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**VERTICAL MIXING AND FLOW PATTERNS IN A
TROPICAL, SHALLOW LAKE**

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**VERTICAL MIXING AND FLOW PATTERNS IN A
TROPICAL, SHALLOW LAKE**

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Executive Summary

This PhD study focused on the vertical mixing and the circulation patterns in a tropical shallow reservoir, Kranji reservoir. The thesis investigated the development of the Surface Mixed Layer (SML) and mechanisms driving the vertical mixing. Another finding is the different circulation patterns during the daytimes and the nighttimes in Kranji reservoir for the experimental study period which was during the Southwest monsoon and when the circulation pattern was wind driven, without the influence of tributary inflows. The spatial distribution of chlorophyll-a concentration as calculated by the Computational Aquatic Ecosystem Dynamics Model (CAEDYM) was lastly qualitatively correlated to the derived circulation patterns.

The vertical mixing analysis was carried out based on field data from three 24-hr Self Contained Autonomous MicroProfiler (SCAMP) measurements. It started with delineating the SML into three layers, a surface mixing layer, a diurnal thermocline and subsurface layer by the maximum temperature gradient, and by a constant temperature drop of 0.02°C from the surface. Kranji Reservoir exhibited a diurnal variation in the SML mixing. A daytime thin SML (<2 m) such as under strong heating was very much affected by the wind shear generated billows and the nighttime SML was dominated by penetrative cooling. The observations of turbulence parameters such as the vertical eddy diffusivities and, the mixing length scale should contribute to the future development of a more realistic biological model in tropical lakes e.g. a better estimate of the timescale of phytoplankton transport. Furthermore, in contrast to two thermoclines existing in temperate shallow lakes, only the diurnal thermocline exists in Kranji reservoir. More active mass exchanges over the water column in tropical shallow lakes are expected.

The temperature calculation based on the one-dimension vertical heat transport equation coupled with Xing et al's (2014) classification of the physical regimes in Kranji reservoir showed that the thermal structure was determined by the vertical mixing and heating during the first two 24-hr SCAMP measurement periods which

was under the influence of Xing et al (2014)'s cold inflow regime. The thermal structure was much affected by the horizontal transport during the last 24-hr SCAMP measurement period which was under the solar radiation dominated regime.

The flow circulation pattern was studied via both field measurement and numerical calculation using the Estuary, Lake, and Coastal Ocean Model (ELCOM). Both Acoustic Doppler Current Profiler (ADCP) and temperature chain measurements indicated a three-layer flow and thermal structure during the daytime and a two-layer structure during the night time. The ELCOM results showed that this diurnal pattern of flow and thermal structure was mainly explained by the upwelling generated by the southerly wind, typical during the Southwest Monsoon, and the horizontal temperature gradient induced by the differential cooling.

Lastly, the ELCOM-CAEDYM (ELCD) calculated spatial distribution of chl_a concentration during the study period corresponding to the Southwest Monsoon and with a further scenario extension to the Northeast Monsoon was found to be closely related to the circulation patterns obtained from the ELCOM calculation. The ELCD results were also in reasonable agreement with the PUB sampling results under both the southerly (Southwest Monsoon) wind and the northerly (Northeast Monsoon) wind.

This work demonstrated that weak changes in physical forcing (such as weak underflows as generated by differential cooling) could be significant to tropical shallow lake systems. This is because the dynamics of these lakes is fragile and variable due to the lack of strong seasonal weather (and variation) and having a more much more significant diurnal variation. Furthermore, this work presents a good example of high level physical-biological coupling in tropical lakes: the hydrodynamic influence on phytoplankton distribution. The ecosystem of the tropical shallow lakes is likely to be closely related to the prevailing monsoonal seasons as demonstrated for Kranji Reservoir where the flow pattern and the distribution of the chlorophyll *a* concentration were largely determined by the wind direction. Such responses of tropical shallow lakes to the monsoon seasons deserve

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- N : the buoyancy frequency of the background stratification [s^{-1}];
- D_T : the thermal diffusivity of water equal to $1.4 \times 10^{-7} \text{ m}^2/\text{s}$;
- ν : the kinematic viscosity of water equal to $10^{-6} \text{ m}^2/\text{s}$;
- K_B : the Batchelor wavenumber [m^{-1}];
- χ_T : the dissipation rate of temperature fluctuation [K^2/s];
- d' : the Thorpe displacement [m];
- Fr_T : the turbulent Froude number;
- Re_T : the turbulent Reynolds number;
- Γ : the mixing efficiency;
- R_f : the flux Richardson number ;
- Ri_g : the gradient Richardson number;
- u : the scale of the fluctuating velocity [m/s];
- l : the scale of the most energetic overturns [m];
- L_C : the central length scale of an overturn [m] ;
- L_K : the Batchelor's length [m];
- b : the vertical buoyancy flux [m^2/s^3];
- ϵ : the dissipation rate of TKE [m^2/s^3];
- α : the thermal expansion coefficient of water equal to $2.57 \times 10^{-4} \text{ }^\circ\text{C}^{-1}$;
- S : the velocity shear rate in vertical direction [m^2/s^2];
- U : the horizontal velocities obtained by aligning the ADCP measured velocities with the principal axes [m/s]
- K_z : the vertical eddy diffusivity [m^2/s];
- Q_{surface} : the surface heat flux [W/m^2];
- Q_{net} : the net radiation [W/m^2];
- Q_l : the latent heat flux by evaporative water flux [W/m^2];
- Q_s : the sensible heat flux [W/m^2];
- ρ_a : the air density equal to $1.2 \text{ kg}/\text{m}^3$;

C_{Pair} : the specific heat of air equal to 1012 J/(kg°C);
 C_s : the bulk transfer coefficient for sensible heat equal to 1.45×10^{-3} ;
 U_{10} : the wind speed at 10 meters above the water surface [m/s];
 T_a : the air temperature at 10 m above the water surface [°C];
 L_E : the latent heat due to evaporation equal to 2.453×10^6 J/kg;
 C_L : the bulk transfer coefficient for the latent heat equal to 1.45×10^{-3}
 q_s : the saturation specific humidity [kg H₂O / kg air] at the measured water surface temperature
 q_a : the specific humidity of atmosphere at 10 m above the water surface;
Pa: the pressure equal to 1012 mbar;
 ρ_0 : the reference density of water equal to 997 kg/m³
 q : the surface energy flux from wind and penetrative cooling [m³/s³];
 u_* : the wind-induced shear velocity [m/s];
 u_f : the buoyancy flux expressed in terms of velocity [m/s];
 C_D : the dimensionless drag coefficient equal to 1.3×10^{-3} ;
 C_T : the non-dimensional coefficient of shear production;
 C_k^f : the non-dimensional coefficient of wind stirring;
 η : the non-dimensional coefficient of convection;
 $\tilde{H}$: the surface cooling rate [W/m²];
 c_p : the special heat capacity of water equal to $4179 \text{ J} / (\text{kg} \cdot ^\circ\text{C})$;
 L_t : the Thorpe length scale [m];
 l : the sizes of the shear induced billows [m];
 ΔU : the velocity difference across the surface layer [m/s];
 g_0' : the reduced gravity across the interface [m/s²];
 J_b^0 : buoyancy flux from the surface buoyancy flux [m²/s³];
 J_T : the vertical transport of the vertical heat flux [W/m²];

$R(z)$: the heat source in the one-dimensional vertical heat transport equation
[W/m²];

B : the buoyancy flux from the tributary inflow [m⁴/s³];

B_{norm} : the normalized inflow-induced buoyancy;

Q : the sum of the inflows from the catchments into the reservoir [m³/s];

T_{lake} : the lake average temperature [°C];

A_s : the surface area of the Kranji Reservoir equal to 3x10⁶ m²;

ΔQ : the net inflow flow rate [m³/s];

J_{chla} : the vertical flux of chlorophyll a concentration [mg/(m²s)];

1. Introduction

1.1 Motivation and General Introduction

Tropical lakes, covering about 10% of lake surface area in the world (Lewis 1996) are unique because of their high water temperature, high irradiance and low variation in irradiance at the low latitude locations. Compared to numerous studies done on temperate lakes, work on tropical lakes is limited (Talling and Lemoalle 1999; Lewis 2000). However, such lakes are of key importance to the dense population in tropical regions as fresh water supply sources. One of the biggest threats to the fresh water system in tropical lakes is nuisance algal blooms that at times can even be harmful to human health. Research shows that thermal dynamics and flow dynamics, through their effects on the light and nutrient accessibility to phytoplankton, have a strong bearing on the growth of phytoplankton (MacIntyre 1993; Huisman et al. 1999; Wong et al. 2007). A better scientific understanding of the thermal and flow dynamics is thus key towards a long-term sustainable approach for maintaining water quality with implications on raw water or recreational water quality, and to better inform on reservoir management.

Kranji Reservoir (1°25'N, 103°43'E) which is located in the northwest corner of Singapore Island (see Figure 1.1) was selected as the study site in the proposed PhD work. Kranji Reservoir is one of 17 Singapore freshwater reservoirs supplying raw water to Singapore. As in many other tropical regions, the climate in Singapore is largely influenced by two monsoonal seasons. Kranji Reservoir located in a relatively undeveloped area in Singapore is free of the high buildings and less impacted by human activities than other parts of the island. The reservoir is also susceptible to algal blooms. It is thus a good example for an in-depth examination of the thermal and flow dynamics of a tropical lake responding to the monsoonal winds and the interaction between the hydrodynamics and the algal growth.

Figure 1.1 Locator map of Singapore (PUB 2011) with a zoom in to Kranji Reservoir and its catchments

1.2 Study site

Kranji Reservoir has a catchment area of about 5600 ha and the surface area of the reservoir is about $2.6 \times 10^6 m^2$. Its average depth is about 6.7 m. Figure 1.1 shows the location of Kranji Reservoir and its catchments. The reservoir is mainly supplied by the runoff brought by four tributaries, namely Sg. KangKar, Sg. Tengah, Sg. Peng Siang, and Sg. Peng Sua, and the direct rainfall on the reservoir. Like many other tropical lakes, Kranji Reservoir is a polymictic lake as diurnal cycles of thermal stratification and destratification were frequently observed at many locations inside Kranji Reservoir (Xing et al. 2014). Problems of algal bloom with a dominance of *Microcystis* were reported in the reservoir in the recent years (Lo 2008; Te and Gin 2011; Xing 2011). Among the few recent studies on Kranji Reservoir, Xing et al. (2012) constructed the water and heat balance for the reservoir both on shorter time scales of days and longer time scales of months and Xing et al. (2014) classified three thermal forcing regimes in Kranji Reservoir (a) solar driven, (b) windy and (c) cold inflow with the solar driven being the most commonly observed regime. However a detailed analysis of the vertical mixing and flow patterns inside Kranji Reservoir was not done which is performed in this work.

1.3 Study objectives and scope

This study focuses on an in-depth understanding of thermal structure and flow structure based on the field observation and the numerical study for a shallow tropical reservoir, that of Kranji Reservoir in Singapore.

Long-term and short-term field work were carried out to assemble a comprehensive, high quality dataset in the Kranji Reservoir simultaneously covering physical, water quality and algal parameters and thereby enable detailed study and fundamental understandings of the thermal variability and flow structures. The dataset combines (i) short term intensive sampled data via two short term (about 3 weeks in duration each) field deployment exercises with (ii) long term (months to years) observations as assembled from PUB¹ data and data collected by SysEng² for PUB. This combination of data observational time frames allows hypotheses derived from analysis of the short term data to be translated into long term observations and phenomena.

The dynamics in the surface mixed layer in Kranji Reservoir was investigated using the data assembled from 24-hr Self Contained Autonomous MicroProfiler (SCAMP) microstructure profiles and the meteorological data measured by a NTU-Stanford met station³ as collected during the short-term field exercises. The evolution and the turbulence structure in the surface mixed layer were examined to understand the fine-scale vertical mixing, evaluate the strength of the vertical mixing diffusivity and as well as its role in the vertical shaping of phytoplankton.

The Estuary, Lake, and Coastal Ocean Model (ELCOM) which is a three-dimensional hydrostatic numerical code, was applied to Kranji Reservoir for the July 2012 field work period to investigate the driving mechanisms of the observed temporal and spatial variations in the flow structures. This code was driven by

¹ PUB is the Public Utilities Board of Singapore responsible for the management of Singapore's reservoirs including Kranji Reservoir;

² SysEng is a private company engaged by PUB for their internal monitoring of the reservoirs including Kranji Reservoir.

³ The field work was sponsored by the Singapore-Stanford Program, a joint collaboration between Nanyang Technological University and Stanford University which allows Stanford field instruments to be used in Singapore.

meteorological data measurements from NTU-Stanford met station, inflow data simulated by the XP Stormwater & Wastewater Management Model (XP-SWMM, XP Software Inc 2005) based on PUB rainfall gauge data in the catchments, and temperature data measured by temperature chains deployed in the main channel and the tributaries. The performance of the ELCOM was evaluated by comparing with the field measurement of temperature and velocity in the horizontal (along-channel), transverse (cross-channel) and vertical directions. The thermal and flow structure calculated by ELCOM were later shown to be in good agreement with the field observation.

Next, the thermal and flow structures in Kranji Reservoir were studied based on the ELCOM results and the field observation. The effect of inflow on the thermal and flow structure was preliminarily investigated based on the field measurements. The flow circulation patterns in Kranji Reservoirs during the daytime as arising from a combination of surface stratification, wind (wind direction) and horizontal temperature difference during periods of no inflow/rainfall was investigated. The importance of differential cooling between the main body and the tributary to the thermal structure and flow structure during the nighttime was also assessed. The potential effect of the vertical mixing and the flow circulation patterns on the distribution of phytoplankton biomass was lastly evaluated.

