed in Manso Reservoir, the average value of the transfer coefficients of heat averaged 2.2 \(\times 10^{-3}\), a 50\% increase above their neutral values. In Lake Tanganyika, atmospheric instability increased the mean annual \(C_E\) by 23\% above its neutral value (Verburg and Antenucci 2010).

Because water vapor pressure differences \((e_{sat} - e_a)\) depend on air, water temperatures, atmospheric pressure, and relative humidity and affect evaporation and instability \((\zeta)\), they are useful for comparing water vapor buoyancy and instability for different waterbodies. For the southern basin of Tanganyika Lake, \(\zeta\) values for April–May were similar to those observed at Manso Reservoir, and \(e_{sat} - e_a\) varied from 9 to 25 hPa with a mean of 15 hPa. For the study period at Manso Reservoir, these values were 9.85 and 31.56 hPa, respectively, with a mean of 16.5 hPa. The corresponding (min, mean, max) \(q_s - q_z\) values, after correcting for the different mean atmospheric pressures (921 and 968.5 hPa for Tanganyika and Manso Reservoir, respectively) were \(5.8 \times 10^{-3}\), \(9.6 \times 10^{-3}\), and \(16 \times 10^{-3}\) for Tanganyika and \(6.3 \times 10^{-3}\), \(10.6 \times 10^{-3}\), and \(20.3 \times 10^{-3}\) hPa for Manso Reservoir, respectively. So, in general, the water vapor influence on \(E\) and on ABL instability in southern Tanganyika and Manso Reservoir are similar, with the values at Manso about 9\% and 27\% higher for minimum and maximum humidity deficits, respectively.

By least square linear regression fitting of \(E\) against wind, \(q_s - q_a\) and their product, and \(H\) against wind, \(T_w - T_a\) and their product, it is possible to get a more quantitative account of the dependences of latent and sensible fluxes to those variables. In Manso Reservoir, wind alone explained 85\% of variance of \(E\), while humidity deficit alone explained only 25\%, and their product 98\% (Table 3). Similar to Manso Reservoir, evaporation was highly dominated by wind at Lake Victoria (MacIntyre et al. 2002) and Lake Tanganyika (Verburg and Hecky 2003), both tropical lakes. As a contrasting example, for Great Slave Lake, a high latitude boreal lake in northern Canada (\(62^\circ N\)), 77\% of the variability of \(E\) is accounted for by the humidity deficit alone, while wind alone explains 33\%, and their product 84\%. Note, however, that
winds and the humidity deficit were not independent variables; high winds were normally associated with cold and dry air masses (Blanken et al. 2000). Such correlation is not present in Manso Reservoir, where $U$ and $q_a - q_s$ were weakly correlated ($r^2 = 0.05$).

For Manso Reservoir, while $T_a - T_w$ has a dominant effect on $H$, the wind effect comes close to the air–water temperature differences in modifying the sensible heat flux (Table 4). A weak but significant correlation of $U$ with $T_a - T_w$ was observed during the observation period ($r^2 = 0.29$). At Lake Victoria, diurnal variations of $H$ were strongly tied to the changes in air–water temperature differences ($r^2 = 0.74$; MacIntyre et al. 2002), the same correlation observed for Manso Reservoir (Table 4). For Lake Valkea-Kotinen, a high latitude lake located in Southern Finland, neither $E$ nor $H$ showed strong correlations with wind alone; correlations between $E$ and $H$ with the products of wind and humidity deficit and air–water temperature differences, respectively, were also reduced (Nordbo et al. 2011). For Manso Reservoir, the almost perfect fit of $E$ against the product $U (q_a - q_s)$; $r^2 = 0.98$; Table 3) and $H$ against $U (T_a - T_w)$; $r^2 = 0.99$; Table 4) shows that $E$ and $H$ can be accurately estimated by these simple linear relations with a root-mean-square error (RMSE) of about 10% of both $E$ and $H$.

