Surface energy fluxes, vertical mixing, and oxygen concentrations in bottom water in Lake Tutira

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NIWA Client Report No: HAM2012-152
Report date: November 2012
NIWA Project: ELF12212
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1. Executive summary

A surface heat flux budget was constructed from atmospheric and thermistor string data collected every 15 minutes at a stationary buoy at the deepest site in Lake Tutira. In addition, manual temperature and dissolved oxygen depth profile data collected in 2008-2011 and 1992-1996 were examined for relationships between net surface fluxes, heat content, temperature stratification and oxygen concentrations.

On average, the surface water temperature was 3.2 °C higher than the air temperature (for the year between 6 June 2009 and 5 June 2010), while the distance between the air temperature sensor and the water surface was only 1.5 m. Such a large surface to air temperature gradient results in large heat losses by latent and sensible heat fluxes.

The average net surface flux ($F_S$) was substantial (-54 W m$^{-2}$, 82% of the evaporative heat flux and 12% of total heat loss). The fact that $F_S$ was negative suggests that on annual average a substantial amount of heat is imported from elsewhere in the lake. The origin of the advected heat is likely the inshore margin of the lake. In shallow locations the water column absorbs heat faster relative to its volume during the daytime, simply because the volume per unit area is smaller. The warmer water is less dense than the cooler surface water in the deeper parts of the lake and proceeds to flow away from the shore and overflows the surface layer in the center of the lake, enhancing both the stratification and the heat content offshore. During the night, when lower air temperatures and absence of solar radiation tend to cool the surface waters, the opposite occurs, with relatively cool water generated at the shallow lake margin, but this mechanism appears less important in Lake Tutira, in view of the observed negative mean $F_S$ at the lake center.

Comparison of $F_S$ with the change in heat content suggested most of the lateral input of heat occurred during September to February, with the maximum input in November. In particular during spring and early summer when air temperatures start to rise, water inshore warms faster than offshore because of the greater volume offshore, supporting the hypothesis that the source for the lateral heat inputs is inshore.

The hypothesis of lateral inputs of heat from inshore lake margins and shallow areas to the deep site of the buoy to explain the large negative $F_S$ can be verified by applying simultaneously the same type of recorder inshore as exists offshore, except that only one thermistor is needed, near the surface. An inshore heat budget with a large positive $F_S$, exceeding local increase in heat content, would support the hypothesis.

Stratification is controlled by the heat content and its distribution in the water column, while $F_S$ controls the change in heat content and in stratification. In spite of the mean negative value of $F_S$, suggesting substantial lateral advection of heat, $F_S$ was found to correlate well with changes in heat content ($R^2 = 0.88$ for monthly data). Heat content correlated well with stratification as indicated by the temperature gradient between 1 m and 40 m depth ($R^2 = 0.67$ in the buoy data and $R^2 = 0.62$ in the manual profile data). $F_S$ also correlated directly with change in stratification as indicated by the daily change in the temperature depth gradient ($R^2 = 0.40$). The heat content explained a large proportion of the variation in the total amount of oxygen in the water column ($R^2 = 0.74$ in 2008-2011) and the oxygen concentration in the bottom water ($R^2 = 0.70$ in 2008-2011).
There was no oxygen in the bottom water for about 8 months of the year. Oxygen concentrations at 40 m depth start to increase as soon as the difference in temperature between the surface and the bottom has become about zero, reaching 6 mg L\(^{-1}\) in only 3 to 6 days. Oxygen concentrations at 40 m depth start to decline again about 7 weeks later when stratification is only just starting to develop.

Surface water was warmer by about 1 degree in 2008-2011 than in 1992-1996, reflecting a significant increase in air temperature in the region. The water at the interface of the metalimnion and hypolimnion was on average cooler in 2008-2011 than in 1992-1996. Negative temperature trends at the top of the hypolimnion resulted from an earlier onset of stratification in spring, later breakdown of stratification in autumn, and reduced vertical mixing, as has been seen in other warming lakes. Increasing surface temperatures and decreasing temperatures at the top of the hypolimnion indicate an increase in the density gradient across the thermocline. An increased density gradient would have resulted in decreased vertical mixing across the thermocline. Probably at least in part as a result of decreased vertical mixing across the thermocline oxygen conditions appear to have deteriorated since the 1990s with a decrease in average bottom water oxygen concentrations.

If algal production in the lake is nutrient limited, then phosphorus (P) is likely the dominant limiting nutrient because the average ratio of total nitrogen to total phosphorus in the surface layer (TN/TP ratio) is well above the average ratio in which algae require nutrients (TN/TP ~ 7), and increased in the surface layer from 19.0 in 1992-1996 to 21.6 in 2008-2012.

The concentrations of nutrients in the lake were higher in 2008-2012 than in 1992-1996, both in the bottom layer and in the surface layer, although no effect could be shown on algal biomass as indicated by the chlorophyll \(a\) concentrations. Nutrients concentrations in the bottom layer, already substantially higher than in the surface layer, increased proportionally more than in the surface layer since 1992-1996. Nutrient loading from the catchment to the lake may have increased since the 1990s, which would explain the increase in nutrient concentrations in the lake. In addition, in view of a decrease in the TN/TP ratio and especially a large decrease in the DIN/DRP ratio in the bottom layer compared with 1992-1996, it seems likely that phosphorus release from the sediment was relatively enhanced under the more severe anoxic conditions in 2008-2011. Loss of fixed nitrogen by denitrification, possibly enhanced as a greater sediment surface area is exposed to low oxygen conditions, may in part explain the reduced TN/TP and DIN/DRP ratios in the bottom layer as well. The ratios of TN/TP and DIN/DRP in the surface layer increased on the other hand, which may indicate reduced nutrient exchange between surface and bottom layers as a result of reduced vertical mixing.
2. Introduction

Temperate lakes of deeper than about 15 or 20 m typically stratify in summer. During stratification the oxygen concentrations in the hypolimnion decline as it is consumed by microbial activity while the stratified water layers prevent replenishment with oxygen from the surface layer. The rate of decline of oxygen in the hypolimnion has been used as an indicator of the trophic state of lakes (Burns and Rutherford 1998). Therefore, an improved understanding of stratification patterns and the interaction with the atmosphere through surface heat fluxes is an asset in the management of lake ecosystems. The purpose of the project was to help understanding of how oxygen concentrations throughout the water column are affected by stratification and vertical mixing and the abundance of algal biomass, and the understanding of how stratification in the lake is affected on diurnal and seasonal time scales by atmospheric forcing and to facilitate recommendations to on-going monitoring and management.

Hawke’s Bay Regional Council (HBRC) maintains automatic monitoring equipment in Lake Tutira that record water temperatures and oxygen in the water column and meteorological data at the surface of the lake. The available data can be used to examine stratification at hourly to seasonal time scales, the impact of stratification on the oxygen inventory and the association of stratification patterns with surface heat fluxes and the heat contents of the lake. In addition, oxygen consumption in bottom waters can be derived and the seasonal change in oxygen distribution in the lake can be linked to the relative abundance of algal biomass (as indicated by chlorophyll a). The request for advice on these matters by HBRC also provided the opportunity to compare temperature and oxygen and nutrient concentrations with historic data collected in 1992-1996 (Burns and Rutherford 1998) which enables us to determine whether significant changes have occurred in the lake.

The monitoring of Lake Tutira is part of the Regional Council SoE water quality monitoring strategy. The role of stratification in the oxygen inventory needs to be better understood to ensure that HBRC is focusing on the best way to manage this valued water body. This project will provide HBRC with an improved understanding of processes that affect oxygen concentrations in bottom waters of Lake Tutira and the complex interactions of atmospheric forcing, physical limnology and phytoplankton biomass, as a basis for defining impacts of catchment-scale anthropogenic activity. The improved understanding is intended to assist in policy and management planning, allowing changes in Council monitoring programmes to be realised that better reflects the monitoring needs of Lake Tutira. This report will be made publicly available through the Envirolink website.
3. Methods

3.1 Site description

Lake Tutira (39° 13' 30" S; 176° 53' 30" E; 39°221 S 176°893E in decimal coordinates; Fig. 1) is a eutrophic lake (McColl 1978; Stansfield 2010). The lake surface area is about 1.7 km$^2$ and mean and maximum depths are about 21 m and 42 m respectively. Surface elevation is 150 m above sea level. Residence time is 2 y and the catchment area (lake excluded) is 27 km$^2$ (McColl 1978).

Figure- 1: Google Earth image of Lake Tutira.
Figure-2: Bathymetry of Lake Tutira (Irwin 1978), with latitude and longitude of the position of the buoy station (red star).
3.2 Data collection: instruments and processing

A buoy with sensors and data recorder was anchored in August 2008 at the deepest site of the lake (Fig. 2) by the University of Waikato. Data were recorded between 9 January 2009 and 4 November 2011. A weather station (Vaisala WXT510) on the buoy, at 1.5 m above the water, recorded wind speed, wind direction, air temperature, relative humidity, air pressure, and solar radiation (280 to 2800 nm; Apogee PYR-S). Water temperature was recorded at the buoy with a Scienterra temperature string with sensors at depths of 0.5 m, 2 m, 5 m, 10 m, 15 m, 20 m, 25 m, 30 m, and 35 m below the surface, and 20 cm above the bottom. A depth pressure sensor was placed on the bottom at the site of the buoy on 12 October 2009. Bottom depth was on average 40.4 m (range 39.9 to 41.5 m). Two dissolved oxygen sensors (D-opto) were placed at 0.5 m and 40 m depth. Chlorophyll a, phycocyanin (blue-green algae pigment) and turbidity sensors (Turner C3) were placed at 0.7 m depth and data recorded in raw fluorescence units, which represent relative and not absolute measurements. The thermistor string was temporarily replaced between June and 12 October 2009 with a string with sensors at identical depths as before, after which the original thermistor string was restored. All data were recorded every 15 minutes, by an iQuest DS-4484 logger. Hourly means were computed as the means of the five data points starting 0.5 hour before the full hour up to 0.5 hour after the full hour. For the estimation of the surface heat fluxes we focused on the most continuous complete data set available, from 6 June 2009 to 18 July 2010. The surface heat fluxes were calculated using hourly input data \((n = 9335)\).