1.3 Outline of the thesis

This thesis is in eight chapters including this Introduction Chapter. In Chapter 2, Literature Review, the vertical mixing mechanisms and flow structure of lakes, the properties of the turbulence, and relevant ELCOM studies are reviewed. At the end of Chapter 2, research questions are proposed to facilitate the understanding of the vertical mixing and flow patterns as well as their impacts on the algal dynamics in a tropical shallow reservoir such as Kranji. Chapter 3 provides details on the field experiments including the deployments, instrumentations and the field activities. In Chapter 4, a description is provided of the surface mixed layer analysis, the definition of the surface mixed layer, the driving mixing mechanism and as well as the vertical mixing in Kranji Reservoir. In Chapter 5, a numerical study of ELCOM

is described with comparison with the field measurement. Chapter 6 provides the thermal structure and flow structure in Kranji Reservoir as ascertained from the numerical study and field observation and with a further detailed description of flow pattern circulations during the Southwest Monsoon in Kranji Reservoir. In Chapter 7 the spatial and temporal distributions of chlorophyll a concentration is linked to the vertical mixing and the circulation flow patterns. Chapter 8 summarizes the major findings and proposes research work that could be done for the future.

2. Literature Review

A large variety of physical processes can be found in lakes in response to the driving forcing such as heating/cooling, wind, lake inflow, and the Earth's rotation. The vertical stratification/mixing, the wind-induced motion and internal wave motion are often the key features found in lakes. Deep temperate lakes have pronounced seasonality of vertical stratification during the summer and winter with complete mixing during the spring and autumn (Hutchinson and Loeffler 1956). The general cyclonic water circulation in large lakes can also present as increasingly affected by the Earth's rotation at increasing latitude (Hutter et al. 2011).

For tropical shallow lakes, the diurnal cycle of stratification and mixing can be more important than the seasonal cycle (Xing et al., 2014; MacIntyre, 2002) due to the lack of a strong seasonal weather (and variation) and the Coriolis forcing is negligible. The subsequent literature review thus focus on the aspects relevant to the PhD work: the vertical mixing responding to diurnal heating and nocturnal cooling, the flow structure influenced by the wind but in the absence of the Coriolis forces, and the Estuary, Lake, and Coastal Ocean Model (ELCOM) later used in this work to calculate the flow patterns in Kranji Reservoir.

2.1 Vertical mixing mechanism in tropical shallow lakes

The diurnal cycle of stratification and mixing was earlier investigated by Talling (1957) and has been reported extensively in tropical African lakes and in Amazonian floodplain lakes (Talling and Lemoalle 1998). During the daytime, the stratification in such lakes is enhanced by more rapid changes of water density at high temperatures and by the high light attenuation coefficient due to the high concentrations of phytoplankton or suspended substances (MacIntyre 2012). During the night time, the thermal stability decreases and the diurnal thermocline is destroyed by the convective cooling (MacIntyre and Melack 1982). We next review the vertical mixing under the background of diurnal heating and nocturnal cooling.

2.1.1 Diurnal stratification

The stratification dynamics in shallow tropical lakes varies significantly with time in response to the weather conditions. Limnological studies on the mixed-layer dynamics include Imberger (1985), Anis and Singhal (2006), Tedford et al. (2014) and Pernica et al. (2014). The surface mixed layer (SML) is defined as the envelope of maximum depths reached by the mixing layer on time scales of a day or more and is the zone that has been mixed in the recent past. Imberger (1985) first divided the surface mixed layer or epilimnion into a surface mixing layer, diurnal thermocline and a subsurface layer in-between. The depth of the diurnal thermocline or the depth of the SML can be found by the maximum density gradient (Imberger 1985). The temperature in the SML tends to be non-uniform with the degree of stratification depending upon time of day (Pernica et al. 2014). The mixing time scale T_{mix} within the SML also depends on its stratification: $T_{mix}=L^2/4K_z$ when the SML is stratified, and $T_{mix}=L/u_*$ or L/w_* . Here L is the depth of SML u_* is the wind-induced shear velocity and w_* is the convection velocity when the SML is unstratified.

In contrast to the mixed layer, the mixing layer is the depth zone being actively mixed from the surface at a given time and generally corresponds to the depth zone in which there is strong turbulence directly driven by surface forcing. Brainerd and Gregg (1995) found the measurement of the overturn scales, i.e. the Thorpe displacements L_T is a good indicator for the surface mixing layer. MacIntyre et al. (2002) chose a temperature difference from the surface of 0.02°C as the depth of the surface mixing layer in Pilkington Bay of Lake Victoria. Tedford et al. (2014) found that results of this temperature criterion and of the L_T criterion are similar and consistent with visual inspection of the microprofile data except when the surface mixing layer rapidly shoaled at the onset of heating or horizontal advection. Lombardo and Gregg (1989) quantified the wind-dominant turbulence within the surface mixing layer by modelling the dissipation rate as $\varepsilon = 1.8u_*^3 / \kappa z$ under the assumption that the TKE production and dissipation are in balance. Here u_* is the wind-induced shear velocity, κ is the von-Karman constant and z is the depth with positive downwards in the absence of stratification. Tedford et al. (2014) further

gave a new similarity scaling as $\varepsilon = 0.6u_*^3 / \kappa z$ during the heating. By considering the energy balance and the stratification, the mixing time scale of the wind-dominated turbulence is give in Imboden and Wüest (1995) $t \approx \frac{5}{3} N^2 h^3 \kappa / u_*^3$ where N is buoyancy frequency, h is the depth of the surface mixed layer.

The turbulence in the stratified regions below SML is closely related to internal waves. It is reviewed in Section 2.2.3 after the barotropic and baroclinic motions are described in Section 2.2.1 and Section 2.2.2 in preparation.

2.1.2 Nocturnal cooling

During night time, convection is initiated by the sinking of heavy cold water parcels from the surface, which deepen and erode the surface mixed layer. For example, strong convection deepened the SML in Wellington Reservoir, Australia with a reservoir with a mean depth of 30 m. The deepening from ~5 m to ~6 m occurred within a three-hour period from 03:00-06:00 (Imberger 1985). It also deepened the SML in Soppensee lake with a mean depth of 12 m, a small Alpine lake, with deepening from ~4 m to ~6 m within a six-hour period from 18:00-0:00 (Jonas et al. 2003). Such deepening rates of the SML driven by nighttime cooling can be tracked by the TKE budget (Fischer 1979) if one dimensional processes are dominating. It is noted that the temperate lakes have two thermoclines, i.e. diurnal thermocline varying on the daily time scales and the seasonal thermocline varying on the monthly time scales (Kannan and Job 1980; Monismith et al. 1990).

Because of the high evaporation rates, mixing caused by the cooling can extend to deeper depths in tropical lakes than in temperate lakes and conditions are less favourable for the establishment of seasonal stratification. MacIntyre et al. (2002) observed the diurnal surface mixed layer deepened to the water bottom (14m) in a shallow embayment, Pilkington Bay in Lake Victoria. The mixing can even extend to the bottom of main basin (60m deep) of Lake Victoria during the southeast monsoon (Fish 1957, Talling 1966, MacIntyre et al. 2014). However MacIntyre and Melack (1982) observed chemical stratification reduced the depth of nocturnal mixing in Sonachi Crater Lake. In addition, wind sheltering reduced the nocturnal

mixing depth because of the decrease in the turbulence kinetic energy for the deepening. Various studies (Imberger 1985; Anis and Moum 1992) showed the dissipation rate of TKE ε averaged over the convective mixed layer is proportional to the surface buoyancy flux J_b^0 : $\langle \varepsilon \rangle = aJ_b^0$ where a ranges from 0.45 (Imberger 1985) to 0.81 (Anis and Moum 1992). Tedford et al. (2014) found the dissipation rate $\varepsilon = 0.56u_*^3 / \kappa z + 0.77J_b^0$ during the cooling. However ε was not well predicted by this scaling during periods of weak and transient atmospheric forcing when buoyancy flux was transitioning between heating and cooling and when the wind speed was near zero. The corresponding mixing time scale of the convection-dominated turbulence could be calculated as $t \approx \frac{5}{2} N^2 h^2 / J_b^0$ where N is the buoyancy frequency and h is the thickness of the surface mixed layer.

2.2 Flow structure

2.2.1 Barotropic motion

Surface seiches occur in an enclosed or partially enclosed water body in the form of standing waves when gravity and the wind/atmospheric pressure induced water level displacement are in hydrostatic equilibrium. The velocity at which the waves propagate are described by the shallow wave velocity $\sqrt{gH}$, where H is the total water depth. The period of the surface seiches in a uniform water depth of H and completely enclosed rectangular basin is described by Merian's equation (Merian 1828):

$$T = \frac{2L}{N_o \sqrt{gH}} \quad (2.2.1)$$

where L is the length of the basin [m] and N_o indicates the number of nodes, points that experience zero vertical motion. N_o equals one for the fundamental mode, two for the bimodal mode, and so forth. Due to the complex geometry of typical lake basins, Merian's equation became inaccurate to characterize the seiches, especially for the higher modes. Various works have been made use of numerical models

based on the shallow water equations to simulate such surface seiches (Rueda and Schladow 2002; Cushman-Roisin et al. 2005; Mans et al. 2011).

Besides the surface seiche, upwelling and downwelling is another important phenomenon in response to the wind-driven surface currents in an enclosed or partially enclosed water body. The wind drives the surface water to converge towards a region (causing downwelling) and to move away from another region (upwelling). This influences the horizontal thermal/flow structure and the biological production in the water system. As upwelling causes water rising up from the thermocline to replace the surface water moved-away, the upwelling zone is colder than the downwelling zone where warm surface water converges. This potentially establishes a horizontal temperature gradient that initiates horizontal convection. The Wedderburn number W was proposed by Thompson and Imberger (1980) to estimate the possibility of upwelling, i.e. the relative importance of the wind force to the baroclinic pressure force. The Wedderburn number is defined as

$$W = g' d^2 / (u_*^2 L) \quad (2.2.2)$$

where g' is reduced gravity, u_* is the water shear velocity induced by the wind, L is the representative length of the lake, d is the depth of the thermocline. Increasing wind-induced shear velocity u_* or decaying stratification (smaller g') causes a decreasing W , which indicates a greater movement of the interface, i.e. upwelling is more likely. For multi-layered systems (e.g. a three-layered system), the lake number L_N (Imberger and Patterson 1989) is better suited to model the response as it accounts for a varying density profile in the stability term. The lake number is expressed as

$$L_N = \frac{St(1 - Z_T / H)}{u_*^2 A_0^{3/2} (1 - Z_V / H)} \quad (2.2.3)$$

where the stability of the lake St is given as

$$St = \frac{g}{\rho_0} \int_0^H (Z_V - z) A(z) \rho(z) dz \quad (2.2.4)$$

and Z_T is the height of the thermocline in m, Z_V is the height to the centre of volume in m, H is the total depth, A_0 is the lake surface area in m^2 . When $L_N < 1$, the wind is

of sufficient strength to induce the upwelling of the thermocline at upwind reaches of the lake (Yeates and Imberger 2003). For $L_N > 1$ the wind-generated turbulence is confined to the surface layer and for $L_N > 10$, there is no tilting of isotherm and therefore internal waves will not exist. Imberger and Patterson (1990) further differentiated the first and second vertical mode by values for W and L_N : mode one is characterized by small values for both W and L_N , while mode two results in small W and large L_N .

Xing et al. (2014) found the Wedderburn number and the Lake number developed from a lake system where a background stratification evolves seasonally may not be suitable for shallow tropical systems like Kranji Reservoir as the SML in Kranji Reservoir can deepen to the full water depth via convective cooling and no seasonal thermocline exists in the water column. Rather Xing et al. (2014) proposed another two dimensionless numbers important to characterizing the stratification dynamics in Kranji Reservoir: parameter I_{IW} representing the relative balance of inflow-induced stratification and mixing by wind stress, and I_{WB} representing the competing effects of the wind mixing and solar heating heating.

2.2.2 Baroclinic motion

In small and stratified lakes most of the energy that passes through the surface mixed layer and enters the main water column is transferred to basin-scale wave motions referred to as internal seiche (Imberger 1998; Wüest and Lorke 2003). The modes of internal seiche are determined by the density stratification and the bathymetry of the basin (Munnich 1996) and classified according to the number of nodal points ($V_i H_j$), where i and j are the number of vertical and horizontal nodes. The most commonly observed mode is V1H1 and its frequency is described as (Spigel and Imberger 1980),

$$T_{12} = \frac{2L}{N_o \sqrt{\frac{g' h_1 h_2}{H}}}, N_o = 1, 2, \dots \quad (2.2.5)$$

where h_1 and h_2 are the thickness of the two layers respectively, H is the total water depth and N_o is the number of nodes.

Observations of high-vertical mode internal seiche are seldomly reported but may occur frequently (Vidal et al. 2007). The two-dimensional governing equations of a stream function ψ can be used to describe the internal wave motions including high-vertical mode internal waves (Munnich et al. 1992; Vidal et al. 2005) in a hydrostatic Boussinesq fluid

$$\frac{\partial^4 \psi}{\partial t^2 \partial z^2} + N^2 \frac{\partial^2 \psi}{\partial x^2} = 0 \quad (2.2.6)$$

where x and z are the horizontal and vertical dimensions, t is time, and velocities $u = -\partial \psi / \partial z$ and $w = \partial \psi / \partial x$. The above equations can be converted to the generalized eigenvalue problem by assuming periodic time dependence $\psi = \Phi(x, z) e^{i(\omega t + \sigma)}$ where ω is the angular frequency of the seiche and σ is a constant. The expression of the converted generalized eigenvalue problem is then given as:

$$\frac{\partial^2 \Phi}{\partial x^2} = \frac{\omega^2}{N^2} \frac{\partial^2 \Phi}{\partial z^2} \quad (2.2.7)$$

Various methods such as Q-Z algorithm or the Schur-Raleigh-Ritz variant of inverse vector can compute the eigenvectors and eigenvalues of the above generalize eigenvalue problem. The eigenvalue is equal to ω^2 and the period of the internal seiche is $T = \frac{2\pi}{\omega}$. The eigenvectors further describe the internal wave field in a specified domain.