Another way of analyzing the relative influences of wind, humidity deficit, and air–water temperature differences on latent and sensible heat fluxes is to evaluate the dependence of the fractional variations of $E$ and $H$ ($dE/E$ and $dH/H$) on the fractional modulations of input variables, wind $U_z$ ($dU_z/U_z$), near surface moisture difference $\Delta q$ ($d\Delta q/\Delta q$), and air–water temperature differences $\Delta T$ ($d\Delta T/\Delta T$). The hourly values of these variables can be used to estimate the average values of the fractional changes, resulting in $dU_z/U_z = 0.68$, $d\Delta q/\Delta q = 0.2$, and $d\Delta T/\Delta T = 0.92$. Therefore, for the latent heat flux variability at Manso Reservoir, wind changes are much more important than the variations of $q_a - q_s$. In comparison with Great Slave Lake, we used wind and humidity values presented by Blanken et al. (2000) to estimate $dU_z/U_z \sim 2$ and $d\Delta q/\Delta q \sim 8$, an opposite condition as observed in Manso Reservoir. Although the fractional changes of wind at Great Slave Lake were much higher than for Manso Reservoir, the humidity deficit fractional changes dominated over wind influences. The wind dominance over the humidity deficit in controlling evaporation variability for a warm environment such as Manso Reservoir is somewhat puzzling considering that $q_a - q_s$ is higher for warmer water and air temperatures compared to temperate and high latitude lakes. However, the isolated effects of wind and humidity temporal changes on evaporation must be considered in the real situation when environmental conditions are changing over relatively short time scales, such as hourly, diel, or even synoptic. For a tropical inland waterbody such as Manso Reservoir, the relative changes of the wind are much higher than relative humidity deficit changes; the humidity deficit is large, but its variance is relative small compared to its mean. In contrast, cold lakes seem to have much more pronounced relative humidity deficit changes than relative wind changes. For the sensible heat flux variability at Manso Reservoir, the air–water temperature differences dominate over the wind, although wind variations are still significant.

As discussed in Lofgren and Zhu (2000), for a given available energy resulting from a net radiational flux (net shortwave minus net longwave fluxes), the Bowen ratio ($B$; the ratio between $H$ and $E$), gives the fraction of this energy that goes into sensible heat, which warms and directly affects the static stability of the atmospheric boundary layer and is a thermal forcing for the lake breeze. When $E$ is much larger than $H$, as in Manso Reservoir, the relative humidity of the near surface atmosphere can be enhanced and the cloud base height lowered. For the observation period at Manso Reservoir, in a diel mean time frame the Bowen ratio varied from a minimum of 0.06 ($H = 7$ Wm$^{-2}$; $E = 112$ Wm$^{-2}$) at midafternoon to a maximum 0.24 ($H = 39$ Wm$^{-2}$; $E = 162$ Wm$^{-2}$) at the end of the nighttime period. This substantial change in the Bowen ratio from midafternoon to nighttime (4-fold) is highly determined by a large increase of sensible heat flux (6.5-fold) produced by the joint effect of increases of both air–water temperature differences and wind speeds during the night. Latent heat flux also increases at night (1.53-fold), but much less than $H$.

The overall low values of $H$ show the predominant role of evaporative heat loss over sensible heating of the atmosphere, an anticipated result considering the high water temperatures. The diel changes of $H$ also indicate that at night, and particularly near sunrise, the effect of $H$ on the near surface atmospheric instability over the lake is considerably higher. Note that the Bowen ratio itself does...
Table 5. Hourly mean incident, emitted, and net longwave heat fluxes and input components. Units: Fluxes (W m$^{-2}$); $T_{air}$ and $T_{water}$ ($^\circ$C)

<table>
<thead>
<tr>
<th></th>
<th>Max</th>
<th>Min</th>
<th>$\Delta$ (%)</th>
<th>Local Time Max</th>
<th>Local Time Min</th>
</tr>
</thead>
<tbody>
<tr>
<td>$L_{w_{inc}}$</td>
<td>0.844</td>
<td>0.825</td>
<td>2.3</td>
<td>23:00</td>
<td>15:00</td>
</tr>
<tr>
<td>$T_{air}$</td>
<td>29.4</td>
<td>23.1</td>
<td>27.3</td>
<td>15:00</td>
<td>05:00</td>
</tr>
<tr>
<td>$T_{water}$</td>
<td>29.4</td>
<td>28.9</td>
<td>1.7</td>
<td>16:00</td>
<td>07:00</td>
</tr>
<tr>
<td>$L_{w_{inc}}$</td>
<td>385.9</td>
<td>370.1</td>
<td>4.3</td>
<td>17:00</td>
<td>09:00</td>
</tr>
<tr>
<td>$L_{w_{emi}}$</td>
<td>461.5</td>
<td>458.4</td>
<td>0.7</td>
<td>16:00</td>
<td>07:00</td>
</tr>
<tr>
<td>$L_{w_{net}}$</td>
<td>88.6</td>
<td>75.5</td>
<td>17.4</td>
<td>09:00</td>
<td>17:00</td>
</tr>
</tbody>
</table>