Full water column heat content \((\Theta, J \text{ m}^{-2})\) was calculated by linear interpolation of water temperatures between sensor depths on the buoy thermistor string.

Temperature \((n = 39)\) and dissolved oxygen \((n = 36)\) depth profiles collected with a WTW Profilino DO Oxi 197 by HBRC, more or less monthly from May 2008 to November 2011, were also examined, and compared with those collected by Burns and Rutherford (1998) between February 1992 and June 1996 \((n = 47\) for temperature and oxygen). Temperature and dissolved oxygen (DO) were measured in 1992-1996 with a YSI 5739 probe with stirrer and a YSI Model 58 DO meter, which gave a direct digital read-out of DO concentration and temperature (Burns and Rutherford 1998). These measurements were taken at depth intervals of 1 m. Where gaps existed in the sampled depths they were rarely more than 1 m. In order to prepare contour graphs of temperature and oxygen concentration and calculate oxygen inventories, data for unsampled depths at 1 m intervals down to 40 m depth were estimated by linear interpolation. Total number of data were \(n = 1880\) for temperature in 2008-2011 (37% by interpolation), \(n = 1880\) for oxygen in 2008-2011 (37% by interpolation), \(n = 1560\) for temperature in 1992-1996 (36% by interpolation), \(n = 1440\) for oxygen in 1992-1996 (42% by interpolation).

3.3 Surface thermodynamic calculations

Latent heat \((E)\) and sensible heat \((H)\) fluxes were estimated from surface water temperature, air temperature, wind speed, relative humidity, and air pressure, with bulk aerodynamic transfer equations (see next section). The latent and sensible heat fluxes are considered positive when there is a net loss of heat from the lake to the atmosphere and negative when heat is entering the lake. Incoming \((L_{\text{in}})\) and outgoing \((L_{\text{out}})\) long-wave radiation were estimated with standard equations from air temperature, relative humidity, cloudiness and water surface temperature (see section 3.6 for the methods), following Brutsaert (1982). \(L_{\text{in}}\)
and $S_a$ were adjusted for surface reflection ($\varepsilon$) and albedo ($\alpha$), respectively (Brutsaert 1982), to estimate the net radiative fluxes. The net long-wave radiation $L$ was estimated as the difference between the absorbed incoming ($L-\varepsilon$) and emitted outgoing long-wave radiative fluxes. Net radiation ($Q$) was estimated as the sum of the net (= absorbed) solar radiation ($S$) and $L$. The net surface heat flux, which represents the change in heat stored in the lake at the site of the measurements, was estimated as

$$F_S = Q - E - H$$

$F_S$ is positive when the lake gains heat and negative when heat is released to the atmosphere. Advection of heat by rivers into and out of the lake and by rainfall was ignored because heat gain and heat loss by hydrological terms are relatively small and more or less cancel out on an annual basis (Schertzer et al. 2000).

The energy balance components were calculated hourly for the period 6 June 2009 – 18 July 2010. This represents the period with the best data availability, allowing estimation of 95% of the hourly fluxes. The only substantial data gaps during this period occurred between 14 September and 2 October 2009, and on 12 – 13 July 2009. The annual means of the energy fluxes were estimated for the year from midnight 5 June 2009 to midnight 5 June 2010.

### 3.4 Bulk aerodynamic transfer method

Sensible and latent heat fluxes are linearly proportional to transfer coefficients, which vary with the local stability of the atmospheric boundary layer (ABL), which is affected by wind velocity and the gradients of temperature and humidity above the water surface. The drag coefficient and the transfer coefficients for latent and sensible heat were adjusted for the stability of the atmospheric boundary layer by applying stability functions (Verburg and Antenucci 2010; see next section). The Matlab code used to estimate $E$ and $H$ is made available online at: www.niwa.co.nz/our-services/software/heat-fluxes-from-lakes.

The algorithm is similar to the types of algorithms that are typically used for the computation of ocean surface fluxes (Zeng et al. 1998; Renfrew et al. 2002; Fairall et al. 2003). The algorithm differs from that applied in studies on the Laurentian Great Lakes (Derecki, 1981; Croley, 1989; Lofgren and Zhu, 2000; Laird and Kristovich, 2002) by providing roughness lengths of temperature and vapor (Brutsaert 1982), by allowing the roughness length of momentum to account for smooth flow (Zeng et al. 1998; Smith, 1988), in the formulation of the stability functions for stable conditions (Imberger and Patterson, 1990), and by adjusting air density for temperature and air pressure.

The measured variables used in the calculations are air temperature ($T$, °C), water surface temperature ($T_o$, °C), relative humidity ($R_h$, %), wind speed ($U_z$, m s$^{-1}$) and air pressure ($p$, mb). Sensible heat ($H$) and latent heat fluxes ($E$) are estimated with bulk aerodynamic methods (in W m$^{-2}$):

$$H = \rho_a C_a C_H U_z (T_o - T)$$

$$E = \rho_a L_v C_E U_z (q_s - q_z)$$

and evaporation (m s$^{-1}$) is estimated as

$$E' = E/\rho_w L_v$$
where $C_H$ and $C_E$ are the transfer coefficients for sensible heat and latent heat, respectively (see next section), $C_a =$ specific heat of air (1005 J kg$^{-1}$ K$^{-1}$) and air density (kg m$^{-3}$) is

$$\rho_a = 100\rho[R_a(T+273.16)]$$

(5)

The gas constant for moist air (J kg$^{-1}$ K$^{-1}$) is given by

$$R_a = 287(1 + 0.608 q_d)$$

(6)

Latent heat of vaporization (J kg$^{-1}$) is

$$L_V = 2.501 \times 10^6 - 2370 T_o$$

(7)

Specific humidity (kg kg$^{-1}$) is

$$q_Z = 0.622 e_a/p$$

(8)

while specific humidity at saturation (kg kg$^{-1}$) is given by

$$q_S = 0.622 e_{sat}/p$$

(9)

The vapor pressure (mb) is

$$e_a = R_h e_s/100$$

(10)

while the saturated vapor pressure at $T_o$ (mb) is

$$e_{sat} = 6.11 \exp^{[17.27 T o/(237.3 + T_o)]}$$

(11)

and the saturated vapor pressure at $T$ (mb)

$$e_s = 6.11 \exp^{[17.27 T/(237.3 + T)]}$$

(12)

Water density (kg m$^{-3}$) is given by Henderson-Sellers (1986) as

$$\rho_w = 10^3 x (1-1.9549 \times 10^{-5}|T_o-3.84|^{1.68})$$

(13)

Neutral transfer coefficients are dimensionless. The neutral drag coefficient is computed from

$$C_{DN} = (u^*/u_d)^2 = \kappa [\ln(z/z_o)]^2$$

(14)

and the neutral latent heat transfer coefficient

$$C_{EN} = \kappa^2/[\ln(z/z_o)\ln(z/z_E)]$$

$$= \kappa C_{DN}^{1/2}[\ln(z/z_o)]$$

(15)

where $z$ is the measurement height (1.5 m), $\kappa$ is the von Karman constant (0.41; Wuest and Lorke 2003). The transfer coefficient for sensible heat ($C_{HN}$) is assumed the same as for latent heat:

$$C_{HN} = C_{EN} (Zeng et al. 1998)$$

(16)

Air shear velocity (m s$^{-1}$) is estimated by

$$u^* = (C_d u_d^2)^{1/2} = \kappa u_d [\ln(z/z_o)]$$

(17)
The roughness length for momentum (m) is given by

\[ z_o = (au^2/g) + (0.11\nu/u) \]  \hspace{1cm} (18)

and the roughness length for vapor (m) by

\[ z_E = z_o \exp(-2.67Re^{1/4} + 2.57) \]  \hspace{1cm} (19)

where \( a \) is the Charnock constant (0.013; following Zeng et al. 1998), and \( g \) is the gravitational acceleration (9.81 m s\(^{-2}\)), and the roughness Reynolds number (m) is

\[ Re = u\cdot z_o/\nu \]  \hspace{1cm} (20)

Roughness lengths for temperature (\( z_T \)) and for vapor (\( z_E \)) are assumed the same (Zeng et al. 1998):

\[ z_T = z_E \]  \hspace{1cm} (21)

Kinematic viscosity of air (m\(^2\) s\(^{-1}\)) can be obtained as

\[ \nu = \frac{\mu}{\rho_a} \]  \hspace{1cm} (22)

where dynamic viscosity of air (kg m\(^{-1}\) s\(^{-1}\)) is given by the linear relation with air temperature (estimated using data in Montgomery, 1947; \( R^2 = 1.0, n = 7, p < 0.00001 \)):

\[ \mu = 4.94 \times 10^{-8}T + 1.7184 \times 10^{-5} \]  \hspace{1cm} (23)

To initialize \( u^* \) from wind speed at 10 m (\( U_{10} \)) an equation in Amorocho and De Vries (1980) is used:

\[ u^* = U_{10}(0.0015(1 + \exp[-(U_{10} + 12.5)/1.56])^{-1} + 0.00104)^{0.5} \]  \hspace{1cm} (24)

followed by a simple iteration loop which computes the roughness length for momentum \( z_o \), using the results to compute \( u^* \) with equation 17. This is fed back into the roughness length for momentum until it converges to within 0.001% of the previous value of \( z_o \) followed by the calculation of the neutral transfer coefficients. \( U_{10} \) in equation 24 is estimated by (Schertzer et al. 2003)

\[ U_{10} = U_2(10/z)^{1/7} \]  \hspace{1cm} (25)

and after obtaining \( z_o \) can be estimated by

\[ U_{10} = U_2\ln(10/z_o)/[\ln(z/z_o)] \]  \hspace{1cm} (26)

### 3.5 Adjustment of transfer coefficients for atmospheric stability

Atmospheric stability is a function of the Obukhov stability length \( L \) (m):

\[ L = \frac{-\rho_a u^3 T_v}{\kappa g \left( \frac{H}{C_a} + 0.61 \left( \frac{T + 273.16}{L_v} \right) E \right)} \]  \hspace{1cm} (27)
which is a measure of the ratio of the reduction of potential energy due to wind mixing and the growth of atmospheric stratification due to the heat flux (Brutsaert, 1982). When $L$ is negative the boundary layer is convective and vertical transport is enhanced. This is expressed by higher transfer coefficients. The opposite is true for stable stratified boundary layers ($L > 0$) when the heat and latent fluxes are reduced. In the neutral case, $L = \infty$ or $-\infty$, the stability parameter

$$\zeta = z L^{-1} = 0$$

(28)

and the atmospheric transfer coefficients are equal to their neutral values.