2.2.3 Internal waves and production of turbulence

In many limnological studies, the turbulence in the stratified regions below SML has been linked to internal waves (Saggio and Imberger 1998, MacIntyre et al. 1999). In a non-rotating system such as a long narrow lake where the effects of the Earth's rotation can be neglected, and when the wind duration $T_w > T_i/4$ (a quarter of the internal seiche time scale), the internal wave field may be decomposed into a

standing seiche, a progressive nonlinear surge, and a dispersive high-frequency internal solitary wave (ISW) (Boegman et al. 2005). When $T_w \ll T_i$, the basin-scale internal waves form at the lee shore, propagate as progressive waves across a lake (Thorpe et al. 1972) and generate high-frequency wave packs as they travel (Boegman et al. 2005). Field observations show that the wind-forced basin-scale waves decay at a rate far greater than can be accounted for simply by internal dissipation (Stevens et al. 1996, Saggio and Imberger 1998). The energy in the internal wave field dissipates rapidly and the high frequency wave energy is transferred to turbulence production via nonlinear steepening, shear instabilities, shoaling and reflection at sloping boundaries and interaction with topography (Horn et al. 2001). Further, the changes in internal wave field and turbulence in lakes can be related to the Wedderburn number or Lake number. When wind forcing further increases and so reduces L_N to ~ 2 for the lake as a whole, nonlinear steepening of the deeper pycnocline could occur and induce significant vertical fluxes (MacIntyre et al. 2009). Horn et al. (2001) developed a phase diagram based on the inverse Wedderburn number (W) and the ratio of mixed layer (h) to basin depth (H) to predict the type of internal waves and their degree of nonlinearity. When $W > 3$, the internal waves are linear damped wave; when $1 < W < \sim 3$ and $h/H < 0.5$ surges and solitons form with nonlinear steepening of the thermocline; when $W \approx 1$ and $0.3 < h/H < 0.5$ Kelvin-Helmholtz waves tend to occur and when $W < 1$ internal bores or hydraulic jumps are expected. A few attempts have been made to quantify the relation between the turbulence intensity and the internal wave field. Alford and Pinkel (2000) found the dissipation rate ε is in reasonable agreement with the change in the effective rate of strain $\partial w / \partial z$ where w is vertical velocity computed from the rate of upward or downward movement of isotherms. The dissipation rate can be evaluated as $\varepsilon_{str} = 500 \left\langle \left(\partial w / \partial z \right)^2 \right\rangle^2$ where $\langle \rangle$ indicate averaging. However a discrepancy was observed between the coefficient used to compute ε between MacIntyre et al (2009) and Alford and Pinkel (2000), which indicates more efforts are required to understand the role of internal wave dynamics in turbulence production. For the turbulence within the benthic boundary, Ivey et al. (1995) modelled the dissipation ε in the benthic boundary layer by assuming that all the

TKE is absorbed at the boundary: $\varepsilon = r \left(\frac{3\alpha^2 N}{4\pi} \sin 4\beta_0 \right)$ where r is the reflection coefficient ranging from 0 to 1, α_0 denotes the amplitude of the velocity of the incident internal wave field m/s, N is buoyancy frequency and β_0 is the bottom slope.

2.2.4 Horizontal convection

Horizontal convections in coastal waters of lakes, reservoir side arms or shallow water bodies with a sloping bottom are driven by imposed horizontal temperature gradients as a result of differential cooling during the night and differential heating during the day. A typical evidence of convectively driven circulation is a warm outflow along the surface and a cool return flow along the bottom (Monismith et al. 1990). A considerable number of studies have investigated the horizontal convection. Farrow and Patterson (1993) and Farrow (2004) provided the asymptotic solution to the convective flow over varying topography in reservoirs and lakes by solving the set of two-dimensional momentum equations, the temperature transport equation and the continuity equation. Mao et al. (2010) and Mao et al. (2009) reported on a scaling analysis for the unsteady natural convection induced by differential heating and differential cooling respectively and provided

- the scale for the pressure difference driving the horizontal convection as based on the balance between the buoyancy term and the pressure term in the vertical momentum equation;
- the scale for the flow velocity driven by the horizontal convection flow as based on the balance between the pressure term and the viscous term in the horizontal momentum equation;
- and the time scale to reach a steady convective flow as based on the balance between conduction and convection in the energy equation.

Research shows the horizontal convection is important to the biological production in the water body. Monismith et al. (2006) pointed out that the horizontal convection enhanced the connection between coral reefs and the ocean and helped alleviate the stress of coral bleaching. Niemann et al. (2004) indicated that the

horizontal convection in the northern Gulf of Aqaba, Jordan transported near-shore phytoplankton and other suspended matter to the adjacent deep ocean. Hughes and Griffiths (2008) listed the following three governing parameters determining the horizontal convection process:

(a) Rayleigh number

$$Ra_{\Delta b} = \frac{\Delta b L^3}{\nu \kappa} \text{ or } Ra_B = \frac{B L^3}{\nu \kappa^2} \quad (2.2.8)$$

(ii) the aspect ratio of the water body

$$A = \frac{H}{L} \quad (2.2.9)$$

(iii) and the Prandtl number

$$Pr = \frac{\nu}{\kappa} \quad (2.2.10)$$

where $\Delta b \left[m / s^2 \right]$ is the imposed specific buoyancy or $B \left[m^3 / s^3 \right]$ is the imposed rate of supply of specific buoyancy per unit box width, ν is the viscosity, κ is the diffusivity, and H and L are the depth and the length of the water body, respectively. The state of horizontal convection can be classified in terms of the range of Prandtl and Rayleigh numbers in which the Rayleigh number is more determinative. The convective flow is steady and laminar at low to moderate Rayleigh number ($O(10^0) \sim O(10^3)$). At moderate to high Rayleigh numbers, the impact of the Prandtl number becomes important. If the Prandtl number exceeds $O(10^0)$, a shallow horizontal intrusion tends to be formed beyond the boundary layer. If the Prandtl number is not large, with increase of Rayleigh number, increasing unsteadiness and eddying appears in the water body above the horizontal boundary layer formed by the convective flow.

2.3 Turbulence

2.3.1 Non- dimensional parameters for turbulence

There are four dimensionless parameters commonly used to describe the state of turbulence in a stratified water body (Ivey et al. 2008): the flux Richardson number R_f , the turbulent Froude number Fr_T , the turbulent Reynolds number Re_T and the turbulent intensity parameter $\frac{\varepsilon}{\nu N^2}$.

The flux Richardson number R_f represents the ratio of the buoyancy flux to the net mechanical energy available to sustain turbulent motions:

$$R_f = \frac{-\frac{g}{\rho_0} \overline{w' \rho'}}{\overline{-u_i w' \frac{\partial u_i}{\partial x_3}}} = \frac{b}{b + \varepsilon} \quad (2.3.1)$$

where b is the buoyancy flux and ε the dissipation rate. The theoretical maximum value of R_f is given in the range of 0.15~0.2 beyond which the turbulence cannot be maintained at steady state, i.e. the shear production of turbulence is not in balance with the buoyancy flux and the dissipation of turbulence (Osborn 1980). If $b < 0$, water column is will mixed by convective motion. Under such circumstance, the flux Richardson number is negative. Microstructure measurements such as that provided by the Self Contained Autonomous MicroProfiler (SCAMP) can provide direct measurements of the flux Richardson number possible by giving the values of ε and b where b is derived from the temperature variance χ_T and the mean vertical

temperature gradient $\frac{\partial T}{\partial z}$ as $b = -\alpha g \langle w' T' \rangle = \alpha g \frac{\frac{1}{2} \chi_T}{\frac{\partial T}{\partial z}}$ (Dunckley et al. 2012).

Alternatively, Imberger and Ivey (1991) modelled the flux Richardson number in terms of the turbulent Froude number and turbulent Reynolds number:

$$R_f = \frac{1}{1 + 3Fr_T^2}, \quad Fr_T > 1.2 \quad (2.3.2)$$

$$R_f = a + bFr_T + c(Fr_T)^2, \quad Fr_T < 1.2 \quad (2.3.3)$$

where

$$a = 0.49 + 1.44c, \quad b = -0.25 - 2.4c, \quad c = \frac{0.25\alpha - 0.49}{\alpha^2 - 2.4\alpha + 1.44}, \quad \alpha = 3.9 \left[\frac{1}{Re_T} \right]^{1/2}$$

Following the above model, the R_f value is capped around 0.25 when $Fr_T \approx 1.2$. Here the turbulent Froude number Fr_T describes the ratio of the turbulent inertia motions to gravitational forces:

$$Fr_T = \frac{u}{Nl} \quad (2.3.4)$$

where $N \equiv \sqrt{-\frac{g}{\rho_0} \frac{d\rho}{dz}}$ is the buoyancy frequency of the background stratification [s^{-1}], u the scale of the fluctuating velocity [m/s], l the scale of the most energetic overturns [m]. The fluctuating velocity u is given by (Luketina and Imberger 1989)

$$u \sim (\varepsilon L_C)^{1/3} \quad (2.3.5)$$

Thus Fr_T can be expressed as below by using L_C as the scale of the overturns (Ivey and Imberger 1991):

$$Fr_T = \left(\frac{\varepsilon}{N^3 L_C^2} \right)^{1/3} \quad (2.3.6)$$

where N and ε are the buoyancy frequency and dissipation rate averaged over the patches, L_C is the central length scale of an overturn [m] which is calculated from the Thorpe scales by shifting them vertically by one half of their values thereby placing them at the centre of the overturn event.

The turbulent Reynolds number Re_T is defined as the ratio of the inertia of turbulent motions to the viscous forces.

$$Re_T = \frac{ul}{\nu} \quad (2.3.7)$$

where ν denotes viscosity in m^2/s . Since u can be re-written as $(\varepsilon L_c)^{1/3}$ (equation (2.3.5)):

$$\text{Re}_T = \left(\frac{L_c}{L_K} \right)^{4/3} \quad (2.3.8)$$

where L_c is the central length scale of an overturn, L_K^{-1} is the Batchelor's wavenumber ($L_K^{-1} \approx \sqrt[4]{\varepsilon/\nu^3}$) at which the inertial sub-range was cut-off from the dissipation sub-range in the TKE spectrum. In addition to modelling the flux Richardson number, Imberger and Ivey (1991) demonstrated that the ranges of the two dimensionless numbers, Fr_T and Re_T vary greatly with time within a lake and further proposed a Fr_T - Re_T diagram for delineating the mixing mechanisms in the lake as shown below in Figure 2.1.

Figure 2.1 Fr_T vs Re_T diagram showing the various flow regimes reproduced from Imberger and Ivey (1991). The processes are diurnal thermocline (SN), wind driven surface layer (WM), intrusion (IF), shear at the base of the surface layer (SS), penetrative convection (PC), thermal in a homogeneous fluid (PT), hypolimnetic mixing (HM) and boundary mixing (BM).

Besides R_f , Fr_T and Re_T , the turbulent intensity parameter $\frac{\varepsilon}{\nu N^2}$ is another non-dimensional parameter that describe the turbulence in the stratified water. It evaluates the relative importance of the destabilizing effects of turbulent stirring to the stabilizing effects resulting from the combined action of buoyancy and viscosity

(Ivey et al. 2008). Shih et al. (2005) defined ranges of $\frac{\varepsilon}{\nu N^2}$ for different turbulent regimes as listed in Table 2.1.

Regime	$\varepsilon/(\nu N^2)$ range	Behavior of turbulence
Diffusive	$\varepsilon/(\nu N^2) < 7$	turbulence decaying
Intermediate	$7 < \varepsilon/(\nu N^2) < 100$	turbulence in equilibrium
Energetic	$\varepsilon/(\nu N^2) > 100$	energetic turbulence, turbulence growing

Table 2.1 Turbulent regimes based on the ranges of $\varepsilon/\nu N^2$, reproduced from Shih et al. (2005)

Another important parameter is the gradient Richardson number Ri_g defined as

$$Ri_g = N^2 / S^2 \quad (2.3.9)$$

where N is the buoyancy frequency and S is the velocity shear rate in vertical direction

$$S^2 = (\partial U / \partial z)^2 + (\partial V / \partial z)^2 \quad (2.3.10)$$

It measures the relative importance of the buoyancy to the velocity shear in the flow. The critical value is 0.25 as for $Ri_g > 0.25$ the flow is buoyancy dominated and for $Ri_g < 0.25$ the flow is shear-dominated. Unlike the above-mentioned dimensionless numbers, the gradient Richardson number is a property of the mean flow rather than the property of the turbulence. It will be later used to discuss the vertical mixing in Kranji Reservoir together with the dimensionless number for turbulence properties.

2.3.2 Length scales of the turbulence

Length scales of the turbulence are also required to characterize the turbulence state and especially for stratified fluids. Some important mixing length definitions are reviewed here.

The largest possible overturn scale in the presence of stratification is characterized by Ozmidov scale (L_o) or Corrisson scale (l_c) above which the turbulence is restricted by the buoyancy or shear while the smallest possible scale is represented by Kolmogorov scale (L_K) at which the viscous effects are important. Expressions for the Ozmidov and Corrisson scales and the Kolmogorov scale are given as

$$L_o = \left(\frac{\varepsilon}{N^3} \right)^{\frac{1}{2}} \quad (2.3.11)$$

$$l_c = \left(\frac{\varepsilon}{S^3} \right)^{\frac{1}{2}} \quad (2.3.12)$$

$$L_K = \left(\frac{\nu^3}{\varepsilon} \right)^{\frac{1}{4}} \quad (2.3.13)$$

For the buoyancy-dominated flow i.e. $Ri_g > 0.25$ (see the definition of Ri_g given in (2.3.9)), the largest eddy size is restricted by the buoyancy and estimated by L_o in (2.3.11) while for the shear-dominated flow $Ri_g < 0.25$, the largest eddy size is restricted by shear stress and estimated by l_c in (2.3.12).