not depend on the atmospheric stability. Because $C_{p}$ has been assumed equal to $C_{cp}$, the modulations of these 2 exchange coefficients produced by variations of ABL instability are equal, and consequently the ratio of $H$ and $E$ is independent of the atmospheric instability. The low $B$ values observed in the afternoon at Manso Reservoir were similar to the average value $\bar{B} = 0.06$ observed at Lake Tanganyika (Verburg and Antenucci 2010); our typically higher nighttime $B$ values were closer to those observed during autumn at some high latitude lakes such as at Lake Huron ($\bar{B} = 0.33$), one of the North American Great Lakes (Lofgren and Zhu, 2000). The average $\bar{B} = 0.17$ for Manso Reservoir is similar to the average value ($\bar{B} = 0.13$) for the Northern Lake Victoria (MacIntyre et al. 2002).

The cloud cover fraction $C$ modulates the incident longwave flux, increasing it from its cloudless condition as given by equation 5. Daytime average diel variability of hourly values of $C$ estimated for Manso Reservoir using the CD99 methodology (Fig. 7) indicate that higher cloud cover fraction was present in the early morning hours ($\sim 0.5$), falling quickly to an almost cloudless condition ($\sim 0.2$) by 11:00–12:00 h and reaching a minimum at 14:00 h when $C$ began to increase toward the end of the afternoon. MacIntyre et al. (2002) also reported for Lake Victoria a higher cloud cover from midnight to morning hours during their observation period of April 1996. We presume that cloud formation over the Manso Reservoir was strongly influenced by ABL instability, which increases with increasing $T_a$; $T_w$ and $H$ (Fig. 6a and 6c). The daytime temporal behavior of $C$ was similar to that of $T_a$; $T_w$ (Fig. 6c). Using equations 4 and 5 with the daytime hourly values of $C$ estimated using CD99 approach and interpolated values at nighttime, we estimated longwave emitted and incident and net longwave fluxes.

Hourly mean net longwave flux (emitted minus incident) was positive (reservoir losing energy) throughout the day, varying from a minimum of 76 W m$^{-2}$ at 17:00 h to a maximum of 89 W m$^{-2}$ at 09:00 h (Fig. 8a; Table 5). Emitted and incident longwave fluxes (Fig. 8b) were almost in phase, with their maximum and minimum at about 16:00–17:00 h and 07:00–09:00 h, respectively. Due to the small diel variability of $T_a$ (Fig. 3a), $L_{w_{emi}}$ shows only a small magnitude variability of about 3 W m$^{-2}$ compared to a much greater change in $L_{w_{inc}}$ of about 16 W m$^{-2}$ (Table 5). The consequence was that the maximum $L_{w_{net}}$ was highly determined by the minimum $L_{w_{inc}}$, which occurs at 09:00 h.

We found that changes of air temperature and cloud cover jointly affect $L_{w_{inc}}$ along the day. The drop in $T_a$ (Fig. 3a) from 00:00 to 04:00 h dominated $L_{w_{inc}}$, although $C$ increases during the period (using the interpolated $C$ values). From 04:00 to 07:00 h, when $T_a$ was near its minimum with a small rate of change, the increase in $C$ dominated and $L_{w_{inc}}$ reached a local maximum. From 07:00 to 09:00 h, although $T_a$ was increasing, the rapid drop in $C$ dominated and $L_{w_{inc}}$ dropped. From 09:00 to 14:00 h, with a strong increase in $T_a$, $L_{w_{inc}}$ increased, although $C$ was dropping. From 14:00 to 17:00 h, the growth of both $T_a$ and $C$ combined to push $L_{w_{inc}}$ to its daily maximum. The local peak in $L_{w_{inc}}$ at 07:00 h, although weak (~4 W m$^{-2}$), was evident in the local minimum of $L_{w_{net}}$.

Fig. 7. Daytime hourly mean values of fraction of cloud cover $C$ estimated using the DC99 approach.