The virtual air temperature (K) is given by

$$T_V = (T + 273.16)[1 + 0.61 q_z]$$

(29)

and the virtual temperature of saturated air at the water surface (K) by

$$T_{oV} = (T_o + 273.16)[1 + 0.61 q_s]$$

(30)

while the virtual air-surface temperature difference is given by

$$\Delta \Phi = T_{oV} - T_V$$

(31)

The stability of the atmosphere above the water surface is accounted for in the mass transfer formulas by adjusting the transfer coefficients using the atmospheric stability functions ($\Psi$ functions; Brutsaert, 1982). There are different sets of $\Psi$ functions for stable and unstable conditions to adjust the transfer coefficients for stability:

**Stability functions for $\zeta > 0$ (stable atmosphere):**

$$\Psi_M = \Psi_T = \Psi_E = -5 \zeta$$

(0 < $\zeta$ < 0.5)

(32A)

$$= 0.5 \ zeta^2 - 4.25 \ zeta^1 - 7 \ ln \ zeta - 0.852$$

(0.5 < $\zeta$ < 10)

(32B)

$$= ln \ zeta - 0.76 \ zeta - 12.093$$

($\zeta$ > 10)

(32C)

**Stability functions for $\zeta < 0$ (unstable atmosphere):**

$$\Psi_M = 2 \ ln \ [(1 + \chi)/2] + \ ln \ [(1 + \chi^6)/2] - 2 \ arctan \ \chi + \pi/2$$

(33A)

$$\Psi_T = \Psi_E = 2 \ ln \ [(1 + \chi)/2]$$

(33B)

$$\chi = (1 - 16 \zeta)^{1/4}$$

(33C)

**Stability dependent transfer coefficients are obtained as:**

$$C_D = \kappa^2/[\ln(z/z_o) - \Psi_M]^2$$

(34)

for the drag coefficient, and

$$C_E = \kappa^2/[\ln(z/z_o) - \Psi_M][\ln(z/z_o) - \Psi_E]$$

$$= \kappa C_D^{1/2}[\ln(z/z_o) - \Psi_E]$$

(35)
for the latent heat coefficient, and
\[ C_H = C_E \] (36)
for the sensible heat coefficient.

Because \( L \) and therefore the \( \Psi \) functions are defined as functions of \( E \) and \( H \), the calculations of \( E \) and \( H \) are initiated with neutral fluxes \( E_N \) and \( H_N \) (by using \( C_{DN} \) and \( C_{EN} \)) and solved by iteration until the values converged (Hicks, 1975; Renfrew et al. 2002). In each iteration \( u^*, z_0, z_T, z_E \), the drag and transfer coefficients and \( H \) and \( E \) are recalculated and used to recalculate \( L \) and the \( \Psi \) functions. The stability calculations were iterated until \( L \) converged to within 0.001% of the previous \( L \) - generally ten iterations were sufficient.

Bounds can be imposed on \(| \zeta |\) (for instance \(| \zeta | = 15\) suggested by Imberger and Patterson, 1990; MacIntyre et al. 2002) because \( C_E, H \) and \( E \) can behave erratically at extreme values of \( \zeta \) (Smith 1988; Fairall et al. 1996), although this usually does not make much difference. The surface heat fluxes were calculated using hourly input data \((n \sim 9335\) for each of the budget components).

### 3.6 Shortwave and long-wave radiative fluxes

Albedo \( \alpha \), the reflectance of solar radiation, increases as the solar angle decreases. To estimate \( \alpha \), a function of solar altitude (Brutsaert 1982) was used:
\[ \alpha = 1.18 \Lambda^{-0.77} \] (37)

Solar altitude \( \Lambda \) (in degrees) was calculated from latitude, the time of the day and Julian day number \((J = \text{day of the year})\) following Martin and McCutcheon (1999) as
\[ \Lambda = \frac{180 \lambda}{\pi} \] (38)
where \( \lambda \) is the solar altitude in radians
\[ \lambda = \tan^{-1}\left[ \frac{\Omega}{\sqrt{1 - \Omega^2}} \right] \] (39)
where
\[ \Omega = \sin\left(\frac{\pi \theta}{180}\right) \sin(\delta) + \cos\left(\frac{\pi \theta}{180}\right) \cos(\delta) \cos(\omega) \] (40)
where \( \theta \) is the latitude (in degrees, negative for southern hemisphere), \( \omega \) is the solar hour angle (radians) and \( \delta \) is the declination of the sun
\[ \delta = \frac{23.45 \pi}{180} \cos\left(\frac{2 \pi}{365} \left(172 - J\right)\right) \] (41)

Reflected short-wave radiation was calculated by
\[ S_{out} = \alpha S_{in} \] (42)
and net short-wave radiation \((\text{W m}^{-2})\) by
Incoming and outgoing long-wave radiation was estimated using air temperature, relative humidity, cloudiness and water surface temperature. Emitted long-wave radiation (W m\(^{-2}\)) was estimated as

\[ L_{\text{out}} = \varepsilon_o \sigma (T_o + 273.16)^4 \]  

(44)

where surface emissivity \(\varepsilon_o = 0.97\) (Brutsaert 1982), and \(\sigma\) is the Stefan-Boltzmann constant (5.6697 \(\times 10^{-8}\) W m\(^{-2}\) K\(^{-4}\)).

Incoming long-wave radiation (W m\(^{-2}\)) was estimated by

\[ L_{\text{in}} = \varepsilon_a \sigma (T_o + 273.16)^4 (1+0.17C^2) \]  

(45)

where \(\varepsilon_a\) is the emissivity of air

\[ \varepsilon_a = 1.24[e_a/(T_o + 273.16)]^{1/7} \]  

(46)

and \(C\) is cloudiness. \(L_{\text{in}}\) increases with cloudiness, which was estimated by examining the distribution of the ratio of daily mean solar radiation (\(S_{\text{in}}\)) at the surface and at the top of the atmosphere (extraterrestrial solar radiation \(S_o\)). Clear sky (0% cloudiness) and fully overcast sky (100% cloudiness) insolation corresponded at the measurement location with 0.60 \(S_o\) and 0.03 \(S_o\) respectively. The values 0.60 and 0.03 therefore corresponded to the maximum and minimum proportions of \(S_o\) reaching the surface (daily \(S_{\text{in}}/S_o\)). Daily mean cloud cover was determined by adapting and rewriting equation 3.55 in Eagleson (1970) as follows:

\[ C = 1 – \left[\frac{(S_{\text{in}}/S_o)-0.03}{0.6-0.03}\right] \]  

(47)

The extraterrestrial solar radiation \(S_o\) (W m\(^{-2}\)) was estimated from latitude, longitude, the time of the day and \(J\) (Martin and McCutcheon 1999) and a solar constant (\(S_C\)) of 1366 W m\(^{-2}\) (IPCC 2001):

\[ S_o = \frac{S_C}{r^2} \left[ \sin\left(\frac{\pi \theta}{180}\right) \sin(\delta) + \frac{12}{\pi} \cos\left(\frac{\pi \theta}{180}\right) \cos(\delta) \sin(h_e) \right] \Gamma \]  

(48)

where \(h_e\) is the solar hour angle (radians) at the end of the period over which \(S_o\) is being calculated (a 1 hour period), \(h_b\) the solar hour angle (radians) at the beginning of the period over which \(S_o\) is being calculated, \(\Gamma = 0\) if the remainder of Eq. 48 returns a negative value and \(\Gamma = 1\) if the remainder of Eq. 48 is positive, and \(r\) is the relative earth-sun distance (Martin and McCutcheon 1999)

\[ r = 1.0 + 0.017 \cos\left[\frac{2 \pi}{365} (186 – J)\right] \]  

(49)

The solar hour angles (radians) at the start and end of the period over which \(S_o\) is being calculated are given by Martin and McCutcheon (1999) as

\[ h_b = \beta + \gamma (2\pi) \]  

(50)

and
\[ h_r = B + \gamma_2 (2\pi) \tag{51} \]

where

\[ \beta = \frac{\pi}{12} \left( \left( h - 1 \right) - \Delta t_s + \gamma_3 12 \right) \tag{52} \]

and

\[ B = \frac{\pi}{12} \left( h - \Delta t_s + \gamma_3 12 \right) \tag{53} \]

where \( h \) is the hour of the day (1 to 24). The coefficient \( \gamma_1 = -1 \) when \( \beta > 2\pi \), \( \gamma_1 = 1 \) for \( \beta < 0 \) and \( \gamma_1 = 0 \) when \( 0 \leq \beta \leq 2\pi \). The coefficient \( \gamma_1 = -1 \) when \( B > 2\pi \), \( \gamma_1 = 1 \) for \( B < 0 \) and \( \gamma_1 = 0 \) when \( 0 \leq B \leq 2\pi \). The coefficient \( \gamma_2 = 1 \) for \( h \leq 12 \) and \( \gamma_2 = -1 \) for \( h > 12 \). \( \Delta t_s \) is the fraction of the 15-degree increment by which the local meridian is west of the standard meridian for the time zone (Martin and McCutcheon 1999):

\[ \Delta t_s = \frac{\gamma_4}{15} \left( 15 \varphi_{ST} - \varphi \right) \tag{54} \]

where \( \varphi \) is the longitude (positive for east, in decimal coordinates), \( \varphi_{ST} \) is the integer nearest to \( \varphi/15 \), and the coefficient \( \gamma_4 = -1 \) for west longitude and \( \gamma_4 = 1 \) for east longitude.