Dillon (1982) used the Thorpe scale to characterize the complete overturn described above with the Thorpe scale given as:

$$L_T = \left\langle d'^2 \right\rangle^{\frac{1}{2}} \quad (2.3.14)$$

The Thorpe scale represents the root-mean-square of the displacements over which fluid elements can be moved adiabatically along with an unstable density profile. Dillon (1982) plotted the relations between Ozmidov and Thorpe scales on log-log scale and found $L_o = 0.8L_T$. However, it can be seen from the Fr_T - Re_T diagram in Figure 2.1 that other factors such as buoyancy control the relation between L_o and

L_T as Fr_T can be rewritten as $Fr_T = \left(\frac{L_o}{L_c} \right)^{2/3}$. Furthermore, a complete overturn can

be found by comparing an observed vertical density profile with a hypothetical

profile re-ordered by the density and with an overturn spanning a region in the re-ordered vertical profile fulfilling the two requirements (i.e. patches in Figure 2.2):

- within the overturn (mixing patch), no fluid having density between the patch minimum and maximum density can be found outside the patch along the vertical profile;
- no values of density found within the patch are found outside of it in the same profile;

Figure 2.2 Example of complete overturns and the overturns are demarcated by the dash lines, reproduced from Figure 1 in Galbraith and Kelley (1996)

Due to the intermittence and patchiness of the turbulence, the turbulence parameters (such as Fr_T , Re_T and $\frac{\varepsilon}{\nu N^2}$) provide indicators of the bulk features of a complete overturn and various algorithms have been proposed for characterization based on the Thorpe and Ozmidov scales (Dillon 1982; Galbraith and Kelley 1996; Piera et al. 2002) and further improved by removal of noise signals. Moum (1996a; b) proposed two conditions for patch detection: 1) patches only contain signals significantly different from their respective noise levels; and 2) the depth integral of the Thorpe displacement over the patch $\int d_T(z) dz$ equal to zero. As to fulfilling the first condition, the water body should be strongly stratified. Piera et al. (2002) later

used wavelet denoising and Thorpe displacement analysis to remove the uncertainty due to the instrument precision from the microstructure profiles and so extract the bulk features of the overturns. This can be applied to the temperature microstructure profiles in weakly stratified water bodies.

2.4 Numerical models

Various three-dimensional numerical models have been developed to calculate the hydrodynamics in lakes and oceans such as Princeton Ocean Model (POM), the Stanford unstructured-grid, nonhydrostatic, parallel coastal ocean model (SUNTANS) and the Estuary, Lake, and Coastal Ocean Model (ELCOM). In this work, the ELCOM will be later used for the calculation of the motions in Kranji Reservoir. ELCOM solves the hydrostatic, Boussinesq, Reynolds-averaged Navier-Stokes equations and scalar transport equations by using a semi-implicit scheme. It has been extensively used in modelling a wide range of problems in lakes as reviewed below in Section 2.4.1.

2.4.1 Application examples for ELCOM

The Estuary, Lake, and Coastal Ocean Model (ELCOM) is developed by the Centre for Water Resource (CWR) at the University of Western Australia (UWA). It has demonstrated reasonable longer-term (7-day) simulations of circulations with density structure in Lake Kinneret, Israel (Laval et al. 2003). ELCOM was successfully applied to reproduce the spatial distribution of temperature and the heat budget in Lake Erie, at the border of Canada and U.S.A. (Leon et al. 2005). Laval et al. (2005) reproduced the basic characteristics in the salinity profiles in a large tropical lake, Lake Maracaibo, Venezuela, a tropical salt-stratified under spatially varying wind forcing with ELCOM.

ELCOM had also been shown to be a good short-term model of hydrostatic internal waves of approximately four 24-hour wave periods of in Lake Kinneret (Hodges et al. 2000) but it showed a rapid divergence of the model and field signals for higher frequencies, which was a consequence of the dissipation occurring at numerical grid and temporal scales in the model. Vidal and Casamitjana (2008) elucidated the

structure and period of the vertical modes in response to periodic winds in the stratified Sau reservoir in Spain by using ELCOM. The ELCOM simulated velocity field, vertical isotherm displacement, evaluated internal wave field and as well as the modal response of the vertical V2 wave (as defined in section 2.2.2) were in good agreement with the field measurement in the Sau reservoir. However damped displacements associated with the V3 modal response was observed because of numerical dissipation.

2.4.2 Numerical techniques in ELCOM

Within one model time step of ELCOM, the following calculations are performed including: (Hodges and Dallimore 2006)

- 1) introduction of surface heating/cooling in the surface layer;
- 2) mixing of scalar concentrations and momentum using a mixed-layer model;
- 3) introduction of wind energy as a momentum source in the wind-mixed layer;
- 4) solution of the free-surface evolution and velocity field;
- 5) horizontal diffusion of momentum;
- 6) advection of scalars and
- 7) horizontal diffusion of scalars.

The equations are discretized in a Cartesian z coordinate mesh using an Arakawa-C grid stencil with velocities defined on cell faces and free-surface height and scalar concentrations defined on cell centres. The semi-implicit formulation is applied for the momentum equation in which the time advancement is a first-order backwards-Euler method for free-surface evolution, the second-order spatial discretization is used for baroclinic and barotropic terms, and a spatially third-order quadratic discretization for the advective terms.

A conservative third-order scheme is used for scalar transport to minimize unphysical oscillations, numerical diffusion, and non-monotonicity in the scalar fields that are a feature of lower-order schemes. As mentioned above, the scalar concentrations are defined at the cell centres while the velocities at the cell faces. The ULTIMATE flux limiting filter applied with third QUICKEST interpolation is used in ELCOM to interpolate the cell face values of scalar concentrations.

In contrast to full three-dimensional models, ELCOM uses a mixing layer model to simulate the vertical mixing based on the mixing energy budget. The available mixing energy TKE_A includes energies generated by wind, bottom drag, cooling and velocity shear. The local required energy E_{req} for mixing is the potential energy determined by the density gradient at the given layer. The mixing fraction η is the time step divided by the shorter of the timescales of convective overturns and turbulence mixing. If $TKE_A > \eta E_{req}$ mixing occurs and the excessive TKE ($TKE_A - \eta E_{req}$), if any is stored in the mixed layer and is included in the energy budget for the next layer. If $TKE_A < \eta E_{req}$, no mixing occurs and the TKE is dissipated.

In ELCOM there are two different Courant-Friedrichs-Lewy (CFL) conditions used for a time step constraint:

- 1) the internal wave CFL obtained by explicit discretization of the baroclinic terms in the momentum equations;
- 2) an advective CFL by explicit discretization of scalar transport;

The internal wave CFL is more restrictive. In ELCOM one discretization technique is applied for low CFL, and a different technique for high CFL as listed in Table 2.2, which can be generalized into the concept of a "hybrid" numerical method.

CFL	Numerical scheme
$0 < CFL < 2$	Quadratic semi-lagrangian discretization
$CFL > 2$	Linear semi-lagrangian discretization/ approach in TRIM

Table 2.2 Numerical schemes in ELCOM under different CFL values (Hodges and Dallimore 2006)

2.4.3 CAEDYM

The Computational Aquatic Ecosystem Dynamics Model (CAEDYM) is an aquatic ecological model also developed by CWR, UWA. It is usually coupled with the hydrodynamic model ELCOM for water quality modelling. CAEDYM includes

processes for primary production, nutrient cycles and metal cycling, phytoplankton dynamics and the movement of sediment (Hipsey et al. 2006). It has been used in numerous studies on water quality modelling in lakes (e.g. modelling nutrient cycles in Prospect Reservoir, Australia and in Lake Burragorang, Australia (Romero et al. 2004)), modelling the impact of zooplankton grazing on phytoplankton in the Swan River Estuary, Western Australia (Griffin et al. 2001) and predicting the algal bloom in Lake Pusiano, Italy (Copetti et al. 2006). In this PhD work, the CAEDYM will be used coupled with the ELCOM to evaluate the spatial distribution of the phytoplankton in terms of chlorophyll a concentration in Kranji Reservoir.

2.5 Research questions for this thesis

To extend the understanding of the vertical mixing and the flow patterns as well as their impacts on the algal dynamics from the above reported works to a tropical shallow lake such as Kranji Reservoir, the following three broad research objectives are postulated below:

Question1: vertical mixing in Kranji Reservoir

- What is the depth of the surface mixed layer?
- What are the properties of the turbulence in Kranji Reservoir?
- What affects the vertical mixing in Kranji?
- How does the vertical transport in Kranji behave under different meteorological conditions?

Question2: flow pattern in Kranji Reservoir

- What are the possible thermal structure and the flow structure in Kranji reservoir?
- How do the meteorological parameters affect the thermal structure in Kranji Reservoir?
- How reservoir controlling parameters such as inflow affects the hydrodynamics in Kranji reservoir?

Question3: interaction between the dynamics of Kranji Reservoir and the distribution of phytoplankton biomass in Kranji Reservoir?

- How does the vertical mixing affect the chlorophyll a concentration distribution?
- How does thermal structure and flow structure affects the distribution of phytoplankton, for example, via light and nutrient accessibility to phytoplankton?
- Seasonal variations in hydrodynamics and effects on the algal growth.

Question 1 about the vertical mixing will be addressed in Chapter 4. Section 4.2 analyzes the surface mixed layer (SML) dynamics in terms of the common definitions of the SML, the driving mixing mechanism as well as the properties of the turbulence within the SML in Kranji Reservoir. Section 4.3 discusses the turbulence and vertical mixing in the water column below the SML. The vertical transport under the different meteorological conditions is investigated in Section 4.3 via calculation of the vertical heat flux and one-dimensional calculation of the vertical heat transport.

Question 2 is addressed in Chapter 6. Section 6.3 provides a detailed description of thermal and flow structure in Kranji Reservoir based on the field measurement during July 2012. The response of the thermal and flow structure to the different external forcing such as inflow and wind are also studied in Section 6.3. The basin-scale flow pattern during the Southwest Monsoon in Kranji Reservoir is next presented in Section 6.4 based on the numerical results of ELCOM.

Question 3 about the interaction between the hydrodynamics and the algal dynamics is addressed in Chapter 7. The impact of the vertical mixing on the phytoplankton vertical distribution is discussed in Section 7.1. The spatial distributions of chlorophyll a concentration during the Southwest Monsoon is linked to the circulation flow patterns in Section 7.2. Furthermore, in Section 7.3, a scenario test under northerly winds typical of Northeast Monsoon is described to show how the seasonal variations in hydrodynamics or meteorological conditions affect the temporal and spatial variation of the phytoplankton in Kranji Reservoir.

3. Field Experiments

3.1 Objectives of the field experiment

Kranji Reservoir, the site of the field experiment, is located in the northwest corner of Singapore at long/lat ($1^{\circ}25'N$, $103^{\circ}43'E$). It has a catchment area of about 5600 ha and a surface reservoir area of about $2.6 \times 10^6 \text{ m}^2$. Its average depth is about 6.7m. The reservoir is mainly supplied by the runoff brought by four tributaries, namely Sg. KangKar, Sg. Tengah, Sg. Peng Siang, and Sg. Peng Sua, and the direct rainfall on the reservoir. Figure 3.1 shows a map of Kranji Reservoir and its catchments.

Figure 3.1 Kranji Reservoir and its catchments and the locations of field measurement locations including K1 and K2. Surface temperature measurements were taken at M1, M2, M3 and M8. A meteorological station, a temperature thermistor chain and an bottom mounted ADCP were deployed at K2. A bottom mounted ADCP was deployed at K1.

A combination of short term, intensively sampled data combined with longer term monitoring was performed. The short term effort is via two field exercises during 19/09/11 to 08/11/11 and 10/07/12 to 27/07/12 respectively. The following short-term, intensively sampled data were collected during the two field experiments and comprised of:

- time series of meteorological variables including air temperature, relative humidity, long wave radiation, short wave radiation, wind speed, wind direction and vapour pressure;
- water temperature at the main body and at tributaries;
- velocities at the main body (K2) and near the junction of the tributaries (K1);
- transect (cross reservoir channel) data obtained by microstructure SCAMP (Self Contained Autonomous MicroProfiler) instrument and CTD (conductivity-temperature-depth) probe;
- 24hr microstructure measurement of temperature, turbulence and chlorophyll a concentration;
- selected water quality parameters tested externally at about twice weekly

The long term data was via assembling PUB rainfall and water quality data, PUB (via SysEng) on-line surface meteorological and water quality data at Kranji 1 (K1) and Kranji 2 (K2) platforms, and NTU high resolution thermister (temperature) data at K2 and in the 4 tributaries. K2 was selected as it is near the deepest point of the reservoir and K1 was selected as it is close to the junction of the three tributaries.

3.2 Short term field exercise: activities and instrumentation

Two field exercises during 19/09 to 08/11, 2011 and 09/07 to 27/07, 2012 were conducted. The Sept 2011 field exercise included the following activities:

- (a) Attachment of a net pyranometer onto the SysEng's platform⁴ at K2 for on-line measurement of the net solar radiation
- (b) Deployment of a thermister chain with an attached depth sensor at K2 and single thermisters at each of the 4 Kranji tributaries
- (c) Deployment of bottom ADCP (Acoustic Doppler Current Profiler) for velocity profiling at K2 on 20/09 and its recovery on 08/11
- (d) Deployment of a surface ADCP at K1 on 03/10 and its recovery on 12/10
- (e) Three 24-hr SCAMP (Self Contained Autonomous MicroProfiler) profiling exercises on 23/09-24/09, 29/09-30/09 and 05/10-06/10 at K2 and two 6-7

⁴ The net radiation data provided by the SysEng platform was not accurate enough for modelling purpose. A NTU pyranometer was installed to obtain net radiation at sufficient accuracy.

hour daytime SCAMP profiling at K2 (27/09) and (04/10), along with fluorometer measurements.

- (f) Various ADCP transects along with CTD/DO/fluorometer dips across the width of the reservoir main channel intersecting K1 and K2 locations on 26/09 and 03/10.
- (g) Six (6) water sampling events, collecting water from four different depths at K2 at noon and 7 pm on 23/09, 29/09 and 05/10 respectively with samples sent to an external accredited laboratory (Setsco) for water quality analysis.

It is noted that item (c) represented a longer duration of about 7 weeks from the originally planned due to the ADCP being available longer. Items (a) and (b) were continued, and form part of the longer term data gathering effort.

An early analysis of Sept 2011 data showed the SysEng deployed ultrasonic anemometer in the K2 platform consistently overestimated the wind speed and the meteorological data was of insufficient accuracy for supporting detailed quantitative analysis. Thus another independent set of meteorological data was obtained during the second field exercise in Jul 2012 to calibrate with the wind measurements from the first field exercise. SysEng further installed a NTU provided R.M Young Model 05305L anemometer for additional long term wind speed measurements from July 2012 to Dec 2012.

The second field exercise over 09/07 to 27/07, 2012 had the following activities:

- (a) Replacement of the SysEng ultrasonic anemometer (an Airmax LB 150) with a newly acquired R.M Young anemometer Model 05305L at the K2 weather station.
- (b) Deployment of a new weather station (including a wind sensor, rain gauge, radiometer CNR1, air temperature sensor and relative humidity sensor) at K2 platform on 09/07 and its recovery on 27/07.
- (c) Deployment of two ADCPs (velocity profiling) at the bottom of K2 and K1 respectively on 09/07 and its recovery on 27/07.