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If $T_a$ and $C$ are considered the independent variables, equation 5 can be used to calculate the fractional weights of each of these 2 variables to a fractional variation of $L_{w_{inc}}$:

$$\frac{dL_{w_{inc}}}{L_{w_{inc}}} = 4 \frac{dT_a}{T_a} + \left( \frac{0.34 C^2}{1 + 0.17 C^2} \right) \frac{dC}{C}.$$  \hspace{1cm} (11)

By equating the 2 right-hand side terms of equation 11, for specific values of $T_a$ and $dT_a$ we can estimate the necessary variation in $C$ ($dC$) that would match the variations in $T_a$ in changing the incident longwave flux. Using the local times of 05:00 and 06:00 h when $C$ was ~0.5 and there was a significant influence of $C$ in $L_{w_{inc}}$ (Fig. 8), the fractional change in $C$ that would have the same influence as the fractional air temperature change would be $dC/C = 1.3\%$. The estimated value from interpolated values of $C$ is ~6\%, 4.6 times greater. An opposite case occurred at 15:00 h, when $C$ was ~0.16 and another minimum in $dT_a$ was observed. In this case the needed $dC/C$ should be 48.3\%, compared to an observed fractional change of 14.4\%.

Air emissivity (Fig. 8c), which also modulates $L_{w_{inc}}$ (equation 5), depends nonlinearly directly on $e_a$ (Fig. 5a) and inversely on $T_a$ (Fig. 3a; Brutsaert 1975). By comparing the diel changes of (Table 5) and $L_{w_{inc}}$ its secondary effect on the incident longwave flux is evident.

The ratio of net longwave to latent heat fluxes varied from 52\% (at 19:00 h) to 74\% (at 14:00 h), and the maximum of $L_{w_{net}}$ was about 127\% higher than the sensible heat flux max. These figures clearly show the importance of the net longwave flux in the overall surface heat flux balance.

The diel variability of mean surface heat flux balance ($S$) for the analyzed period (Fig. 9) shows that the mean diel surface heat balance at Manso Reservoir was positive (lake warming) only for a period of about 8 h around solar noon. The diel cooling phase starts in midafternoon (16:30 h) with the increase in the wind and lasts until early morning (08:00 h); it is highly determined by the evaporative losses ($E$). For Lake Victoria the average period of positive surface heat flux balance was shorter, ~6 h around noon when winds tend to be weaker. Soon before and after this period, winds were high enough to enhance latent heat loss to 200–350 Wm$^{-2}$, making the heat balance negative.

At Manso Reservoir, a daily mean net heat flux of ~59.2 Wm$^{-2}$ was obtained by a numerical integration of $S$ along the day. The same procedure, but using neutral stability conditions (see curve $S_N$ in Fig. 9), leads to a

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**Fig. 8.** Diel variability of (a) net Longwave heat flux, $L_{w_{net}}$; (b) emitted $L_{w_{emi}}$ and incident $L_{w_{inc}}$; and (c) air emissivity $e_a$.

**Fig. 9.** Diel variability of the surface heat flux balance ($S$). Also displayed: net surface solar shortwave flux (and the sum of latent ($E$), sensible ($H$), and longwave ($L_w$) fluxes. For comparison, net surface heat flux under neutral conditions ($S_N$) is presented. Mean diel values calculated from all hourly dataset acquired by SIMA from 22 April to 20 May 2007.
Mean diel variability of surface energy fluxes over Manso Reservoir

Diel mean net heat flux of $-21.2$ Wm$^{-2}$, a clear indication of the significant role of the atmospheric instability on the calculated heat fluxes and net balance for Manso Reservoir.

Changes in surface water layer temperatures are not solely determined by the net surface heat flux; they also depend on the vertical profile of light attenuation, on the depth of the mixed layer (which is not constant throughout the day), as well as on vertical (upwelling and downwelling) and horizontal advection, inflow and outflow, and entrainment processes (Kim 1976, Imberger 1985, Imberger and Patterson 1990, MacIntyre et al. 2002). If diel mean variability of $T_w$ is highly correlated with $S$, however, then in principle, $S$ should be the dominant factor. In comparing the diel variability of $T_w$ (Fig. 3a) and $S$ (Fig. 9), $T_w$ started to increase at 08:00 h, coinciding with the beginning of a positive phase of $S$. The increase of $T_w$ ended at 17:00 h when $S$ became zero. From this time up to 07:00–08:00 h, when $S$ was negative and fairly constant, $T_w$ decreased until its minimum value at 07:00–08:00 h. Thus, at least qualitatively, we can state that the observed changes of $T_w$ were strongly controlled by the net surface heat flux.