\( S_o \) was averaged over the day before applying to Eq. 47. Surface reflectance for incoming long-wave radiation was set to 0.03 (Brutsaert 1982) which means that 97% of incoming long-wave radiation is absorbed by heat uptake by the lake. Net long-wave radiation (W m\(^{-2}\)) was estimated as

\[ L = 0.97L_{in} - L_{out} \tag{55} \]

Net radiation (W m\(^{-2}\)) is the sum of net solar radiation \( S \) and net long-wave radiation \( L \)

\[ Q = S + L \]

\[ = S_{in}(1 - \alpha) + 0.97L_{in} - L_{out} \]

\[ = H + E + F_S \tag{56} \]

### 3.7 Water quality

Water samples for concentrations of nutrients and chlorophyll \( a \) were collected more or less monthly in the surface layer and in the bottom layer in February 1992 – June 1996 (by Burns and Rutherford 1998) and in May 2008 – July 2012 (by HBRC), with sampling depths of both layers chosen in the same way in both periods. When the lake was more or less isothermal samples of the surface layer were collected at 10 m depth. When the lake was stratified samples of the surface layer were a composite of 4 subsamples collected at different depths in the epilimnion (at depths of 20 cm, and \( \frac{1}{4}, \frac{1}{2}, \frac{3}{4} \) of the full depth of the epilimnion). The bottom layer was sampled generally at 25-30 m depth.
4. Results

4.1 Buoy data

Daily mean air temperature varied from 4.1 °C to 25.3 °C (mean 13.4 °C; Fig. 3), wind speed from 0.8 to 6.1 m s\(^{-1}\) (mean 2.2 m s\(^{-1}\)), relative humidity from 45% to 94% (mean 75%), and atmospheric pressure from 968 mb to 1019 mb (mean 997 mb). Air temperature was highest in February and lowest in June-August and seasonal maxima and minima in air temperature lagged behind those in solar radiation. There was no seasonality in air pressure and relative humidity but wind speed tended to be slightly higher in summer (Fig. 3).

The water level fluctuated frequently between October 2009 and November 2011, with about 1.6 m between the minimum and maximum levels (Fig. 4). Peaks in lake level were clearly linked to peaks in rainfall measured at the buoy (Fig. 4).

The yearly maximum hourly temperature at the surface (Fig. 5) was 26.42 °C, in February in 2010 and in January in 2011. There was a data gap from 7 January to 6 May 2011 and it is possible the maximum was not captured that year. The yearly minimum hourly temperature at the surface was 9.1 °C at the end of July in 2009 and slightly higher in August 2011 (9.5 °C). There was a data gap during the cool period in 2010 (19 July – 28 September). The lowest water temperature recorded in the near bottom water was 8.86 °C and the maximum temperature near the bottom was 11.26 °C.

From the end of February the mixed layer gradually deepened, becoming 10 m deep by the beginning of April, 15 m deep by the beginning of June, 20 m deep after the third week of June and the lake became thermally homogenous in early July (Fig. 5). As the mixed layer deepens, the deep water is initially warmed by mixing with the warmer shallower water, before cooling resumes. As a result, the bottom temperature is generally at its annual maximum at the end of the stratification, when the lake becomes thermally homogenous.

Bottom water cooling in July-August 2011, after stratification ended, was stronger than in 2009 and 2010, suggesting strong cooling by heat loss at the surface continued after the lake became thermally homogenous, resulting in strong vertical mixing in the winter of 2011. There was a data gap in 19 July – 28 Sept 2010 but the high bottom water temperature when recording resumed suggested less cooling had occurred in the interim during the winter in 2010, as in 2009 (Fig. 5). The bottom water temperature was higher in the stratified season of 2010-2011 than in the previous year (Fig. 5), and at the start of the recorded data, at the beginning of October, it was in fact already higher than it was at the end of the stratification in the previous season, when bottom temperature is generally highest of the year. This suggests that in the winter of 2010 no reset of the bottom temperature occurred during the data gap, in contrast to the strong bottom water cooling observed in the winter of 2011. Air temperature and other atmospheric data are not available either for July-August 2010; therefore the cause of the differences in bottom cooling between the years cannot be examined.
Figure-3: Daily mean meteorological data.
Figure-4: Daily mean bottom depth derived from the pressure sensor at the buoy and rainfall between 12 Oct 2009 and 4 November 2011.

Figure-5: Daily mean water temperatures.
4.2 Water temperature sensor accuracy

The rated accuracy of the water temperature sensors was ±0.1 °C. Some of the sensors may have drifted since they were calibrated. As a result, while typically water temperature at a deeper depth is expected to be usually cooler or at most the same as shallower temperatures, because the opposite would suggest a density instability which is usually small and short-lived, this was not the case for a number of the water temperature sensors (Fig. 6). The water temperature at 35 m depth was on average 0.03 °C higher than at 30 m depth (n = 16,360 hourly averaged values) which is a small difference relative to the rated accuracy of the sensor. However, the difference was >0.1 °C for 30% of the time and the maximum difference was 0.27 °C, exceeding the expected accuracy of the sensor. Because we don’t know what the actual difference in temperature was between 30 and 35 m depth the error may be larger (but not smaller). From 4 July 2011 recorded water temperature at 35 m depth was continuously higher than at 30 m depth, on average by 0.18 °C, and was even higher than at 25 m depth. This pattern also occurred throughout August 2009. The 15 m sensor consistently recorded substantially lower temperatures than the 20 m sensor during the second half of 2011 (on average 0.15 °C lower, maximum 0.25 °C lower), and in August until mid-September 2011 even lower temperatures than any other sensor. During August 2009, recorded bottom temperatures were higher than at 25-30 m depth and during 20 August – 12 September 2011 higher than at 30 and 15 m depth (Fig. 6).

From 12 October 2009 until 4 November 2011, 70% of the time the water temperature recorded at 2 m depth was higher than at 0.5 m depth (on average by 0.06 °C, maximum 0.28 °C), suggesting that either or both sensors had drifted (most likely the 0.5 m sensor, see Fig. 7). While prolonged higher temperatures at deeper depths are not realistic, where this occurred the difference between the two thermistor sensors was usually within a range of twofold the accuracy of both thermistors (±0.2 °C). On both example dates in Fig. 7 (25 May 2010 and 24 December 2010) the temperature recorded at 0.5 m depth was about 0.1 °C lower than at 2 m depth (Fig. 7). Furthermore, on 25 May 2010 (Fig. 7) recorded temperature at 10 m depth was higher than at 5 m and 0.5 m depth throughout the day.

The data presented in Fig. 4 and Fig. 5 are daily means which would mask spikes of shorter duration in the data. Occasional spikes in temperatures of up to about 3 degrees above the daily mean occurred in water temperatures at depths below where large diel effects can be expected (Fig. 7). The variability in temperature at 10 m depth on 24 December 2010 appeared to be reflected to some extent by variability in temperature at 15 m depth. On 25 May 2010 only the temperature recorded at 15 m depth showed strong variability (Fig. 7). It is not clear whether these infrequent spikes present sensor malfunction or lateral flow of relatively warm water.
Figure-6: Daily mean water temperatures.

Figure-7: Water temperatures every 15 minutes on 24 December 2010 (top panels) and 25 May 2010 (bottom panels). The right hand panels focus on the near surface.
4.3 Heat content, stratification and oxygen concentrations

During summer the temperature at 0.5 m depth can increase by up to about 4 °C during the diurnal heating period (Fig. 8). As the surface temperature increases on warm days the temperatures at 2 m and 5 m depth diverge. Typically after a few warm and calm days, when the wind speed increases the warmer surface water is mixed downward and temperature at 5 m depth increases until it is similar to that at 2 m depth. Days with strong diel surface water temperature increases agreed with high heat gain ($F_S$) during the middle of the day, while lower midday surface temperature occurs when either heat gain was lower during the day or cooling was strong during the previous night (Fig. 8).

The diel maxima in $F_S$ precede surface temperature maxima by several hours. For instance, in 6 to 17 December 2009 the maximum in $F_S$ occurred on average 4 hours before the maximum in the surface water temperature, but the lag time varied considerably, from 1.5 to 8.5 hour between these 12 days (see Fig. 8 for the data). The correlation was greatest with a delay of 5 hours ($r^2 = 0.10$, $p < 0.00001$). Perfect correlation between $F_S$ and surface temperature is not expected, because the value of $F_S$ is related to change in heat content, not to actual heat content. In addition, a proportion of the heat is mixed downward and mixing intensity varies, depending on stratification and wind forcing.

Often, variability in temperature at 5 m depth was anti-phased with surface temperature, with the temperature at 5 m increasing only at night when the surface cools and reduced near surface stratification allows mixing downwards of the heat gained at the surface during the day (Fig. 8, middle panels). Diel variability in near surface temperatures is usually far less during winter (Fig. 8).

Oxygen concentrations at 0.5 m depth were on average 9.8 mg L$^{-1}$ (Fig. 9). Oxygen concentrations at 0.5 m depth generally peaked at the end of daylight time close to 5 pm (Fig. 10), driven by O$_2$ production by photosynthesis, and were lowest at 5 am. The diurnal peak in dissolved oxygen at 0.5 m depth more or less coincided with the maximum in water temperature at that depth. Variability in dissolved oxygen at 0.5 m depth during the day is less during winter (Fig. 10) when photosynthetic productivity is probably less than in summer. Hourly mean values ranged from 3.8 to 18.0 mg L$^{-1}$, not including data between January and June of 2010.
Figure-8: Diel variability in water temperatures and $F_S$. Water temperatures at 0.5 m, 2 m and 5 m depth (every 15 minutes) and the hourly net surface heat flux ($F_S$) in summer (6-17 December 2009, top panels; and 3-14 February 2010, middle panels) and winter (14-25 July 2009, bottom panels; temperatures at 2 m and 0.5 m depth were identical in the data base).
Figure-9: Hourly mean oxygen concentrations at 0.5 m and 40 m depth.