- (d) Three 24-hr SCAMP profiling exercises on 11/07-12/07, 18/07-19/07, 25/07-26/07 at K2 and two 6-7-hr daytime SCAMP profiling at K2 on 10/07 and 16/07, along with CTD (conductivity-temperature-depth) measurements.
- (e) Various ADCP transects along with CTD/DO/fluorometer dips across the width of the reservoir main channel intersecting K2 locations on 17/07.
- (f) ADCP surface profiling and SCAMP profiling at a point of the main channel (water depth about 5.1 m) for about 5 hours on 23/07.
- (g) Six (6) sampling events, collecting water from four different depths at K2 on 10/07, 12/07, 16/07, 19/07, 23/07 and 26/07 respectively with samples sent to an external accredited laboratory (Setsco) for water quality analysis.

The following field instruments from NTU and Stanford were used in the short term field deployment exercise. Collectively they provide highly accurate measurements of temperature, velocity, turbulent microstructure and water level change. These instruments comprise:

- a) Seabird SBE-56 thermistors
- b) Teledyne RDI Acoustic Doppler Current Profiler (ADCP)
- c) Self-Contained Autonomous Microprofiler (SCAMP)
- d) Seabird SBE-19 Conductivity-Temperature-Depth recorder (CTD)
- e) Turner Self-Contained Underwater Fluorescence Apparatus (SCUFA)
- f) RBR depth sensors DR-1050P
- g) R.M. Young anemometer Model 05305L
- h) Kipp & Zonen CNR1 Net Radiometer

and the existing met sensor AirMar LB150 on the PUB K2 platform. Figure 3.2 shows pictures of the various instruments.

a. SBE 56

c. SCAMP

d. CTD & e. SCUFA

b. ADCP

f. RBR DR-1050P (pict from RBR website)

g. R.M. Young anemometer Model 05305L

h. Kipp & Zonen CNR1

i. AirMar LB 150 (pict from AirMar website)

Figure 3.2 Various instruments used in the field work

Brief descriptions of the instruments are given below as:

Seabird SBE-56 thermistors: These provided temperature data with accuracy of 0.002 °C and a 30 s sampling was used.

Teledyne RDI ADCPs: These provided profile measurements of flow velocity and direction. The ADCPs had a frequency range of 250 kHz to 3000 kHz and 1200 kHz was used in the fieldwork.

Self contained autonomous MicroProfiler (SCAMP): A SCAMP instrument was used to vertically profile the microstructure temperature, fluorescence and turbulence in the reservoir. It samples at a rate of 100 Hz while ascending (as used here) or descending through the water column at a rate of 10 cm/s. The fluorescence signals were calibrated to the chlorophyll-a measurements of the water quality sampling data.

Seabird SBE-19 CTD: This provided measurements of water temperature, conductivity and pressure. The CTD was used to cross validate the SCAMP data and to ensure the quality of the datasets. The CTD was programmed for profiling at 1 Hz while being manually lowered/raised through the water column.

Self-Contained Underwater Fluorescence Apparatus (SCUFA): A SCUFA submersible fluorometer from Turner Designs was further attached to the CTD for fluorescence measurements. The sampling rate of SCUFA was also set at 1 Hz and the fluorescence signals were calibrated to the chlorophyll-a measurements from the water quality sampling events.

Depth Sensors: Two pressure sensors RBR DR-1050P were used for water depth measurement. One was deployed at K2 to measure the water level fluctuation and the other was kept at NTU to track variations in barometric pressure. The sampling interval of pressure sensor was 30 seconds and the accuracy is 0.05% of full scale.

R.M. Young 05305L Wind Sensor: The measurement range of the wind sensor 05305L was from 0~50 m/s with an accuracy of ± 0.2 m/s or 1% of reading. The threshold value of the wind sensor is 0.4 m/s. The sampling interval of the anemometer was set at 30 seconds. The wind direction however was measured by

the wind sensor AirMar LB150 of the SysEng K2 platform. This was because the R.M. Young wind sensor as mounted on the floating K2 platform rotated with the platform, while the AirMar LB 150 had an internal compass to compensate for platform rotation in the wind direction measurements.

Kipp & Zonen CNR 1 Net Radiometer: This radiometer allowed for the measurement of the net (total) radiation at the water surface. The most accurate measurement mode of independently measuring four radiation components was used. These components comprised (i) solar/short-wave radiation by two pyranometers, one for measuring incoming radiation from the sky (CM3Up), and the other which faces downward for measuring the reflected solar radiation (CM3Dn) and (ii) long wave radiation measured by two pyrgeometers, one for measuring the long-wave radiation from the sky (CG3Up), the other from the surface (CG3Dn).

The measurement parameters and the corresponding instruments are summarized in Table 3.1 and Table 3.2

Parameter	Instrument	Measurement Range	Accuracy
Temperature	SCAMP	0 to 30°C	±0.050°C
	SBE 56	-5 to 35°C	±0.002°C
	SBE 19 CTD	-5 to 35°C	±0.005°C
Conductivity	SCAMP	5x10 ⁻² to 9.0 S/m	±0.2% full scale (9 S/m range)
	CTD	0 to 9.0 S/m	±0.0005 S/m
Fluorescence	SCAMP	N.A.	Accuracy depends on the calibration curve. Here it is calibrated with values from the water quality sampling using the same field samples and dilutions of these samples
	CTD SCUFA	0.02 ~10 ⁴ µg/l	
Depth	RBR DR 1050P	0-20 meters	±0.05% full scale
	CTD	0~20 meters	±0.1% full scale
Light intensity	LiCor LI192	400-700 nm	Temperature dependence: ±0.15% per °C maximum; Azimuth: <±1% error over 360° at 45° elevation;
Velocity	Teledyne RDI ADCP	±10m/s	±1% of measured velocity, ±0.5 cm/s

Table 3.1 Measurement parameters and the corresponding instruments

Additional Meteorological Instruments installed on the K2 Platform for the July 2012 Field Exercise			
Parameter	Instrument	Measurement Range	Accuracy
Wind speed	R.M. Young Model 05305L Wind Sensor	0~50 m/s	±0.2 m/s, 1% of the readings; Threshold value for the propeller: 0.4 m/s
Net radiation	Kipp & Zonen CNR1	300~2800(short wave)nm; 4500~42000(long wave) nm	± 10% of the daily sums;
Air temperature	Vaisala Inc. HMP45C-L5	-39.2°C~60°C	±0.3°C at ~30°C
Relative humidity	Vaisala Inc. HMP45C-L5	0.8~100% RH	±3.5% RH (90%~100%RH at ~30°C)
Wind direction (PUB K2 Met Station)	AirMar LB 150	0~360°	Compass accuracy: 3° root mean square (RMS); Wind direction accuracy without precipitation: 2° RMS; with precipitation: 5° RMS

Table 3.2 The additional instruments used in 2012 and their measuring parameters

The additional small met station (referred to as the NTU-Stanford Met Station) with the equipment listed above in Table 3.2 is shown together with the PUB K2 weather station in Figure 3.3 during the July 2012 field work. The sampling frequency of this new met station was set as 30 seconds and the logging frequency was five minutes. The wind sensor was mounted 2.4 meters above the water surface to avoid interference from the housing of the K2 platform.

Figure 3.3 New additional weather station (horizontal pole) and the PUB weather station during the Jul 2012 field exercise. (Note: SysEng also attached an additional NTU provided R.M. Young wind sensor since the Jul 2012.)

3.3 Long term field experiment: activities and instrumentation

The long term dataset comprised of the available PUB rainfall and water quality data, PUB (via SysEng) on-line water quality data at Kranji 1 (K1) and Kranji 2 (K2) platforms, and temperature data at K2 and in the 4 tributaries from the NTU installed SBE 56 thermisters. The deployment of the thermisters was over Aug 2011 to Dec 2012.

PUB via SysEng had earlier converted two former Kranji platforms from an earlier NTU Kranji project into profiling platforms. The two converted platforms, termed

Kranji 1 (K1) and Kranji 2 (K2) stations had locations as marked in Figure 3.1. Measured water parameter profiles included dissolved oxygen (DO), turbidity, chlorophyll-a (APHA: Pt 10200H), NH_4 (APHA 4500- $\text{NH}_3(\text{H})$), NO_3 (APHA 4500- $\text{NO}_3(\text{I})$), temperature and conductivity. The analyses were conducted by an external accredited laboratory (Setsco). K1 only collected water quality data while K2 collected both water quality data and surface meteorological data that included wind speed, wind direction, net radiation, relative humidity, air temperature and total incoming short wave radiation. The met data sampling rate was about 1 minute and the logging interval was five minutes, i.e. the data was averaged over 5 minutes, which was too sparse for the surface heating analysis. All measured water data were sent remotely via GPRS/GSM technology to a shore station, also installed by SysEng.

As mentioned, the wind speed measurements from K2 platform measured by the AirMar LB150 during the Sept 2011 field exercise was found to be inaccurate with consistently high wind speeds recorded. SysEng subsequently installed a NTU provided R.M Young Model 05305L anemometer for additional wind speed measurements over July 2012 to Dec 2012. The R.M Young anemometer was also used during the July 2012 field exercise and this data was used by SysEng to recalibrate the earlier AirMar LB150 wind speed measurements. The new wind sensor was also mounted at about 3.3 meters above the water surface, i.e. mounted significantly higher than the AirMar LB150 that was mounted directly on the K2 housing and thus potentially could be affected by air flow obstructions presented by the housing. An example of the AirMar LB150 data is plotted in Figure 3.4 below where clearly incorrect high values above 8 m/s and up to 10 m/s are seen⁵.

⁵ The typical wind speed for the Singapore/Kranji region ranges from 0-6 m/s with an average of ~1.27m/s (Xing 2011). The AirMar LB 150 high wind speed shown in Figure 3.4 was also seen on other days indicating its inaccuracy.

Figure 3.4 Earlier (AirMar LB 150) SysEng wind speed in (a) and corresponding Young wind speed in (b)

The water inflows from the tributaries into Kranji were calculated from an existing XP-SWMM model (Xing 2011) using as input PUB rain gauge data. The rain gauge data was available for five locations across Kranji's catchment areas as shown in Figure 3.5.

Figure 3.5 Rain gauges marked with rectangular at five locations across Kranji catchment areas

Kranji Reservoir is known to exhibit 3-dimensional hydrodynamic behaviour with horizontal variations in temperature and flow as driven by inflows and thus the inflow temperature could not be neglected (Xing et al. 2014). Thus for a better understanding of the thermal structure in the reservoir, measurements of vertical temperature structure as well as the horizontal temperature distribution were needed. The locations of K2 in the main channel of the reservoir and the four tributaries (Sg. KangKar, Sg. Tengah, Sg. Peng Siang, and Sg. Peng Sua) were selected as the deployment sites for long term temperature measurements.

The deployment of the thermistor chain at K2 is illustrated in Figure 3.6 and described in Table 3.3. The temperature chain at K2 comprised one longer temperature chain weighted down from a bottom anchor and one shorter temperature chain fixed with respect to the surface via a surface buoy. A steel chain was used to loosely connect the two temperature chains. This configuration ensured that the first thermistor of the shorter chain always measured the water temperature at -0.5 m below the surface regardless of the fluctuations in the water level. The vertical resolution of the first four thermistors was set to be smaller at 0.25 m depth spacing, versus the 1m depth interval used in an earlier 2007 deployment (Lo 2008). This is in order to better capture the change across the thin surface mixed layer. In the lower water column the depth resolution increased with depth from 0.5 m to 1.5 m. As for the tributaries, one thermistor was deployed at -0.5 m below the surface at each tributary. The sampling interval for the thermistor chains and thermistors was set at 30 s.

Moorings	Depth (m)	Deployed instruments	Long-term data collection	Duration
K2	~10.5	<ul style="list-style-type: none">• 14 thermistor loggers,• 1 RBR depth sensor;• PUB profiling platform	<ul style="list-style-type: none">• Water temperature;• Water level;• Surface meteorological data• Water quality data	Sept 2011 to Dec 2012
K1	~6	PUB profiling platform	<ul style="list-style-type: none">• Water quality data	
M1 ⁶	1.7	1 thermistor	<ul style="list-style-type: none">• Water temperature	
M2	1.7	1 thermistor		
M3	1.1	1 thermistor		
M8	2.3	1 thermistor		

Table 3.3 Setup of thermistors for the long-term data collection

⁶ locations of M1, M2, M3 and M8 are indicated in Figure 3.1.

Figure 3.6 Deployment schematic of the thermistor chain near K2

3.4 ADCP flow field measurements during the short term field exercises

Three ADCPs were used for the short term field exercise, two on loan from Stanford University and one from NTU. The two Stanford ADCPs were deployed during the Sept 2011 field work: one at the water bottom near K2 looking upwards and the other was used to perform horizontal transects via a boat. The Mode 12

sampling was selected for ADCP mooring at K2 in Sept 2011 with one ping per ensemble and 80 subpings per one ping. The time interval between the ensembles was 12 seconds, which provided an accuracy of velocity measurement of ± 0.78 cm/s.

The NTU ADCP was first deployed at the water surface near K1 facing downwards during the last week of 2011 field work. Only the Mode 1 was available for NTU ADCP, which did not allow for subpings. It was subsequently moored at the bottom at K2 after the field work for an additional deployment period 10/10-08/11 2011. A depth cell size of 0.5m and 20 pings per ensemble were set for NTU ADCP. The ensemble interval was set as 10 seconds giving an accuracy of 2.20 cm/s.

In the July 2012 field work, one Stanford ADCP was moored at the water bottom near K2 and the NTU ADCP was moored at the bottom near K1. The second Stanford ADCP was used for the transects. The Stanford ADCP mooring at K2 was set in the same configuration as in 2011 with the accuracy of 0.78 cm/s. The NTU ADCP deployed at the bottom of K1 took measurements with the depth cell of 0.5m. Each ensemble contained 10 pings and there was no subping capability. The ensemble interval was set as 10 seconds with an accuracy of 2.20 cm/s.