The negative diel surface heat balance during the study period indicates that Manso Reservoir is losing energy and should present a gradual decrease in temperature. The daily mean surface water temperatures for the observation period dropped an estimated $-0.1$ °C d$^{-1}$ (Fig. 10). As a comparison, Lake Argyle, a large lake in Western Australia at about the same latitude (16.5°S) as Manso Reservoir, showed a similar temperature decrease during the month of May (figure 5 in Imberger and Patterson 1990). Superimposed on the gradual temperature decrease, 2 temperature drops can be noted (days 117 and 129) associated with cold front passages; however, on a scale of a few days, the frontal effect decreased and water temperature was observed to recover to previous values.

To relate more accurately the net surface heat flux to the diurnal variations observed in $T_w$, the Effective Surface Heat Flux ($S_{eff}$) must be considered. $S_{eff}$, the part of $S$ remaining for surface mixed-layer warming after removing the fraction of shortwave flux that propagates to deeper layers (Kim 1976, Imberger 1985, MacIntyre et al. 2002), is a function of the mixed-layer depth and of the light extinction coefficient $K$, a measure of the water clarity. An estimate of $K$ (m$^{-1}$) is possible from measurements of Secchi disk depth (SD) using the empirical relation $K = 1.1$ SD$^{0.73}$ (Williams et al. 1980). From a set of SD measurements obtained in Manso Reservoir from 19 March to 18 July 2007, we derived a mean SD of $-3$ m (characteristic of a mesotrophic waterbody) and a mean $K$ of 0.5 m$^{-1}$.

The vertical resolution of the available water temperature dataset was not sufficient to accurately determine the surface mixed-layer depth. From other studies and data collected in other similar reservoirs in the region, a diurnal mixed-layer depth between 0.5 and 1 m can be assigned. We estimated $S_{eff}$ assuming a 1 m depth diurnal mixed layer, $K = 0.5$ m$^{-1}$, and that 50% of the net solar radiation that penetrates the water is absorbed in the 0.5–1 m surface layer (Jerlov 1968, Kim 1976). At the maximum of incident shortwave flux just beneath the surface, $Sw(1 - \alpha_{sw}) = 650$ Wm$^{-2}$, the heat flux available for warming the mixed layer would be reduced from 440 Wm$^{-2}$, as given by the surface heat balance ($S$; Fig. 9), to about 107 Wm$^{-2}$; most of shortwave radiation penetrates to deeper layers. MacIntyre et al. (2002) reported an $S_{eff}$ maximum of 175 Wm$^{-2}$ near noon during late April for Lake Victoria when above water $Sw$ was near 1000 Wm$^{-2}$. For the night period $S_{eff} = S$.

**Discussion**

Large reservoirs and lakes undoubtedly play an important role as fresh water storage systems for hydroelectric energy generation and human–industrial–agricultural usage. Brazil is a large country with a relatively large population (currently about 200 million people), where the major fraction of its energy grid is hydroelectricity and its largest cities depend on reservoirs for their water supply; therefore, a better understanding of the physical workings of reservoirs and lakes and how these waterbodies interact with atmosphere via the surface energy fluxes is crucial. Furthermore, some of these large reservoirs in Brazil were built on previously heavily vegetated areas and consequently might be significant

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**Fig. 10.** Daily mean surface water temperatures at Manso Reservoir derived from hourly values acquired from 22 April to 20 May 2007 (solid line). Linear least square fit (dashed line).
sources of GHG emissions, the rate of which depends on factors such as the mixing dynamics and vertical stratification of the water column, both of which are dependent on the surface energy fluxes. The work described here was part of an effort to collect in situ data in Manso Reservoir and from it derive a better description of each heat flux component and the net surface heat flux in a mean diel time frame. We especially emphasize understanding the relative importance of each term of the surface energy budget and how the magnitude of each is controlled by input of physical variables throughout the diurnal cycle. The measured and computed variables for the austral autumn cooling period of Manso Reservoir was summarized and compared with other tropical and higher latitude inland waterbodies (Table 6).