Figure-10: Oxygen concentrations every 15 minutes. Periods correspond with those in Fig. 8.
Oxygen concentrations at 0.5 m depth fluctuated strongly during the day between January and June of 2010 (Fig. 9) and the recorded percentage saturation reached up to 250% during that time. There was much less variability in dissolved oxygen during the day in January-June 2011 compared with 2010 (Fig. 10). For instance, in 3-14 February 2011, the range in dissolved oxygen was less than 1 mg L\(^{-1}\) over the day, while in the same period in 2010 it ranged from 8 mg L\(^{-1}\) at night to 16 mg L\(^{-1}\) at the end of the afternoon (Fig. 10). The high variability in oxygen concentrations during the day in the first half of 2010 was probably an artefact caused by biofouling of the oxygen sensor due to lack of cleaning. There was no evidence for a prolonged algal bloom during these months, to explain high oxygen production by high photosynthetic rates during the daytime. Chlorophyll \(a\) concentrations, as measured at the buoy, were not particularly high during this period (see below).

The maximum oxygen concentration at 40 m depth was 9.3 mg L\(^{-1}\) (Fig. 9) and the maximum saturation about 80%. There was no oxygen at 40 m depth for about 8 months of the year (Figs. 9 and 11). Oxygen concentrations at 40 m depth start to increase as soon as the difference in temperature between the surface and the bottom has become about zero, and are initially ramped up very quickly, reaching 6 mg L\(^{-1}\) in only 3 to 6 days. Oxygen concentrations at 40 m depth start to decline again about 7 weeks later when stratification is only just starting to develop (Fig. 11). By the time oxygen at 40 m has disappeared the temperature gradient between the surface and the bottom water is about 4 °C, still less than a third of its annual maximum of about 14 °C.

The temperature gradient between the surface and the bottom water (on average at 40.2 m depth) was at its maximum of about 14 °C in December-February and was nearly zero for about 5 to 6 weeks in July-August. The annual average temperature gradient between the surface and the bottom was 6.9 °C. The water surface temperature was higher than the air temperature almost the whole year, except occasionally during winter when a reverse temperature gradient frequently occurred (Fig. 11). The temperature difference between the water surface (0.5 m depth) and the air was highest in December to April, about 4 to 5 °C, and was least in June-October (on average 2.2 °C in the months with sufficient data). On average, the surface water temperature was 3.2 °C higher than the air temperature (for the year between 6 June 2009 and 5 June 2010), while the distance between the air temperature sensor and the water surface was only 1.5 m. Such a large surface to air temperature gradient results in large heat losses by latent and sensible heat fluxes.

The total heat content per m\(^2\) at the site of the buoy lagged slightly behind the temperature gradient between the surface and bottom water, with the minimum heat content occurring when stratification just started to develop in August and the maximum when stratification started to decline (Fig. 12) at the end of February. In other words, after the lake had become fully mixed the lake continued to cool for a few more weeks, which means that the whole water column cooled. In summer, when the maximum stratification was reached, the lake continued to absorb heat from the atmosphere for another 4 or 5 weeks. The increase in heat content during December 2009 – February 2010, when stratification remained similar, was mostly absorbed at depths below 5 m, especially around 10 m depth (Fig. 5). In spite of the lag, the correlation between heat content and the temperature gradient was strong \((R^2 = 0.67, p < 0.000001)\).
Figure-11: Daily mean differences between surface temperature ($T_s$), air temperature ($T_a$) and bottom temperature ($T_b$), and oxygen concentrations at 40 m depth.

Figure-12: Full water column heat content compared with the difference in temperature between surface (0.5 m depth) and the bottom.
The concentration of chlorophyll $a$ as measured at the site of the buoy was highest in August-November 2009, and was generally relatively low in January-May of each year Fig. 13).
Figure-14: Daily mean radiative fluxes. Net solar radiation ($S$), incoming longwave radiation ($L_{in}$), outgoing longwave radiation ($L_{out}$) and net radiation ($Q$).
Figure-15: Daily mean sensible ($H$) and latent heat ($E$) fluxes and the net surface flux ($F_s$).
4.4 Surface flux energy balance

One year of data were analyzed to construct an annual surface flux balance, between 6 June 2009 and 5 June 2010. In this period 95% of the data were available to estimate the fluxes. Daily mean absorbed solar radiation ranged from 4 to 289 W m\(^{-2}\) (Fig. 14) and the annual average was 113 W m\(^{-2}\). The daily mean incoming longwave radiation ranged from 245 to 403 W m\(^{-2}\) (Fig. 14) and the annual average was 113 W m\(^{-2}\) (Table 1). The daily mean outgoing longwave radiation ranged from 349 to 434 W m\(^{-2}\) (Fig. 14) and the annual average was 388 W m\(^{-2}\) (Table 1). Daily mean net radiation ranged from -63 to 189 W m\(^{-2}\) (Fig. 14) and the annual average was 27 W m\(^{-2}\) (Table 1).

The atmospheric boundary layer (ABL) was 85% of the time unstable, enhancing loss of heat by sensible and latent transfer. The daily mean sensible heat flux ranged from -33 to 65 W m\(^{-2}\) (Fig. 15) and the annual average was 15 W m\(^{-2}\) (Table 1). The daily mean latent heat flux ranged from 1 to 271 W m\(^{-2}\) (Fig. 15) and the annual average was 66 W m\(^{-2}\) (Table 1).

Finally, the daily mean net surface flux ranged from -288 to 89 W m\(^{-2}\) (Fig. 15) and the annual average was -54 W m\(^{-2}\) (Table 1), amounting to a large net loss of heat from the lake at the central site of the buoy station. The total heat loss by sensible and latent fluxes and outgoing longwave radiation was 468 W m\(^{-2}\) (Table 1). The net heat loss at the surface (\(F_S\)) amounted to 12% of the total heat loss (Table 1).

Table 1: Estimated annual mean surface heat fluxes (W m\(^{-2}\)) at the site of the buoy station in Lake Tutira. \(F_S\) is also given as a percentage of total heat loss. \(\Delta\Theta = \) heat storage change, \(A = \Delta\Theta - F_S\) = horizontal advection (positive when heat is imported). For explanation of other symbols see Methods. The annual mean is for 6 June 2009 - 5 June 2010.

<table>
<thead>
<tr>
<th>Heat fluxes</th>
<th>W m(^{-2})</th>
</tr>
</thead>
<tbody>
<tr>
<td>(S_{in})</td>
<td>124.6</td>
</tr>
<tr>
<td>(\alpha_{S_{in}})</td>
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</tr>
<tr>
<td>(L_{in})</td>
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<td>(\varepsilon_{L_{in}})</td>
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<td>(L_{out})</td>
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<tr>
<td>(H)</td>
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<tr>
<td>(E)</td>
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<tr>
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<td>(\Delta\Theta)</td>
<td>4.3</td>
</tr>
<tr>
<td>(A)</td>
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</tr>
<tr>
<td>Total heat loss</td>
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</tr>
<tr>
<td>(F_S) (%)</td>
<td>12</td>
</tr>
</tbody>
</table>

The water column at the site of the buoy was warmer on midnight of 5 June 2010 compared with a year before, on average by 0.80 °C, and there was a net heat gain of 4.3 W m\(^{-2}\) over the year (Table 1). At midnight of 5 June 2010 the upper 15 m was about 2.3 °C warmer, while bottom water from 30 m down was 0.2 °C cooler. In other words, the lake was more stratified at the end of the year used for the estimation of the surface fluxes. The fact that the water column actually warmed over the year while \(F_S\) was negative means that lateral or horizontal heat import at the site of the buoy (\(A\)) was larger (58.6 W m\(^{-2}\)) than the net amount of heat lost at the surface (-54.3 W m\(^{-2}\)), because \(A = \Delta\Theta - F_S\) (Table 1).
The annual average loss of water by evaporation was 855 mm y$^{-1}$, as derived from $E$ (Eq. 4). This is higher than the estimate of 684 mm y$^{-1}$ of McColl 1978, which was extrapolated from the Tongoio Meteorological Station without measurements on the lake. Most evaporation occurred during summer and the pattern is almost identical to that of $E$ (Fig. 15), because $\rho_w$ and $L_v$ in Equation 4 vary very little.

The lake gained heat in the months September to February and lost heat in March to July (Fig. 16). In August 2009 the lake gained heat but in August 2011 there was a net loss of heat. Monthly mean $F_S$ correlated strongly with the monthly change in heat content (Fig. 16), as expected ($p < 0.00001$). However, the fact that the intercept of the regression does not pass through the origin of both axes (i.e., where both $F_S$ and $\Delta \Theta$ are zero) and that $F_S$ is most of the time negative suggests that a substantial amount of heat is advected by currents from elsewhere in the lake towards the site of the measurements.
4.5 Temperature and dissolved oxygen – Historical context

Isotherm depths estimated from more or less monthly temperature depth profiles from May 2008 to November 2011 show shallow mixed layers during summer, with the hypolimnion below 15 to 20 m depth, and a fully mixed water column during winter (Fig. 17 and Fig 18). Between February 1992 and June 1996 (data from Burns and Rutherford 1998) patterns were similar although the stratified period was somewhat longer in the more recent data series (Fig. 17).

Dissolved oxygen from monthly depth profiles from May 2008 to November 2011 (Fig. 18) show high concentrations in the winter time down to the bottom and low oxygen water during summer reaching as shallow as 5 to 10 m depth. Occasionally water with dissolved oxygen below 5 mg L$^{-1}$ reached close to the surface. However, the almost completely hypoxic to anoxic full water column on the first two sampling days in May and June 2008, with surface concentrations of 3.4 to 4.3 mg L$^{-1}$, was not seen at any other time. Stansfield (2010) remarked that these results were unreliable. Hypoxic water (DO < 4 mg L$^{-1}$) reached less shallow depths during summers in 1992-1996, with the top of the hypoxic volume generally below 10 m depth (Fig. 18).