The transects conducted in Jul 2012 were done in a slightly different way from the earlier Sept 2011 transects to allow for a longer averaging time and corresponding improved velocity measurement precision. Eight points were picked up across the width of the reservoir's main channel intersecting K2. The boat stopped at each point and the ADCP was submerged into the water taking a 15-minute measurement. For velocity measurement setup, Mode 12 was selected, the depth cell size was 0.5m, the ensemble contained 1 ping with 6 subpings per ping, and the interval between ensembles was 2 seconds with an accuracy of 2.84 cm/s. Also at these locations, water temperature profiles and chlorophyll a concentration profiles were obtained via both CTD dips and SCAMP profiling.

The measurement activities performed with the ADCPs over both short term field exercises are summarized in Table 3.4 and Table 3.5.

	Equipment	Deployment	Bin size	Duration
Vertical velocity structure	2011 field work			
	Teledyne RDI Workhouse ADCP 1200 KHz (Stanford 1)	Moored at the bottom at K2 (10.5m deep)	0.5 m	3 weeks (20/09- 10/10 2011)
	Teledye RDI Workhouse ADCP 1200 KHz (NTU)	Moored at the water surface at K1		1 week (03/10- 10/10 2011)
		Moored at the bottom at K2		1 month (10/10- 08/11 2011)
	2012 field work			
	Teledyne RDI Workhouse ADCP 1200 KHz(Stanford 1)	Moored at the bottom at K2	0.5m	3 weeks (10/07- 27/07 2012)
	Teledyne DRI Workhouse ADCP 1200 KHz (NTU)	Moored at the water bottom at K1		3 weeks (10/07- 27/07 2012)

Table 3.4 Field activities performed with ADCP to measure the vertical velocity structure during the 2012, 2011 field work

	Equipment	Deployment	Bin size	Sampling information	Duration
Horizontal velocity structure	2012 field work				
	TDRI Workhouse ADCP 1200 KHz (Stanford 2)	15-minute ADCP measurement at 8 selected points across the width of the reservoir main channel intersecting K2; Along with the velocity measurements, SCAMP and CTD profiles were taken at these points.	0.5m	CTD measurements taken at 20 m, 45 m, 112 m, 165m, 217m and 247 m from the east bank (the pump station)	17/07

Table 3.5 Field activities performed with ADCP to measure the horizontal velocity structure during the 2012, 2011 field work

ADCP velocity data processing

Two ADCPs were deployed at K1 and K2 as marked in Figure 3.1. The horizontal velocities U, V were obtained by aligning the velocities with the principal axes. The velocity along the channel U was positive towards the north and negative towards the south, while the cross channel velocity V was positive towards the east and negative towards the west. The 15-minute average was also performed on the velocities.

The ADCPs were deployed at the water bottom during the 2012 field work and faced up to make measurements. As such surface wind or other external surface factors might potentially affect the measurements near the water surface. A quality control process was performed on the ADCP data by setting a cut-off distance near the surface. This was done via an echo data check looking for sudden rises in the

backscattered signal that would likely be caused by external surface factors. The echo intensity as a measure of the intensity of the backscattered echo for each depth cell was measured at K2 and at K1 and depicted in Figure 3.7 and Figure 3.8. There were sudden increases above -1.15 meters at K2 and above -1.1 meters at K1 respectively. Thus the ADCP data above -1.15m at K2 and above -1.1m at K1 were considered as questionable data and were removed from the velocity profiles.

Figure 3.7 ADCP backscatter intensity measured at K2 during 10/07-25/07 2012

Figure 3.8 ADCP backscatter intensity measured at K1 during 10/07-26/07 2012

3.5 SCAMP measurements during the short term field exercises

The 24-hour SCAMP profilings conducted at K2 over the two short time field exercises are summarized in Table 3.6. Measurements were taken automatically within SCAMP as it rose up at a pre-calibrated speed of 10 cm/s and with the sampling frequency set at 100 Hz. The resulting vertical resolution was thus approximately 1 mm. Two SCAMP profiles were obtained every 15 minutes.

During the 2012 measurement, CTD was also used in the 24-hr profiling activities. The CTD measures temperature, conductivity, pressure, fluorescence (chlorophyll-a concentration) and only the upcast data were used since the CTD was lowered rapidly but raised at a slow upcast rate of $O(10\text{cm/s})$. One CTD profile was taken in every 30 minutes.

	Location	Deployment	Duration
Microstructure profiling of turbulence, temperature, fluorescence	2011 Field Work		
	K2	<ul style="list-style-type: none"> • 2 profilings every 15 minutes; • Sampling at 100Hz while ascending at 10cm/s 	<ul style="list-style-type: none"> • Three 24hr measurements on 22/09-23/09; 29/0/-30/09 and 05/10-06/10 • One 6-7 hr measurement on 27/09
	K1		One 6-7hr measurement on 04/10
	2012 Field Work		
	K2	<ul style="list-style-type: none"> • 2 SCAMP profilings every 15 minutes; • One CTD profiling every 30 minutes 	<ul style="list-style-type: none"> • Three 24hr measurements on 11/07-12/07; 18/07-19/07; 25/07-26/07

Table 3.6 24-hr SCAMP profiling measurements during the 2012, 2011 field work

3.6 Water sampling events during the short term field exercises

Water samples were collected from K2 at a frequency of twice a week at four different depth: 0 (i.e. surface), 2, 4, 6 m below the surface. The following nine parameters were tested by an external accredited laboratory (Setsco): TN, TP, TDN, TDP, TSS, NO₃, NH₄, PO₄ and chlorophyll-a concentration (following Pt 10200H Spectrophotometric Determination).

Surface water samples collected on-site were also used to calibrate the fluorometers in the SCAMP and SCUFA. The surface water samples were further diluted on-site using distilled water into ¼ strength; ½ strength and ¾ strength on site and both SCAMP and SCUFA fluorescence measurements taken. Together with the original water sample (full strength) and pure distilled water (zero strength), there were five points for each calibration curve. The fluorescence readings were later plotted against the laboratory measured chlorophyll-a concentration to yield calibration curves with good linearity as shown in Figure 3.9.

Figure 3.9 Typical calibration curves of both SCAMP fluorometer and SCUFA

4. Vertical Mixing in a Tropical Shallow Reservoir

Vertical mixing is of particular importance to the health of the lakes. The transport of the chemical or nutrients are controlled largely by the mixing processes and thus are linked to the primary production of the lake system. In this chapter, the vertical mixing in different layers i.e. surface mixed layer, thermocline and hypolimnion was investigated based on the three 24hr SCAMP measurements during July 2012 at K2 in Kranji Reservoir and other hydro-meteorological data over that period. In the end, the following questions are addressed:

- What is the depth of the surface mixed layer?
- What are the properties of the turbulence in Kranji Reservoir?
- What affects the vertical mixing in Kranji?
- How does the vertical transport in Kranji behave under different meteorological conditions?

4.1 Data processing

The analysis of the properties of vertical mixing in Kranji Reservoir was based on the three 24-hr SCAMP experiments in July 2012, specifically during 11:00 11/07-11:00 12/07, 11:00 18/07-11:00 19/07 and 11:00 25/07-11:00 26/07. The field experiment during which the SCAMP measurements were taken is described in Section 3.5.

4.1.1 SCAMP data processing

The buoyancy frequency, the dissipation rate of TKE and the dissipation rate of temperature fluctuation are extracted from the SCAMP data. The details in deriving the above mentioned parameters from the SCAMP data are described below.

The buoyancy frequency N is defined as

$$N \equiv \sqrt{-\frac{g}{\rho_0} \frac{d\rho}{dz}} \quad (4.1.1)$$

where ρ_0 is the reference density of water equal to 997 kg/m³. The buoyancy frequency N was straightforwardly obtained from the SCAMP temperature microstructure as the salinity is negligible in Kranji Reservoir, a freshwater reservoir.

Estimation of temperature variance χ_T and the dissipation rate of TKE ε is accomplished by the data processing software for SCAMP (Precision Measurement Engineering). The processing algorithm was proposed by Ruddick et al. (2000) and modified by Steinbuck et al. (2009). It determines the k_B such that the Batchelor spectrum based on k_B plus SCAMP noise best matches the observed power spectrum from each segment in a maximum likelihood sense. In this algorithm, χ_T is computed by integrating the vertical gradient spectra in (4.1.1)

$$\chi_T = 6D_T \int_0^{k_L} S_{obs}(k)dk + 6D_T \left\{ \int_{k_L}^{k_n} [S_{obs}(k) - S_n(k)]dk \right\} + 6D_T \int_{k_n}^{\infty} S_B(k)dk \quad (4.1.2)$$

where S_{obs} is the observed vertical temperature gradient spectrum, S_n is a model noise spectrum, D_T is the molecular diffusivity of heat ($=1.4 \times 10^{-7}$ m²/s), k_L is a low wavenumber limit for the Batchelor portion of temperature gradient spectrum and k_u is the highest resolved wavenumber. The first term and the third term on the right hand side are the integral of the best-fit Batchelor spectrum at unresolved low/high wavenumbers. Fast Fourier transform (FFT) with Hanning windowing is used for computing the spectrum. SCAMP generates a slight amount of electronic noise in the fast temperature channel and in the fast temperature gradient channel. The noise spectrum model approximates the noises as broad-band white noise and determines noise spectrum (S_n) from a power spectrum estimate for electronic circuits usage, and an electronic six-pole low-pass filter with a cut-off frequency of 45 Hz.

The dissipation rate of TKE ε can be expressed in terms of Batchelor wavenumber (Batchelor 1959) as

$$\varepsilon = (2\pi k_B)^4 \nu D_T^2 \quad (4.1.3)$$

where D_T is the thermal diffusivity of water ν the kinematic viscosity of water ($=10^{-6}$ m²/s), and k_B is the Batchelor wavenumber. k_B is estimated by fitting the Batchelor spectrum plus the noise spectrum, $S_B + S_n$ to the observed temperature

gradient spectrum S_{obs} of vertical segments containing 256 points (this covers a vertical range of ~0.2 m) and the TKE dissipation rate ε at that depth is computed via equation (4.1.3).

Under isotropic and stationary conditions, the observed Batchelor spectrum of temperature fluctuations in one dimension can be written as (Gibson and Schwarz 1963)

$$S_B(k) = \left(\frac{q}{2}\right)^{1/2} \frac{\chi_T f(\alpha)}{k_B D_T} \quad (4.1.4)$$

Here k is the wavenumber, k_B is the Batchelor wavenumber that can be obtained by fitting the measured power spectrum to the theoretical spectrum, D_T the molecular thermal diffusivity, ν the molecular kinematic viscosity, q the rate of strain proportionality constant taken to be 3.4 and $\alpha = (2q)^{1/2} k / k_B$. The function $f(\alpha)$ is defined as $f(\alpha) = \alpha \left(e^{-1/2\alpha^2} - \alpha \int_{\alpha}^{\infty} e^{-1/2\beta^2} d\beta \right)$. Iteration is used to calculate χ_T . χ_{T_1} for the first iteration is computed via (4.1.2) based on the measured spectrum. The observed Batchelor spectral $S_{obs}(k)$ is next derived via substituting k_B and χ_{T_1} into equation (4.1.4) and χ_T profile is obtained by an inverse transform of the spectrum. To evaluate the goodness of Batchelor fit, the maximum likelihood estimation (MLE) method is used and the normalized maximum likelihood parameter NC_{11} is calculated as

$$NC_{11} = \frac{1}{N} \sum_{i=1}^N \ln \left\{ \frac{d}{S_B(k_i; k_B, \chi_T) + S_n(k_i)} \chi_d^2 \left[\frac{dS_{obs}(k_i)}{S_B(k_i; k_B, \chi_T) + S_n(k_i)} \right] \right\} \quad (4.1.5)$$

where χ_d^2 is a chi-squared distribution with d degrees of freedom and N is the number of spectral points. Following Dunckley et al. (2012), the regions where vertical temperature gradient was smaller than 10^{-3} were not used as it may result high spike in the calculation of buoyancy b later.. The k_B of the best fit is used to estimate ε using equation (4.1.3).

4.1.2 Patch detection

As reviewed in Section 3.3, because of the intermittence and patchiness of the turbulence, the turbulence parameters, Fr_T , Re_T and $\frac{\varepsilon}{\nu N^2}$, indicate the bulk features of a complete overturn as determined by patches.

The patches were detected based on the centred displacements (L_c) derived by Precision Measurement Engineering (PME) software developed earlier by Mike Head for the SCAMP measurement. The patches are determined by smoothing the histogram-shaped centered displacements by a 4th order Butterworth low pass filter with a cut-off frequency of 1Hz. The bounds of each patch were defined as $L_c < 0.0005\text{m}$ (Pernica et al. 2014). Typical patches detected based on the centered displacements are shown in Figures 4.1 to 4.3 for the profiles taken at 11:06 11/07, 15:27 and 22:35 11/07

Figure 4.1 Patch detection results of 11:06 11/07 a) the raw and denoised density (density -1000 kg/m³) profiles with the blue line denoting the raw density profile denoised by Mike Head's software with a 2nd order Butterworth low pass filter, the cut-off frequency of 40Hz and the red line denoting the sorted monotonic density profile b) Thorpe displacement L_t (blue lines) and the centred displacement L_c (the black lines) with the red line showing the smoothed curve by the 4th order Butterworth filter, the cut-off frequency of 1Hz analysis c) the defined patches

Figure 4.2 Patch detection results of 15:27 11/07 a) the raw and denoised density (density -1000 kg/m³) profiles with the blue line denoting the raw density profile denoised by Mike Head's software with a 2nd order Butterworth low pass filter, the cut-off frequency of 40Hz and the red line denoting the sorted monotonic density profile b) Thorpe displacement L_t (blue lines) and the centred displacement L_c (the black lines) with the red line showing the smoothed curve by the 4th order Butterworth filter, the cut-off frequency of 1Hz analysis c) the defined patches

Figure 4.3 Patch detection results pf 22:35 11/07 a) the raw and denoised density (density -1000 kg/m³) profiles with the blue line denoting the raw density profile denoised by Mike Head's software with a 2nd order Butterworth low pass filter, the cut-off frequency of 40Hz and the red line denoting the sorted monotonic density profile b) Thorpe displacement L_t (blue lines) and the centred displacement L_c (the black lines) with the red line showing the smoothed curve by the 4th order Butterworth filter, the cut-off frequency of 1Hz analysis c) the defined patches

4.1.3 Calculation of turbulence parameters

The following parameters were computed to characterize turbulence properties based on the patch-averaged values which are the rms values over the vertical extent of one overturn identified in Section 4.1.2: turbulence intensity parameter

$\frac{\varepsilon}{\nu N^2}$, turbulent Froude number Fr_T , turbulent Reynolds number, mixing efficiency

Γ , flux Richardson number R_f and gradient Richardson number Ri_g . The physical meanings of the parameters have been reviewed in Section 2.3. The buoyancy frequency N , the dissipation rate of TKE ε discussed later in Figures 4.7 to 4.12 are directly derived from SCAMP without patch averaging.