The dominant component of the surface heat flux emissions (turbulent and radiational) at Manso Reservoir was the evaporative loss (latent heat). Higher $E$ values occur normally at night when winds are stronger and lower at midafternoon when winds are weaker. For the observed period, the average $E$ was 137.7 Wm$^{-2}$, similar to that observed for other temperate and tropical inland waterbodies. For April 1996 at Pilkinton Bay, northern Lake Victoria (0°18′N), and excluding high wind events, $E$ ~50–100 Wm$^{-2}$ (MacIntyre et al. 2002). Using a much more time- and space-comprehensive dataset of Lake Victoria, MacIntyre et al. (2014) expanded the previous findings and verified that flux values showed significant both east–west and north–south variability across the lake, revealing a strong seasonal influence of the regional monsoon meteorological regimes. Typical reported values were $E$ ~50 Wm$^{-2}$ at night with lower winds, and 100–150 Wm$^{-2}$ during the daytime with higher winds. The southeast monsoon had a strong influence in the north–south asymmetry of variables and fluxes. At the southern end of the lake, winds increased and relative humidity was reduced. The net effect was a monthly average $E$ in excess of 300 Wm$^{-2}$ in the south region and in the morning hours. For the tropical Lake Tanganyika (6°S), values of $E$ ~70 and ~180 Wm$^{-2}$ were reported for northern and southern portions of the lake, respectively (Verburg and Hecky 2003); 4: Verburg et al. (2011); 5: Liu et al. (2009); 6: Lofgren and Zhu (2000); 7: Rouse et al. (2008); 8: Blanken et al. (2000); 9: Nordbo et al. (2011); 10: Derecki (1976)

Table 6. An overview of surface flux components of Manso Reservoir and some temperate and higher latitude inland waterbodies.

<table>
<thead>
<tr>
<th>Source</th>
<th>$E$ (Wm$^{-2}$)</th>
<th>$H$ (Wm$^{-2}$)</th>
<th>$L_{w}$ (Wm$^{-2}$)</th>
<th>Bowen ratio</th>
<th>Main Influence over $E/H$</th>
<th>Source</th>
</tr>
</thead>
<tbody>
<tr>
<td>Manso Reservoir (14°50′S)</td>
<td>137.7</td>
<td>23.5</td>
<td>71.4</td>
<td>0.17</td>
<td>Wind/∆T</td>
<td>1</td>
</tr>
<tr>
<td>Lake Victoria (0°18′N–2°S)</td>
<td>50–100 (N)</td>
<td>0–20</td>
<td>30–60</td>
<td>0.13 (N)</td>
<td>Wind/Wind</td>
<td>2</td>
</tr>
<tr>
<td>Lake Tanganyika (3.5–9°S)</td>
<td>70–110 (N)</td>
<td>7(N)</td>
<td>58(N)-</td>
<td>~0.05</td>
<td>Wind/-</td>
<td>3</td>
</tr>
<tr>
<td>Ross Barnet Res. (32°N)</td>
<td>~100</td>
<td>24</td>
<td>~70</td>
<td>0.34</td>
<td>∆e/∆T</td>
<td>4</td>
</tr>
<tr>
<td>Great Lakes (41–48°N)</td>
<td>~50</td>
<td>~35</td>
<td>67</td>
<td>0.7</td>
<td>∆e/∆T</td>
<td>5; 10</td>
</tr>
<tr>
<td>GSL/GBL (62°N/66°N)</td>
<td>~80/~10</td>
<td>70/~8</td>
<td>—</td>
<td>0.88/0.8</td>
<td>∆e/-</td>
<td>6; 7</td>
</tr>
<tr>
<td>Valkea-Kotinen Lake (61°N)</td>
<td>2-25</td>
<td>0–20</td>
<td>—</td>
<td>0.74</td>
<td>∆e/∆T</td>
<td>8</td>
</tr>
</tbody>
</table>

$∆T = \text{air/water temperature difference}; \ ∆e = \text{humidity deficit}$

1: MacIntyre et al. (2002); 2: MacIntyre et al. (2014); 3: Verburg and Hecky (2003); 4: Verburg et al. (2011); 5: Liu et al. (2009); 6: Lofgren and Zhu (2000); 7: Rouse et al. (2008); 8: Blanken et al. (2000); 9: Nordbo et al. (2011); 10: Derecki (1976)
example, $E \approx 50 \text{ Wm}^{-2}$ was reported for the Great Lakes (Lofgren and Zhu 2000); for Great Slave Lake (GSL) and Great Bear Lake (GBL; 66°N), a global average value of $E \approx 40 \text{ Wm}^{-2}$ was reported, with values for GSL larger than for GBL (Blanken et al. 2000, Rouse et al. 2008); and for Lake Valkea-Kotinen (61°N), a small lake in Finland, $E \approx 2$–25 Wm$^{-2}$ (Nordbo et al. 2011). In comparison, the evaporative losses for Manso Reservoir are typical of other tropical lakes, and its maximum and minimum diel values are more similar to those of southern Lake Victoria.