The stratification as reflected by the difference in temperature between 1 and 40 m depth was slightly stronger on average in 2008-2011, compared with 1992-1996 (Fig. 19). The annual minimum heat content was on average higher in 2008-2011 compared with 1992-1996 (Fig. 19).

On average, both the surface layer and the bottom water was warmer in 2008-2011 compared with 1992-1996 (Fig. 20). In contrast, the water at the lower limit of the mean metalimnion has cooled, resulting in an increase in the density gradient across the thermocline. The water temperature at 1 m depth adjusted for seasonality, as given by the residuals of the fit of water temperature by day number (Fig. 21A), was on average 0.96 °C higher in 2008-2011 than in 1992-1996 ($t$ test, $p < 0.005$). The heat content adjusted for seasonality, as given by the residuals of the fit of heat content by day number (Fig. 21B), was on average 0.04 GJ m$^{-2}$ higher in 2008-2011 than in 1992-1996 but the difference was not significant ($t$ test, $p > 0.05$). The mean water column heat content did not change significantly because the water temperatures between 8 and 18 m depth decreased while the surface water and bottom water temperatures increased. The temperature gradient between 1 and 40 m depth adjusted for seasonality, as given by the residuals of the fit of water temperature by day number (Fig. 22A), was on average 0.66 °C higher in 2008-2011 than in 1992-1996 ($t$ test, $p < 0.005$). The temperature gradient between 1 and 15 m depth, adjusted for seasonality (Fig. 22B), was on average 1.20 °C higher in 2008-2011 than in 1992-1996 ($t$ test, $p < 0.0001$). The temperature gradient between 1 and 12 m depth, adjusted for seasonality (Fig. 22C), was on average 1.30 °C higher in 2008-2011 than in 1992-1996 ($t$ test, $p < 0.00005$). The increase in the temperature gradient equates to an increase in the density gradient in the metalimnion, which would have resulted in reduced vertical mixing across the thermocline.
Figure-17: Isotherms calculated from temperature profiles and dissolved oxygen in February 1992 – June 1996.
Figure-18: Isotherms and dissolved oxygen in May 2008 – November 2011.
Figure-20: Mean temperature profiles in 1992-1996 (n = 43) and 2008-2011 (n = 32) with time periods chosen so that full years were averaged.

Figure-22: Water temperature gradients for day of the year. A) The difference in water temperature at 1 m and 40 m depth for day of the year, in 1992-1996 and 2008-2011. The 4th order polynomial fit is for 1992-1996 ($R^2 = 0.94$). B) The difference in water temperature at 1 m and 15 m depth, for day of the year, in 1992-1996 and 2008-2011. The 4th order polynomial fit is for 1992-1996 ($R^2 = 0.87$). C) The difference in water temperature at 1 m and 12 m depth, for day of the year, in 1992-1996 and 2008-2011. The 4th order polynomial fit is for 1992-1996 ($R^2 = 0.83$).
The average DO, the difference in DO between 1 and 40 m depth, and DO at 40 m depth. Daynumber 1 is the first of January in 1992 and 2008.
The mean dissolved oxygen concentration (DO) was less in 2008-2009 compared with 1992-1996 (Fig. 23). For data from a comparable period for which reliable data exist in both data sets (day number 204 to 1423 in Fig. 23. Data of the first two unreliable oxygen profiles in 2008 [Stansfield 2010] were excluded), the mean DO was 5.9 mg L\(^{-1}\) in 1992-1996 and only 5.0 mg L\(^{-1}\) in 2008-2011. Similarly, the mean water column DO adjusted for seasonality (Fig. 24), was on average 1.13 mg L\(^{-1}\) lower in 2008-2011 than in 1992-1996 (\(p < 0.0005\)), as given by the residuals of the fit of mean DO by day number. The mean difference in DO between the surface and 40 m depth was higher in 2008-2009, 6.7 mg L\(^{-1}\), compared with 6.1 mg L\(^{-1}\) in 1992-1996. Consistent with this result, the mean DO at 40 m depth was less in 2008-2009, 2.9 mg L\(^{-1}\), compared with 3.5 mg L\(^{-1}\) in 1992-1996. The lower sampling frequency of the manual profiles (Fig 23) does not capture the rapidity of the reoxygenation of the bottom water shown by the buoy (Fig. 9) when the lake becomes fully isothermal. Because oxygen near the bottom goes to zero for long periods of time, with an abrupt increase when bottom water ventilation starts in winter, removing seasonality in DO in bottom water or in the difference in DO between the surface and 40 m depth by fitting polynomials or other types of standard fits was not possible.

Especially in the winter of 2011 oxygen in bottom water was apparently less replenished by vertical mixing than in previous years, as shown by the low oxygen concentrations at 40 m and the large differences in oxygen between the surface and the bottom water (in August to October, or day numbers 1276-1402 in Fig. 23; see also Fig. 18), although there was no evidence for reduced mixing in the winter of 2011 from the surface to bottom temperature gradient (Fig. 19). This is peculiar because bottom water cooling in July-August 2011 was stronger than in 2009 and 2010, as evidenced by temperatures recorded at the buoy, suggesting stronger mixing than in the two previous winters (Fig. 5). In addition, the oxygen data recorded at 40 m depth at the buoy showed that oxygen was replenished to the same level as in 2009 (the maximum in 2010 occurred during the data gap), although the depletion following the maximum was perhaps more abrupt than in 2009 (Fig. 9). Therefore, one or two of the manual measurements of bottom water oxygen concentrations around September 2011 may have been unrealistically low.

From Fig. 23 it is clear that the oxygen depletion rate at 40 m depth during summers in 2008-2011 was similar to that in 1992-1996. Also at 25 m depth oxygen depletion rates during summer did not differ (\(t\) test, \(p > 0.05\)) between 2008-2011 (44 to 50 microgram L\(^{-1}\) d\(^{-1}\), \(n = 4\)) and 1992-1996 (47 to 64 microgram L\(^{-1}\) d\(^{-1}\), \(n = 4\)).

The heat content explained variance in total oxygen inventory (given by the mean oxygen concentration in the 40 m water column) better (\(R^2 = 0.74\) in 2008-2011, \(R^2 = 0.65\) in 1992-1996 and \(R^2 = 0.67\) for the combined data; \(p < 0.000001\) [Fig. 25 and Fig. 26]) than the temperature gradient between 1 and 40 m depth (\(R^2 = 0.40\) in 2008-2011, \(R^2 = 0.26\) in 1992-1996). The heat content also explained variance in oxygen concentration at 40 m depth quite well (\(R^2 = 0.70\) in 2008-2011, \(R^2 = 0.57\) in 1992-1996), better than the temperature gradient between 1 and 40 m depth (\(R^2 = 0.40\) in 2008-2011, \(R^2 = 0.26\) in 1992-1996). Lastly, the heat content explained variance in the difference in oxygen concentration between 1 m and 40 m depth (\(R^2 = 0.45\) in 2008-2011, \(R^2 = 0.64\) in 1992-1996 and \(R^2 = 0.56\) for the combined data; \(p < 0.000001\) [Fig. 25]) better than the temperature gradient between 1 and 40 m depth (\(R^2 = 0.36\) in 2008-2011, \(R^2 = 0.38\) in 1992-1996). Data of the first two unreliable oxygen profiles in 2008 (Stansfield 2010) were excluded.
The positive relationship between heat content and the temperature gradient between 1 and 40 m depth was strong ($R^2 = 0.62$ both for the 2008-2011 data and for the 1992-1996 data), in spite of the occasional lag of the heat content behind the temperature gradient (Fig. 27).

The daily mean $F_S$ correlated well ($R^2 = 0.40$) with the daily change in the temperature gradient between 1 and 40 m ($\Delta T_{S Tb}$) measured at the buoy, with the temperature gradient increasing for $F_S > 0$ ($t$ test, $p < 0.000001$) and vice versa (Fig. 28). Daily mean $F_S$ explained less of the variance in the daily change in heat content ($\Delta \Theta$, $R^2 = 0.30$).

**Figure-24:** The mean water column oxygen concentration (DO), for day of the year, in 1992-1996 and 2008-2011. The 4th order polynomial is for 1992-1996 ($R^2 = 0.88$).
Figure-26: Synchronicity in the total heat content and average dissolved oxygen (DO) concentration in the water column from manual profile data. A) in 1992-1996 ($R^2 = 0.65$) and B) in 2008-2011 ($R^2 = 0.74$). Day number 1 is the first of January in 1992 and 2008. Note inverted scale for DO.
Figure-27: Synchronicity in the total heat content and the temperature gradient between 1 and 40 m depth from manual profile data. A) in 1992-1996 ($R^2 = 0.62$) and B) in 2008-2011 ($R^2 = 0.62$). Day number 1 is the first of January in 1992 and 2008.

$y = 0.01x + 0.35$
$R^2 = 0.40$

Figure-28: The change in the temperature gradient between 1 and 40 m depth against $F_S$. 
4.6 Chlorophyll $a$ and nutrient concentrations

In 2008-2012 chlorophyll $a$ showed maximum abundance in early summer but this trend was not present in the 1992-1996 data (Fig. 29). Average chlorophyll $a$ was higher in September 2008 – July 2012 (5.3 mg m$^{-3}$) than in February 1992 – January 1996 (4.2 mg m$^{-3}$ with periods chosen so that the data sets present more or less full years) but the difference was not significant (two-tailed $t$ test, $p > 0.05$). There was also no significant change in Secchi depth (average 3.8 m) between 1992-1996 and 2008-2012 (not shown). Data of chlorophyll $a$ concentrations were too scattered over the year for residuals of trend fitting of results by day of the year (as for instance for temperature, Fig. 21) to be useful for adjusting for seasonality ($R^2 = 0.06$ for a 4th order polynomial). In 1992-1996 *Ceratium hirundinella* dominated phytoplankton biomass, followed by *Aulacoseira granulata* and *Anabaena* (Burns and Rutherford 1998). In 2008-2012 algal composition was not examined.