The Fr_T is calculated as (Ivey and Imberger 1991):

$$Fr_T = \left(\frac{\varepsilon}{N^3 L_c^2} \right)^{1/3} \quad (4.1.6)$$

where N and ε are the buoyancy frequency and the dissipation rate averaged over the patches and L_c is the central length scale of an overturn.

The turbulent Reynolds number is calculated as

$$Re_T = \left(\frac{L_c}{L_K} \right)^{4/3} \quad (4.1.7)$$

where L_c is the central length scale of an overturn, L_K^{-1} is the Batchelor's wavenumber.

The last term is the mixing efficiency Γ defined as

$$\Gamma = \frac{R_f}{1 - R_f} = \frac{b}{\varepsilon} \quad (4.1.8)$$

where R_f is the flux Richardson number given by

$$R_f = \frac{b}{b + \varepsilon} \quad (4.1.9)$$

with b being the vertical buoyancy flux defined below and ε the dissipation rate of TKE. The flux Richardson number is calculated in two different ways: direct measurement via equation (4.1.9) and the model proposed by Ivey and Imberger (1991) as reviewed in Section 2.3. Both R_f values are used in the later analysis and comparisons.

If the production of potential energy is assumed to be in balance with dissipation rate of temperature fluctuation and the turbulence is steady and isotropic, the vertical buoyancy flux b can be calculated via $-\overline{w'T'}$ as

$$b = -\alpha g \langle w'T' \rangle = \alpha g \frac{\frac{1}{2} \chi_T}{\frac{\partial \bar{T}}{\partial z}} \quad (4.1.10)$$

where α is the thermal expansion coefficient of water ($=2.57 \times 10^{-4} \text{ }^\circ\text{C}^{-1}$). The χ_T used in 4.1.10 is the rms value over one patch. The background temperature gradient $\frac{\partial \bar{T}}{\partial z}$ need to be calculated carefully as it changes with the selection of depth range. Here the background temperature gradient over a patch in the monotonic profile (e.g. between the red lines in Figure 4.2c) was used to obtain a more representative temperature gradient. With this R_f and Γ were straightforwardly computed following the definitions given in equations (4.1.8) and (4.1.9).

The gradient Richardson number Ri_g at the given depth is defined as

$$Ri_g = N^2 / S^2 \quad (4.1.11)$$

where N is the buoyancy frequency and S is the velocity shear rate in vertical direction

$$S^2 = (\partial U / \partial z)^2 + (\partial V / \partial z)^2 \quad (4.1.12)$$

The horizontal velocities U, V are obtained by aligning the ADCP measured velocities with the principal axes and 15 minute averages performed. Due to the high vertical resolution of SCAMP (1mm) and that the ADCP bins were set as 0.5m, spline smoothing was used to obtain a better fit than e.g. linear interpolation in the estimation of the velocity shear at the depth corresponding to the value of N .

Besides the non-dimensional parameters, the strength of vertical transport characterized by the vertical eddy diffusivity K_z is computed from the microstructure measurements under the assumption of isotropic and steady turbulence (Osborn and Cox 1972):

$$K_z = \frac{1}{2} \chi_T \left/ \left(\frac{\partial \bar{T}}{\partial z} \right)^2 \right. \quad (4.1.13)$$

where χ_T denotes the dissipation rate of temperature variance and $\frac{\partial \bar{T}}{\partial z}$ is the temperature gradient over the vertical range of the patch/overtun.

4.1.4 Calculation of the net surface heat fluxes

The net surface heat flux was estimated based on 15-minute averaged meteorological data by using the Matlab Air-Sea Toolbox Version 2 from Pawlowicz et al. (2001). Xing et al. (2012) closed the heat budget for Kranji reservoir in 2007 with the atmospheric stability taken into account. The value of the Charnock constant (=0.018) used followed Xing et al. (2012). The net surface heat flux $Q_{surface}$ is given as:

$$Q_{surface} = Q_{net} + Q_s + Q_l \quad (4.1.14)$$

where Q_{net} is the net radiation [W/m^2], the sum of net longwave radiation and net shortwave radiation measured by the Kipp & Zonen CNR N1 radiometer; Q_l latent heat flux by evaporative water flux [W/m^2], and Q_s is the sensible heat flux [W/m^2] which is the convective-conductive energy flux either from atmosphere to water (positive) or from water to atmosphere due to the temperature gradient (negative). It should be noted here that the positive energy flux indicates the flux coming from atmosphere to water while the negative flux from water to atmosphere. The calculation of the sensible heat flux Q_s follows the standard bulk formula (Hicks 1972):

$$Q_s = \rho_a C_{p_{air}} C_s U_{10} (T_s - T_a) \quad (4.1.15)$$

where ρ_a is the air density of 1.2 kg/m^3 , $C_{p_{air}}$ is the specific heat of air, equal to $1012 \text{ J/(kg}^\circ\text{C)}$, C_s is the bulk transfer coefficient for sensible heat with values from the Air-Sea Tool box, U_{10} is the wind speed at 10 meters above the water surface, T_s is the surface water temperature measured by the temperature logger SBE 56 deployed at -0.5 m below the surface at K2 and T_a is the air temperature at 10 m above the water surface. Equation (4.1.16) below was used to convert the wind speed measured by the met station at 3 m above the surface to U_{10} :

$$U_{10m} = U_{3m} \left(\frac{10}{3} \right)^{0.27} \quad (4.1.16)$$

Similarly, the air temperature measured at 2.2 m above the water surface was converted to T_a by

$$T_{a|10m} = T_{a|2.2m} - 0.0098(10 - 2.2) \quad (4.1.17)$$

The latent heat flux is calculated by (Hicks 1972):

$$Q_L = \rho_A L_E C_L U_{10} (q_s - q_a) \quad (4.1.18)$$

where L_E is the latent heat of evaporation, C_L is the bulk transfer coefficient for the latent heat, q_s is the saturation specific humidity (kg H_2O / kg air) at the measured water surface temperature and q_a is the specific humidity of atmosphere at 10 m above the water surface. The value of C_L follows the calculation in Air-Sea Tool box with the atmospheric stability taken into account.

The specific humidity at the measured water surface temperature T_s can be computed by the following equations:

$$ew = 6.1121 \left(1.0007 + 3.46 \times 10^{-6} Pa \right) \times \exp \left(\frac{17.502 T_s}{240.97 + T_s} \right) \quad (4.1.19)$$

$$q_s = \frac{0.62197 ew}{Pa - 0.378 ew} \quad (4.1.20)$$

where Pa is the pressure equal to 1012 mbar. The specific humidity of the atmosphere at 10 m above the water surface q_a can be computed by using $T_{a|10m}$ in the above equations.

The surface buoyancy flux J_b^0 is defined as (Monismith and MacIntyre 2009)

$$J_b^0 = \frac{\alpha g Q_{surface}}{\rho_0 c_p} \quad (4.1.21)$$

and $Q_{surface}$ is the net surface heat flux in W/m^2 (positive downwards), ρ_0 is the reference density of water equal to 997 kg/m^3 , c_p is the specific heat capacity

($=4179 \text{ J / (kg } ^\circ\text{C)}$) and α is thermal expansion coefficient of water ($= 2.57 \times 10^{-4} \text{ } ^\circ\text{C}^{-1}$).

4.1.5 Calculation of the nighttime surface mixed layer deepening

The TKE budget was used to evaluate the growth of the SML thickness $h(t)$ during the night time. The calculation followed the equations of the turbulent kinetic energy (Fischer 1979) below:

$$(C_T q^2 + \alpha \Delta T g h)(dh / dt) = C_\kappa^f q^3; \quad (4.1.22)$$

The first term on the LHS arises from the deepening forcing (penetrative cooling and wind) of the SML and the second term on the LHS denotes the potential energy contained by the SML. The change of the SML depth $h(t)$ with time was then integrated forward in time (15 minutes time step was used) as

$$dh = \frac{C_\kappa^f q^3}{(C_T q^2 + \alpha \Delta T g h)} dt \quad (4.1.23)$$

In the above ΔT is taken as the temperature difference between the surface and the base of the SML, and the coefficients $C_T = 0.5$, $C_\kappa^f = 0.13$, and $\eta = 1.23$ (in equation (4.1.24) below) are given in Fischer (1979). The term q is the surface energy flux from wind and penetrative cooling given by the expression:

$$q = (u_f^3 + \eta^3 u_*^3)^{1/3} \quad (4.1.24)$$

with the wind-induced shear velocity u_* given as

$$u_* = \sqrt{\frac{C_D \rho_a}{\rho_0}} U_{10} \quad (4.1.25)$$

and the buoyancy flux expressed in terms of velocity u_f

$$u_f = \left(\frac{\alpha g h \tilde{H}}{c_p \rho_0} \right)^{1/3} \quad (4.1.26)$$

where C_D is the dimensionless drag coefficient ($= 1.3 \times 10^{-3}$), ρ_a is the air density ($= 1.2 \text{ kg / m}^3$), ρ_0 is the reference water density ($= 997 \text{ kg / m}^3$), U_{10} is the wind

speed in m/s at 10 m above the surface calculated by equation (4.1.16), α is the thermal expansion coefficient of water ($= 2.57 \times 10^{-4} \text{ } ^\circ\text{C}^{-1}$), $\tilde{H}$ is the surface cooling rate [W/m] and c_p is the special heat capacity of water ($= 4179 \text{ J/(kg} \cdot ^\circ\text{C)}$).

4.2 Surface mixed layer in a tropical shallow reservoir⁷

The surface mixed layer (SML), also referred to the epilimnion such as that existing in Kranji Reservoir, is an essential component of the lake's or reservoir's ecosystem, interacting directly with the external meteorological forcing, and being actively mixed by the wind-induced turbulence and by the surface heat exchanges. The SML can be characterized into three layers (Imberger 1985, Pernica et al. 2014): a mixing upper surface layer, a diurnal thermocline where reduced mixing occurs and which sometimes extends to the bottom and a subsurface layer. The mixing within the SML determines the light intensity experienced by phytoplankton and hence primary production and further influences the primary production of the water system. The mixing behaviour within the SML in the context of a thermal dynamics of a shallow, tropical lake is investigated in this chapter using the field data acquired as described in Chapter 3, and particularly the data from the July 2012 field exercise.

4.2.1 Meteorological conditions during SCAMP Profiling

The three 24-hr SCAMP profiling activities in July 2012 were carried out under the different meteorological conditions shown in Figure 4.4. The meteorological data were measured by the deployed weather stations at K2. There were two small rainfall events of about 1 mm/hr on the reservoir during the afternoon and the night of 11/07 and the flow from the catchments came into the lake at a rate with maximum of $5 \text{ m}^3/\text{s}$ at around 20:00 as estimated via XP-SWMM calculations using Kranji rain gauge data as input (Xing et al. 2012). Along with the afternoon rainfall, the net heat flux (calculations described in Section 4.1.4), wind speed and the air temperature dropped significantly during the afternoon of 11/07 and the net heat

⁷ A part of this Chapter was submitted as an extended abstract to Physical Processes in Natural Waters (PPNW) 2013, Gold Coast, Australia.

flux fluctuated around zero. As a consequence, the temperature difference between the water surface and the water bottom was the smallest on 11/07-12/07 amongst the three 24-hr periods. The other two 24hr periods of 18/07-19/07 and 25/07-26/07 exhibited the similar weather conditions along with strong heating above 400 W/m^2 during 12:00-16:00 and accompanied by a strong wind of about 3 m/s . There was no rainfall and correspondingly no inflow during these two periods. Rainfall and inflow events occurred daily in the three days before 18/07 whereas no rainfall or inflow in the three days before 25/07-26/07. These conditions were reflected in the 2°C higher air temperature and 0.4°C larger starting temperature differences across the water column during 25/07-26/07 than those during 18/07-19/07.

Figure 4.4 Meteorological conditions during the three 24-hr SCAMP profiling activities

4.2.2 Layers in the SML

The SML can be divided into three layers (Imberger 1985; Pernica et al. 2014): a mixing upper surface layer, a diurnal thermocline where reduced mixing occurs and

which sometimes extends to the surface and a subsurface layer. The layers in the SML can be differentiated by two criteria: the maximum temperature gradient that tracks the diurnal thermocline and the temperature decrease $\Delta T = 0.02^\circ\text{C}$ from the surface water temperature that tracks the surface mixing layer. Here a temperature difference $\Delta T = 0.02^\circ\text{C}$ was chosen based on its usage in Pilkington Bay, a shallow (9 m) embayment of the tropical Lake Victoria in East Africa (MacIntyre et al. 2002). Temperature data collected from three 24-hr series of SCAMP microstructure profiling in July 2012 was used to set the ΔT point and to determine the maximum temperature gradient point. Because of the high vertical resolution of the SCAMP measurement, the temperature gradient was simply obtained by using finite difference: the difference between the adjacent vertical temperatures divided by the vertical range.

The layer structure within the SML derived from the above two definitions are shown in Figure 4.5. The diurnal thermocline by maximum $\partial T / \partial z$ is via dotted black lines and the surface mixing layer by ΔT via dotted blue lines. The circled red lines in Figure 4.5 denote the SML depth calculated by the TKE budget. The circled red line tracks the dotted black lines well. In addition, we superimposed the surface mixing layer on the L_t plots in Figure 4.6. The surface mixing layer by ΔT are consistent with the shallowest maximum in the L_t profiles. These results of TKE budgets and L_t plot support the criterion suitable for the definitions of surface mixed layer and surface mixing layer in Kranji reservoir. The SML depth as well as the depths of the layers varies diurnally. A thin daytime epilimnion of ~ 0.5 m to 2 m was observed with the diurnal thermocline extending to the surface and a thicker nighttime epilimnion was observed with the distinct three layers: surface mixing layer, subsurface layer and the diurnal thermocline close to the water bottom on the nights of 11/07-12/07 and 25/07-26/07 (see Figure 4.5). The vertical mixing in the SML in the context of the diurnal variation is discussed in details in the next section 4.2.3.