Wind was the main factor controlling evaporative heat losses in Manso Reservoir; the humidity deficit only explained about 25% of the variability of $E$. A similar dominance of wind on the latent heat was observed for Lake Victoria (MacIntyre et al. 2002, 2014) and for Lake Tanganyika (Verburg and Hecky 2003, Verburg et al. 2011). This high wind dominance over $E$ was not observed in the Ross Barnet Reservoir; $E$ was not correlated with $U$ (Liu et al. 2009). In that reservoir, stronger wind events were associated with both positive and negative $H$ and $E$, which were caused by intrusions of cold or dry or warm and moist air masses, respectively. For the same air–water temperature differences, relative humidity, atmospheric pressure, and wind, the higher temperatures at tropical lakes increase the humidity deficit ($q_w - q_a$) and evaporation above those in high latitude and colder environment lakes. In tropical lakes, however, the fractional changes in humidity deficit ($\Delta (q_w - q_a) / (q_w - q_a)$) tend to be smaller than the fractional changes of wind ($\Delta U / U$), making the wind the dominant factor of changes on evaporation in tropical lakes. For higher latitude, cold-environment lakes, the role of wind increases, and the humidity deficit tends to control latent heat variability (Blanken et al. 2000, 2003, Nordbo et al. 2011). Manso Reservoir conforms to the scenario described earlier for tropical lakes with high evaporation rates that are mostly controlled by wind, as observed in Tanganyika Lake and Victoria Lake.

The second most important heat loss term for Manso Reservoir was the net longwave flux, with an average net longwave emitted flux of 71.4 Wm$^{-2}$. Not many publications include calculation or measurements of longwave fluxes over lakes and reservoirs, with a few exceptions. Derecki (1976), estimated $L_{\text{net}}^{\text{lw}}$ for Lake Erie based on 17 years of data, reporting an average net longwave for the full period of 48 Wm$^{-2}$ and for the autumn month of October of 67 Wm$^{-2}$, which is similar to our average estimate for Manso Reservoir. For Lake Victoria, a mean net longwave heat loss of 50 Wm$^{-2}$ with an autumn October range between 30–60 Wm$^{-2}$ was reported by MacIntyre et al. (2002, 2014). For northern and southern Lake Tanganyika, annual mean net longwave values of 58 and 71 Wm$^{-2}$, respectively, were reported by Verburg et al. (2011). Mean net longwave flux at Manso Reservoir, although about 30% higher, can be considered in the same range as for Lake Victoria. Manso Reservoir mean longwave flux for autumn is similar to the annual average value for southern Tanganyika. To reiterate the importance of net longwave radiation to the surface energy budget, its value was >50% of $E$ throughout the diel cycle.

Sensible heat flux was the smallest component of the surface heat flux losses in Manso Reservoir, particularly in the afternoon when the Bowen ratio decreased to 0.06. Diurnal average sensible heat flux was 28.5 Wm$^{-2}$, and combined with an average $E = 137.7$ Wm$^{-2}$ results in an average $B = 0.17$. Typical values of $H$ for Lake Tanganyika are 9–10 Wm$^{-2}$ (assuming a mean $E = 141$ Wm$^{-2}$ and $B = 0.06$; Verburg and Antenucci 2010). For Lake Victoria, excluding the high wind events, sensible heat flux was generally low, with $H = 0$–20 Wm$^{-2}$ (MacIntyre et al. 2002), roughly half of the diurnal range observed at Manso Reservoir ($H = 6$–39 Wm$^{-2}$). Using the average values of $H$ (10 Wm$^{-2}$) and $E$ (75 Wm$^{-2}$) for Lake Victoria, the mean $B$ was ~0.13, close to that for Manso Reservoir. A significantly higher sensible heat flux monthly mean for autumn ($H \approx 70$ Wm$^{-2}$ in October) was reported by Rouse et al. (2008) for the high latitude lakes GSL and GBL (66°N). Lakes Superior, Erie, Michigan, Ontario, and Huron show a smaller average value of $H \approx 35$ Wm$^{-2}$ for October (Lofgren and Zhu 2000). For the small, high latitude lake Valkea-Kotinen, sensible heat fluxes were in the range of 0–20 Wm$^{-2}$ during the day (Nordbo et al. 2011); however, for the similar-sized and high latitude Skeeter Lake (65°N), $H$ and $E$ were roughly double of those at Valkea-Kotinen due to nearly 2-fold higher winds (Spence et al. 2003, Nordbo et al. 2011).