In 2008-2012 surface layer TP showed maximum abundance in spring or early summers but in 1992-1996 in late summer (Fig. 29). Average surface layer TP was higher in 2008-2012 (20.5 mg m$^{-3}$) than in 1992-1996 (16.5 mg m$^{-3}$, with periods chosen so that the data sets present more or less full years) but the difference was not significant ($t$ test, $p > 0.05$). However, the difference was weakly significant ($p = 0.06$) for residuals of a 4th order polynomial trend fit ($R^2 = 0.41$) of TP results by day of the year to adjust for seasonality, with higher concentrations in 2008-2011 than in 1992-1993 (Fig. 30).

In 2008-2012 surface layer TN showed maximum abundance around mid-summer but in 1992-1996 there was less clear seasonality (Fig. 29). Average surface layer TN was higher in 2008-2012 (442 mg m$^{-3}$) than in 1992-1996 (313 mg m$^{-3}$, with periods chosen so that the data sets present more or less full years) and the difference was highly significant ($t$ test, $p < 0.000001$). As a result of a proportionally larger increase in TN since the 1990s the TN/TP ratio increased from 19.0 to 21.6.

Surface layer dissolved reactive phosphorus (DRP) showed maximum abundance during the vertical mixing period in winter and this pattern was most clear in 2008-2012 (Fig. 31). Average surface layer SRP was higher in 2008-2012 (4.6 mg m$^{-3}$) than in 1992-1996 (2.4 mg m$^{-3}$) and the difference was significant ($t$ test, $p < 0.05$), but this may have more to do with changes in detection limits than with actual change. Also surface layer NO$_3$ showed maximum abundance during winter, always in August (Fig. 31). Average surface layer NO$_3$ was higher in 2008-2012 (20.6 mg m$^{-3}$) than in 1992-1996 (9.8 mg m$^{-3}$) and the difference was weakly significant ($t$ test, $p < 0.1$). Adjusting for seasonality by comparing residuals of a 4th order polynomial fit of NO$_3$ concentrations by day of the year did not improve significance of the difference between 1992-1996 and 2008-2011 ($t$ test, $p < 0.1$). Surface layer dissolved inorganic nitrogen (DIN = NO$_3$+NH$_3$) showed maximum abundance during midwinter, always in August (Fig. 31). Average surface layer DIN was higher in 2008-2012 (49.9 mg m$^{-3}$), especially at the winter maxima, than in 1992-1996 (20.0 mg m$^{-3}$) and the difference was significant ($t$ test, $p < 0.01$). As a result of a proportionally greater increase in DIN compared with DRP, the DIN/DRP ratio increased in the upper layer from 8.5 to 10.8.
Figure-29: Concentrations in the surface layer of chlorophyll \( a \) and nutrients. A) chlorophyll \( a \), B) total phosphorus (TP) and C) total nitrogen (TN) in February 1992 – June 1996 and May 2008 – November 2011. Day number 1 is the first of January in 1992 and 2008.
Figure-30: Total phosphorus (TP) in the surface layer, for day of the year, in 1992-1996 and 2008-2011. The 4th order polynomial is for 1992-1996 ($R^2 = 0.41$).
Figure-31: Nutrient concentrations in the surface layer. A) dissolved reactive phosphorus (DRP), B) nitrate (NO$_3$) and C) dissolved inorganic nitrogen (DIN = NO$_3$ + NH$_3$) in February 1992 – June 1996 and May 2008 – November 2011. Day number 1 is the first of January in 1992 and 2008.

Bottom layer TP increased during summer and dropped again rapidly as the mixing period started, and this pattern was prominent both in 2008-2012 and in 1992-1996, although no peak occurred in 1993 and in 2009 (Fig. 32). Average bottom layer TP was higher in 2008-2012 (32.1 mg m$^{-3}$) than in 1992-1996 (20.3 mg m$^{-3}$) and the difference was significant ($t$ test, $p < 0.01$). There was no clear seasonality in bottom layer TN (Fig. 32). Average bottom layer TN was higher in 2008-2012 (507 mg m$^{-3}$) than in 1992-1996 (343 mg m$^{-3}$) and the difference was highly significant ($t$ test, $p < 0.000001$). TN increased proportionally less than TP in the bottom water and as a result the bottom layer TN/TP ratio decreased from 16.9 in 1992-1996 to 15.8 in 2008-2012, in contrast to the increase in the TN/TP ratio that occurred in the surface layer. It appears that TP now accumulates more in the bottom water but a smaller proportion of the accumulated TP reaches the epilimnion.

Bottom layer DRP behaved as TP, increasing during summer and a rapid drop as the mixing period started, both in 2008-2012 and in 1992-1996 (Fig. 33). As for TP, no DRP peak occurred in 1993 and in 2009. Average bottom layer DRP was higher in 2008-2012 (19.6 mg m$^{-3}$) than in 1992-1996 (5.8 mg m$^{-3}$) and the difference was significant ($t$ test, $p < 0.0001$). Bottom layer NO$_3$ was much more variable than TN, and was in general highest in summer, and dropping by late summer (Fig. 33). The seasonal pattern was therefore different from...
that of DRP, possibly as an effect of denitrification. Average lower layer NO$_3$ was higher in 2008-2012 (121 mg m$^{-3}$) than in 1992-1996 (50 mg m$^{-3}$) and the difference was significant ($t$ test, $p < 0.0001$). DIN in the bottom layer, roughly half of which was NH$_3$, was quite variable with generally concentrations increasing during summer until fall (Fig. 33). Average lower layer DIN was higher in 2008-2012 (212 mg m$^{-3}$) than in 1992-1996 (110 mg m$^{-3}$) and the difference was significant ($t$ test, $p < 0.00001$). As a result of a proportionally greater increase in DRP, the DIN/DRP ratio decreased from 19.0 in 1992-1996 to 10.8 in 2008-2012 in bottom water, while it increased in the upper layer from 8.5 to 10.8. Therefore in 2008-2012 it was the same in the surface layer as in the bottom water while in 1992-1996 it was twice as high in the bottom water compared with the surface layer.
5. Discussion

On average, the surface water temperature was 3.2 °C higher than the air temperature (for the year between 6 June 2009 and 5 June 2010), while the distance between the air temperature sensor and the water surface was only 1.5 m. The difference between temperature of the water surface and the air is similar to that in other New Zealand lakes (unpublished data). Such a steep surface to air temperature gradient results in large heat losses by latent and sensible heat fluxes.

The average net surface heat flux was a substantial net loss (-54 W m⁻², 82% of \( E \) and 12 % of total heat loss, \( L_{\text{surf}}E+H \)). The fact that the average \( F_S \) was negative suggests that on annual average a substantial amount of heat is imported from elsewhere in the lake. The origin of the advected heat is likely the shallow inshore margin of the lake, or the northern half of the lake which is much shallower than the southern end where the buoy is located (Fig. 2). In shallow locations the water column can absorb heat faster relative to its volume during the daytime, simply because the volume per unit area is smaller. The resulting warmer water is less dense than the cooler surface water in the deeper parts of the lake and proceeds to flow away from the shore and overflows the surface layer in the centre of the lake, enhancing both the stratification and the heat content offshore. The mechanism is well known in lakes, and has been studied in particular in shallow embayments (MacIntyre et al. 2002) and reservoir side-arms (Monismith et al. 1990), and is illustrated in Figure 34B.

However, lateral heat exchange has been quantified in only few lake studies on annual time scales (Verburg et al. 2011). During the night, when lower air temperatures and absence of solar radiation tend to cool the surface waters, the opposite occurs, with relatively cool water generated at the shallow lake margin (Fig. 34A; Wells and Sherman 2001), but differential cooling appears less important in Lake Tutira in terms of net heat transport, in view of the observed negative mean \( F_S \) at the lake center.

If \( Q \), \( H \) and \( E \) were the only inputs and outputs of heat at the site of the buoy then \( F_S = \Delta \theta \) would be expected, at all times, meaning that the lake warms when \( F_S \) is positive and cools when \( F_S \) is negative. However, this was not the case (Fig. 16) which is best explained by the proposed lateral inputs of heat. \( F_S \) and \( \Delta \theta \) diverged most from parity in September to February, when the lake warmed at the site of the buoy (Fig. 16). Maximum divergence occurred in November. This suggests that most of the lateral input of heat occurred during these months. In particular during spring and early summer when air temperatures start to rise, water inshore warms faster than offshore because of the greater volume offshore, supporting the hypothesis that the source for the lateral heat inputs is inshore. The lateral inputs of heat from shallow locations to surface water in deeper locations advances the onset of stratification offshore and the onset of deoxygenation of deep water.

The hypothesis of lateral inputs of heat from inshore lake margins and shallow areas to the deep site of the buoy to explain the large negative \( F_S \) can be verified by applying simultaneously the same type of recorder inshore as exists offshore, except that only one thermistor is needed, near the surface. An inshore heat budget with a large positive \( F_S \) exceeding local increase in heat content, would support the hypothesis.
Figure-34: Differential heating and cooling. A). Differential cooling which generally occurs at night and which brings cool water along the bottom to deeper areas. B) Differential heating which generally occurs during the day and which brings warm water via the surface to deeper areas (Monismith et al. 1990). Arrows indicate the direction of the net heat flux per m$^3$ (nota bene: not per m$^2$) with larger arrows indicating a larger flux.

In spite of the low values for $F_S$, both $Q$ and $F_S$ may be overestimated (in other words, $F_S$ should actually be even more negative than estimated) if the water temperature recorded at 0.5 m depth was lower than the actual temperature, a possibility in view of the fact that water temperature recorded at 0.5 m depth was often lower than the temperature at 2 m depth from June 2009 onwards.