High latitude lakes have a wide range of values for $H$ and $E$, with changes of sign ($+/-$) of monthly mean fluxes with the progression of seasons. Particularly during synoptic events occurring from fall through spring, which are associated with high winds, high latitude lakes tend to show higher sensible fluxes than tropical lakes (Derecki 1981, Verburg and Atenucci 2010). For large, low latitude lakes such as Tanganyika Lake and Victoria Lake (Verburg and Hecky 2003, MacIntyre et al. 2014), however, winds during the monsoon period are similar to those observed over northern temperate lakes during synoptic events (Liu et al. 2009). Perhaps more representative than absolute values for $H$ and $E$, high latitude lakes contrast with tropical ones by having much higher Bowen ratios, which can be positive or negative, and in some cases even higher than unity. In Manso Reservoir, although air–water temperature difference variability tends to dominate the sensible heat flux changes, wind variability is also significant. While air–water temperature differences alone explained about 75% of $H$ variance, wind alone explained about half of the variability of $H$. In contrast to Manso
Reservoir, the variability of hourly values of $H$ in Lake Victoria are dependent on both wind and air–water temperature differences, with a slight dominance of wind influence. For Ross Barnett Reservoir, the Great Lakes, and Valkea-Kotine, diurnal variations in $H$ were strongly tied to variations in $T_a - T_e$ (Lofgren and Zhu 2000, Liu et al. 2009, Nordbo et al. 2011). Similarly, in Manso Reservoir and Lake Victoria, both wind and air–water temperature differences are correlated with variations in sensible heat flux. Higher latitude lakes tend to have sensible heat flux variability mostly controlled by the air–water temperature differences.

For the observation period at Manso Reservoir, the daily integrated net surface heat flux was negative ($S = -59.2 \text{ Wm}^{-2}$), with the reservoir losing more energy at night than it was gaining during sunlight hours. This net daily energy loss was consistent with an observed gradual decrease of daily mean surface water temperature in the reservoir, an expected result for the austral autumn period of observation. Imberger and Patterson (1990) reported a negative net heat flux of about $-30 \text{ Wm}^{-2}$ for Canning Reservoir ($\sim 32^\circ \text{S}$) for the austral autumn, a value obtained using the bulk aerodynamic method for $E$ and $H$ and neutral transfer coefficients. We can therefore presume that, had the corrections for instability been used, the net heat flux would have been larger (in magnitude) and closer to that inferred for Manso Reservoir. The ABL over Manso Reservoir was unstable nearly all the time, and the instability parameter $\zeta$ was negative during almost all of the observation period. This finding was consistent with the surface water persistently warmer than air, wind typically low, and a hot and humid atmospheric boundary layer over the lake, conditions that resulted in a significant enhancement of turbulent fluxes ($E$ and $H$), which had the effect of increasing the magnitude of the net daily heat loss ($S$). The net daily heat loss calculated without the instability effect was reduced 64% from that including ABL instability.

The large change in Bowen ratio from a minimum of 0.06 in midafternoon to a maximum of 0.24 late at night implies that the energy stored in the upper mixed layer of Manso Reservoir is drained during day, mostly in the form of evaporative loss. At night, the evaporation loss is even higher, but sensible heat loss that occurs with a warming of the ABL cannot be neglected.

The present study adds to the published literature on surface energy budgets for tropical inland waterbodies, showing the similarities between Manso Reservoir and other tropical lakes and the contrasts with temperate and higher latitudes lakes. The mechanistic approach contributes to the analysis of possible effects of climate change on the energy exchanges between lakes or reservoirs and the atmosphere and to the interpretation of hydrodynamics within the water column. In a future scenario of higher air temperatures and lower relative humidity, evaporative losses, which are the dominant term in the surface heat budget, may become even larger. With evaporation peaking at night, we should expect nighttime increases in the mixed layer deepening, or even total water column mixing in shallower places, with consequently larger fluxes of GHGs (Crill et al. 1988). We anticipate that studies at additional locations within Manso Reservoir would show similar patterns but could show some spatial differences in heating and cooling, particularly in secondary sidearms of the reservoir, inducing exchange flows within the reservoir (Monismith et al. 1990, Xing et al. 2014). Studies in other seasons combined with time-series temperature measurements within the reservoir are still needed to describe the seasonal variability in the surface energy budget and how this changing budget moderates mixed-layer dynamics and controls the variability of GHG emissions by the reservoir.

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References

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