Stratification is controlled by the heat content and its distribution in the water column, while $F_S$ controls the change in heat content and in stratification. In spite of the mean negative value of $F_S$, suggesting substantial lateral advection of heat, $F_S$ was found to correlate well
with changes in heat content ($R^2 = 0.88$ for monthly data). Heat content correlated well with stratification as indicated by the temperature gradient between 1 m and 40 m depth ($R^2 = 0.67$ in the buoy data and $R^2 = 0.62$ in the manual profile data). $F_S$ also correlated directly with change in stratification as indicated by the daily change in the temperature depth gradient ($R^2 = 0.40$). The heat content explained a large proportion of the variation in the total amount of oxygen in the water column ($R^2 = 0.74$ in 2008-2011) and the oxygen concentration in the bottom water ($R^2 = 0.70$ in 2008-2011).

Surface water was on average about 1 °C warmer in 2008-2011 than in 1992-1996. The increase in surface water temperature reflected a regional increase in air temperature. Air temperature in the region has increased significantly since 1930 (ANOVA, $p < 0.000001$ using annual means; unpublished data in CliFlo, NIWA’s climate data base), and mean air temperature at Napier, about 30 km south from Lake Tutira, was about 0.5 °C higher in 2008-2011 than in 1992-1996. The water at the interface of the metalimnion and hypolimnion was on average cooler in 2008-2011 than in 1992-1996. Negative temperature trends at the top of the hypolimnion, as has been seen in other warming lakes (Rosner et al. 2012), resulted from an earlier onset of stratification in spring, later breakdown of stratification in autumn, and reduced vertical mixing. On the other hand, the increase in bottom temperatures can be explained by less severe winter cooling events.

Increasing surface temperatures and decreasing temperatures at the top of the hypolimnion indicate an increase in the density gradient across the thermocline. An increased density gradient would have resulted in decreased vertical mixing across the thermocline. Probably at least in part as a result of decreased vertical mixing across the thermocline oxygen conditions appear to have deteriorated since the 1990s with a decrease in average bottom water oxygen concentrations. Increased algal productivity caused by increased nutrient loading, and resulting in enhanced oxygen consumption in bottom water when organic matter decomposes after sinking out from the epilimnion, may also have contributed to a decrease in bottom water oxygen concentrations and mean water column oxygen concentrations. However, no direct evidence for increased algal productivity or enhanced oxygen consumption could be demonstrated.

If algal production in the lake is nutrient limited, then P is likely the dominant limiting nutrient because the average TN/TP ratio is well above the average ratio in which algae require nutrients (TN/TP ~ 7), and increased in the surface layer from 19.0 in 1992-1996 to 21.6 in 2008-2012.

The concentrations of nutrients in the lake were higher in 2008-2012 than in 1992-1996, both in the bottom layer and in the surface layer, although no effect could be shown on algal biomass as indicated by the chlorophyll $a$ concentrations. As a result of the increased nutrient concentrations the mean trophic state of the lake changed from mesotrophic with a mean Trophic Level Index (TLI; Verburg et al. 2010) of 3.8 in 1992-1996 to eutrophic ($TLI = 4.2$) in 2008-2012. However, mean chlorophyll $a$ and TP concentrations in 2008-2012 were still only about half the mean concentrations measured in the surface layer in 1973-1977 (Teirney 2009). TN concentrations were not recorded in 1973-1977 but the mean NO$_3$ concentration in the surface layer was 4.4 times higher in 1973-1977 than in 2008-2012. In 1973-1977 the lake was eutrophic ($TLI = 4.9$) based on the mean chlorophyll $a$ and TP concentrations. The TLI component for the TN concentration contributed more to the TLI than the components for chlorophyll $a$ and TP concentrations in 1992-1996 and in 2008-
2011, suggesting that the TLI in 1973-1977, if based on all three, would probably have been in the supertrophic range (TLI = 5 to 6).

Nutrient concentrations in the bottom layer, already substantially higher than in the surface layer, increased proportionally more than in the surface layer since 1992-1996. Nutrient loading from the catchment to the lake may have increased since the 1990s, explaining the increase in nutrient concentrations in the lake. In addition, in view of the decrease in the TN/TP ratio and especially the large decrease in the DIN/DRP ratio in the bottom layer, it seems likely that phosphorus release from the sediment is relatively enhanced under the more severe anoxic conditions in 2008-2011. Loss of fixed nitrogen by denitrification, possibly enhanced as a greater sediment surface area is exposed to low oxygen conditions, may in part explain the reduced TN/TP and DIN/DRP ratios in the bottom layer as well. The increase of the ratios of TN/TP and DIN/DRP in the surface layer, on the other hand, may indicate reduced nutrient exchange between surface and bottom layers as a result of reduced vertical mixing.

6. Acknowledgements

The lake buoy station was installed by the University of Waikato and was co-sponsored by Genesis, Fish and Game and Hawke’s Bay Regional Council. The buoy station is maintained by Hawke’s Bay Regional Council and Limnotrack. Oxygen and temperature depth profile data and data of nutrient and chlorophyll a concentrations were provided by Nina von Westernhagen. Anathea Albert helped with data processing and contour plotting of the oxygen and temperature depth profile data. This study was funded by Envirolink HBRC_144 and by MSI programme C01X1005-Management of cumulative effects of stressors on aquatic ecosystems.
7. Glossary of abbreviations and terms

\[ T = \text{air temperature} \]

\[ T_o = \text{water surface temperature} \]

\[ R_h = \text{relative humidity} \]

\[ U_z = \text{wind speed} \]

\[ p = \text{air pressure} \]

\[ \Delta T_{S-Tb} = \text{difference between surface and bottom water temperature} \]

\[ \Delta \Delta T_{S-Tb} = \text{change in the difference between surface and bottom water temperature} \]

\[ \Delta T_{S-Ta} = \text{difference between temperature of the water surface and the air at 1.5 m height} \]

\[ \text{DO} = \text{dissolved oxygen} \]

\[ F_S = \text{net surface flux} \]

\[ \Theta = \text{heat content} \]

\[ \Delta \Theta = \text{change in heat content} \]

\[ A = \text{horizontal advection} \]

\[ Q = \text{net radiation} \]

\[ S_{in} = \text{solar radiation (short-wave)} \]

\[ S = \text{net (= absorbed) solar radiation} \]

\[ L_{in} = \text{Incoming long-wave radiation} \]

\[ L_{out} = \text{outgoing long-wave radiation} \]

\[ \varepsilon = \text{surface reflection of long-wave radiation} \]

\[ \alpha = \text{albedo} \]

\[ L = \text{net long-wave radiation} \]

\[ E = \text{Latent heat flux} \]

\[ H = \text{sensible heat flux} \]

\[ E^* = \text{evaporation} \]

\[ C_H = \text{transfer coefficient for sensible heat} \]

\[ C_E = \text{transfer coefficient for latent heat} \]

\[ C_a = \text{specific heat of air} \]

\[ \rho_a = \text{air density} \]
\( R_a \) = gas constant for moist air
\( L_V \) = latent heat of vaporization
\( q_s \) = specific humidity at saturation
\( q_L \) = specific humidity
\( e_{sat} \) = saturated vapor pressure at \( T_0 \)
\( e_a \) = vapor pressure
\( e_s \) = saturated vapor pressure at \( T \)
\( \rho_w \) = Water density
\( C_{DN} \) = neutral drag coefficient
\( C_{EN} \) = neutral latent heat transfer coefficient
\( C_{HN} \) = transfer coefficient for sensible heat
\( z \) = measurement height
\( \kappa \) = von Karman constant
\( g \) = gravitational acceleration
\( u^* \) = air shear velocity
\( z_0 \) = roughness lengths for momentum
\( z_E \) = roughness length for vapor
\( a \) = Charnock constant
\( Re \) = roughness Reynolds number
\( z_T \) = roughness lengths for temperature
\( \nu \) = Kinematic viscosity of air
\( \mu \) = dynamic viscosity of air
\( U_{10} \) = wind speed at 10 m
\( L \) = Obukhov stability length
\( \zeta \) = stability parameter
\( T_V \) = virtual air temperature
\( T_{oV} \) = virtual temperature of saturated air at the water surface
\( \Delta \Phi \) = virtual air-surface temperature difference
\( \Psi_M \) = atmospheric stability function for momentum

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\( \Psi_T \) = atmospheric stability function for temperature
\( \Psi_E \) = atmospheric stability function for vapor
\( \chi \) = term in the atmospheric stability functions
\( C_D \) = Stability dependent drag coefficient
\( C_E \) = Stability dependent latent heat coefficient
\( C_H \) = Stability dependent sensible heat coefficient
\( J \) = day of the year
\( \Lambda \) = solar altitude (degrees)
\( \lambda \) = solar altitude (radians)
\( \Omega \) = component of the equation for solar altitude
\( \theta \) = latitude
\( \omega \) = solar hour angle
\( \delta \) = declination of the sun
\( S_{out} \) = solar radiation reflected at the surface
\( \varepsilon_o \) = surface emissivity
\( \sigma \) = Stefan-Boltzmann constant
\( \varepsilon_a \) = emissivity of air
\( C \) = cloudiness
\( S_o \) = solar radiation at the top of the atmosphere
\( S_C \) = solar constant
\( h_e \) = solar hour angle at the end of the period over which \( S_o \) is being calculated
\( h_b \) = solar hour angle at the start of the period over which \( S_o \) is being calculated
\( \Gamma \) = correction factor in the equation for \( S_o \)
\( r \) = relative earth-sun distance
\( \gamma_1 \) = coefficient in the estimation of \( S_o \)
\( \gamma_2 \) = coefficient in the estimation of \( S_o \)
\( \gamma_3 \) = coefficient in the estimation of \( S_o \)
\( \gamma_4 \) = coefficient in the estimation of \( S_o \)
\( \Delta t_8 \) = fraction of the 15-degree increment by which the local meridian is west of the standard meridian for the time zone
$\beta$ = component of the equation for the solar hour angle

$B$ = component of the equation for the solar hour angle

$h$ = hour of the day

$\varphi$ = longitude

$\varphi_{ST}$ = standard meridian for the time zone

TN = total nitrogen

TP = total phosphorus

DIN = nitrate + ammonia

DRP = dissolved reactive phosphorus

TLI = Trophic Level Index

CliFlo = NIWA’s climate database
8. References


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