Figure 4.5 Surface mixed layer in Kranji Reservoir: SML by $\partial T / \partial z$ plotted via dotted black lines; SML by ΔT via dotted blue lines; TKE calculated SML deepening by the circled red line.

Figure 4.6 Thorpe scale L_t (m) during the three 24hr SCAMP measurements with dotted black lines denoting the SML by $\partial T / \partial z$ and thick black lines denoting the SML by $\Delta T = 0.02^\circ C$

4.2.3 Mixing mechanisms within the SML

Two relevant issues relating to mixing mechanisms was next investigated:

- what were the mixing mechanisms in the SML
- how much mixing (mixing efficiency) resulted from these mechanisms?

The discussion of the mixing mechanism is started with the computed SCAMP results. The buoyancy frequency (N), the dissipation rate of TKE (ε), the gradient Richardson number (Ri_g) and the velocity shear (S^2) are plotted in Figures 4.7 to 4.12 with Figures 4.7 to 4.9 exhibiting the daytime measurements (11:00-19:00) and Figures 4.10 to 4.12 exhibiting the rest of the 24-hour measurements. Measured temperature contours with contour level of $\sim 0.1^\circ\text{C}$ and alongshore (i.e. along-channel) horizontal velocity are also plotted in Figures 4.7 to 4.12 panels a & b for a better understanding of the thermal and mean flow structures and their correlation with the vertical mixing features. In the plots, diurnal thermocline by maximum $\partial T/\partial z$ is denoted via dotted black lines and the surface mixing layer by ΔT via dashed black lines. It is noted here the diurnal thermocline fluctuated greatly during the daytimes as shown in Figures 4.7 to 4.9. It could be because the diurnal thermocline was much dependent on the varying depth range between two adjacent temperature measurements as the SCAMP rose up and occasional temperature fluctuations e.g. due to thermals can induce large temperature gradients. In addition, Figures 4.7 to 4.12 panel b (the velocity), panel e ($\log_{10}(Ri_g/0.25)$) and panel f ($\log_{10}(S^2)$) only show the data below -1.15 meters as the data above this depth was discarded due to errors which has been discussed in Section 3.4.

As shown in Figures 4.7 and 4.9 panel b, the flow structure was three-layer during the daytimes with the surface and bottom underflow towards the dam (positive) and the middle flow towards the junction (negative). During the night time, the SML followed the flow structure denoted by the longshore velocity in Figures 4.10-4.12 panel b and the base of the diurnal thermocline was close to where the maximum shear stress (Figures 4.10-4.12 panel f). Especially during the early evening of 11/07 and the night time period of 25/07-26/07, the SML tracked the strong cold underflow layer near the water bottom, because the bottom stratification induced by the underflow was larger than the decaying surface stratification due to the cooling.

The SML was also consistent with the depth where the thin (red) belt of large $Ri_g > 0.25$ occurred during the nighttime measurements in Figures 4.10-4.12 panel e.

Figure 4.7 Distribution of buoyancy frequency, TKE dissipation rate, gradient Richardson number and shear stress in log scales along with water temperature and velocity during the day time period from 11:00-18:00 11/07

Figure 4.8 Distribution of buoyancy frequency, TKE dissipation rate, gradient Richardson number and shear stress in log scales along with water temperature and velocity during 11:00-18:00 18/07

Figure 4.9 Distribution of buoyancy frequency, TKE dissipation rate, gradient Richardson number and shear stress in log scales along with water temperature and velocity during 11:00-18:00 25/07

Figure 4.10 Distribution of buoyancy frequency, TKE dissipation rate, gradient Richardson number and shear stress in log scales along with water temperature and velocity during 19:00 11/07-11:00 12/07

Figure 4.11 Distribution of buoyancy frequency, TKE dissipation rate, gradient Richardson number and shear stress in log scales along with water temperature and velocity during 19:00 18/07 -11:00 19/07

Figure 4.12 Distribution of buoyancy frequency, TKE dissipation rate, gradient Richardson number and shear stress in log scales along with water temperature and velocity during 19:00 25/07 -11:00 26/07

Next, the following dimensionless variables were computed to facilitate the description of the mixing properties: the gradient Richardson number (Ri_g), turbulent intensity parameter ($\frac{\varepsilon}{vN^2}$), turbulent Froude number (Fr_T), turbulent Reynolds number (Re_T) and the mixing efficiency (Γ) evaluated by direct SCAMP measurements and by Ivey and Imberger (1991)'s model. The physical meanings of the dimensionless numbers have been reviewed in Section 2.3 and the process of computing the numbers from the SCAMP data has been described in Section 4.1. The computation results are plotted in Figures 4.13 to 4.18 with Figures 4.13 to 4.15 exhibiting the daytime measurements (11:00-19:00) and Figures 4.16 to 4.18 exhibiting the rest of the 24-hour measurements.

Figure 4.13 Distribution of turbulence parameters in log scale: a. ε/vN^2 , b. Fr_T , c. Re_T ; d. Thorpe scale (m), e. Γ calculated following Imberger & Ivey, f. Γ calculated by direct measurement of R_f during 11:00 -18:00 11/07

Figure 4.14 Distribution of turbulence parameters in log scale: a. ε/vN^2 , b. Fr_T , c. Re_T ; d. Thorpe scale (m), e. Γ calculated following Imberger & Ivey, f. Γ calculated by direct measurement of R_f during 11:00-18:00 18/07

Figure 4.15 Distribution of turbulence parameters in log scale: a. ε/vN^2 , b. Fr_T , c. Re_T ; d. Thorpe scale (m), e. Γ calculated following Imberger & Ivey, f. Γ calculated by direct measurement of R_f during 11:00- 18:00 25/07

Figure 4.16 Distribution of turbulence parameters in log scale: a. ε/vN^2 , b. Fr_T , c. Re_T ; d. Thorpe scale (m), e. Γ calculated following Imberger & Ivey, f. Γ calculated by direct measurement of R_f during 19:00 11/07 -11:00 12/07

Figure 4.17 Distribution of turbulence parameters in log scale: a. ε/vN^2 , b. Fr_T , c. Re_T ; d. Thorpe scale (m), e. Γ calculated following Imberger & Ivey, f. Γ calculated by direct measurement of R_f during 19:00 18/07-11:00 19/07

Figure 4.18 Distribution of turbulence parameters in log scale: a. ε/vN^2 , b. Fr_T , c. Re_T ; d. Thorpe scale (m), e. Γ calculated following Imberger & Ivey, f. Γ calculated by direct measurement of R_f during 19:00 25/07-11:00 26/07

Thin daytime SML as affected by billows from surface winds

The surface layer defined by $\partial T / \partial z$ had a thickness of about 1 m during the daytime stratification. The SML by $\partial T / \partial z$ was generally in good agreement with the yellow belts of the large buoyancy frequency (N of $\sim O(10^{-1.7})$) at the base of the surface in Figures 4.7 to 4.9 panel c. The Ri_g was < 0.25 and $\frac{\varepsilon}{vN^2}$ was > 1000 above the depth of maximum $\partial T / \partial z$, which is indicative of Kelvin Helmholtz billows driven by shear within the thin daytime stratified SML. The values of Fr_T and Re_T within the thin daytime SML were denoted as black triangles for 11/07-12/07, open circles for 18/07-19/07 and open squares for 25/07-26/07 in Figure

4.19. 84% of Fr_T and Re_T values were in the range of 0.8~3 and in the range of 10^4 ~ 10^3 respectively, demarcated by the red box, which was the Fr_T - Re_T regime given in Imberger and Ivey (1991) for the turbulence produced by the shear billows. Similarly 54% of Fr_T and Re_T values and 52% Fr_T and Re_T values during 11:00-18:00 of 18/07 and 25/07 respectively were in the Fr_T - Re_T regime of turbulence produced by shear billows.

The sizes of the shear-induced billows could be estimated by (Imberger 1985)

$$l = 0.3 \Delta U^2 / g_0', \quad (4.2.1)$$

where ΔU denotes the velocity difference across the surface layer and g_0' is the reduced gravity across the interface between the surface layer and the thermocline. The average ΔU was about 0.03 m/s as shown in the velocity colorplots in Figure 4.7 to 4.9 panel b. g' was about 6×10^{-4} m/s² as evaluated by $g' = \alpha \Delta T g$ and assuming ΔT of about 0.2 °C between the surface and the thermocline (estimated from Figure 4.7 to 4.9 panel a). The length scale from the shear-induced turbulence was thus about 0.5 m, comparable to the surface layer of 0.5~1 m thick during strong heating. It was thus concluded that the thin surface layer was very much affected by the billows generated by the combined effects of the wind-induced surface velocity shear and strong stratification.

Figure 4.19 Fr_T vs Re_T plot within the SML with the triangles for Fr_T and Re_T values during 11:00-18:00 11/07; the circles during 11:00-18:00 18/07; the squares during 11:00-18:00 25/07; the box with solid lines demarcates the range of Fr_T and Re_T for shear induced billow turbulence ($0.8 \leq Fr_T \leq 3$ and $10 \leq Re_T \leq 4000$), the box with dash lines demarcates the range of Fr_T and Re_T for wind induced turbulence ($1 \leq Fr_T \leq 10$ and $10^2 \leq Re_T \leq 10^3$) and the box with dotted lines demarcates the range of Fr_T and Re_T for decaying turbulence ($Fr_T \leq 0.5$ and $10 \leq Re_T \leq 10^4$) following Imberger and Ivey (1991)

The nighttime epilimnion affected by the penetrative cooling

The layers within the SML became distinct during the nighttime (19:00-07:00). The SML was significantly deepened and encompassed the whole water column during the nighttimes of 11/07-12/07 and 25/07-26/07. The nighttime diurnal thermocline deepening process was well calculated by the TKE budget analysis as shown in Figure 4.5, indicating that the SML deepening process was one-dimensional. The large energetic overturns occurred in the surface mixing layer with N about $\sim O(10^{-3}) \text{ s}^{-1}$ (Figure 4.10 to 4.12 panel c), $\varepsilon / \nu N^2 \sim O(10^3)$ (Figure 4.16 to 4.18 panel a) and $L_t > 1\text{m}$ (Figure 4.16 to 4.18 panel d). Interestingly, similar with the diurnal thermocline, $Ri_g > 0.25$ in Figure 4.10 to 4.12 panel d was also observed at these

depths defined by $\Delta T = 0.02$ (the surface mixing layer). Two belts of $Ri_g > 0.25$ were driven by two different mechanisms during the night times: one due to the front of the overturns with small buoyancy frequency $N \sim O(10^{-3}) \text{ s}^{-1}$ and small shear stress $S^2 \sim O(10^{-5}) \text{ m}^2/\text{s}^2$ disturbing the weakly stratified subsurface layer with large buoyancy frequency $N > 0.01 \text{ s}^{-1}$ and large shear stress $S^2 > O(10^{-3}) \text{ m}^2/\text{s}^2$ and the other due to the diurnal thermocline.

The Fr_T vs. Re_T within the SML is plotted in Figure 4.20. As shown, there were a significant proportion of Fr_T and Re_T values in the decaying turbulence regime ($Fr_T < 0.5$ and $10 \leq Re_T \leq 10^4$ given by Imberger and Ivey (1991)). If removing the Fr_T and Re_T values corresponding to decaying turbulence, the percentages of the penetrative cooling caused mixing were 47% during 19:00 11/07-07:00 12/07 (open triangles), 54% during 19:00 18/07-07:00 19/07 (open squares) and 52% during 19:00 25/07- 07:00 26/07 (closed triangles).

The TKE budget analysis also showed the dominance of one-dimensional penetrative cooling in the nighttime SML deepening. The associated net surface energy fluxes of the three SCAMP night time (19:00 to 7:00 of the next day) profiling are listed in Table 4.1. During the nighttimes of 18/07-19/07 and 25/07-26/07, more than 90% of the energy in deepening the SML came from the penetrative cooling.

Figure 4.20 Distribution of Fr_T and Re_T of the overturns above the depth where $T - T_{\text{surf}} = 0.02^\circ\text{C}$; open triangles denote Fr_T and Re_T values of overturns during 19:00-22:00 11/07; open squares during 22:00 18/07-01:00 19/07 and closed triangles during 00:00-02:00 26/07. The box with solid lines demarcates the range of Fr_T and Re_T for penetrative cooling ($1 \leq Fr_T \leq 2$ and $10 \leq Re_T \leq 10^4$) and the box with dash lines demarcates the range of Fr_T and Re_T for decaying turbulence ($Fr_T < 0.5$ and $10 \leq Re_T \leq 10^4$) following Imberger and Ivey (1991)

The mean value of ε and its 95% confidence level (in the parentheses) as computed by the bootstrap method with 1000 samples (Efron and Gong 1983) are shown in Table 4.1. The buoyancy was calculated based on the measurements from meteorological data including wind, relative humidity and vapour pressure. These measurements were less variable and more stable than the measurement of ε and introduced much smaller uncertainties into the calculation of the ratio. Thus, only the uncertainty from the ε was taken into consideration. The mean ratio of ε / J_b^0 was in the range of 0.58-0.74, which is typical for nighttime convections with similar values of 0.61-0.72 reported by Shay and Gregg (1986) and 0.64 by Caughey and Palmer (1979) for convective ocean studies and for a convective atmospheric boundary layer, respectively.

Similarly, the median value of Γ and its 95% confidence level (in the parentheses) as computed by the bootstrap method with 1000 samples (Efron and Gong 1983) are shown in Table 4.2. The median value of night time Γ over the period from

19:00 -07:00 over the SML as listed in Table 4.2 was about 0.13-0.17 from direct measurement and 0.09-0.14 from the model calculation of Imberger and Ivey (1991). The corresponding daytime averaged Γ values over the SML was 0.04-0.17 from the direct measurement for the three days and 0.06-0.12 following the model of Imberger and Ivey (1991). The night time mixing from the penetrative convection was much more efficient than those from daytime wind-induced shear during the last two SCAMP days.

	Mean surface energy flux from penetrative cooling u_f^3 ($\times 10^{-8} \text{ m}^3/\text{s}^3$)	Mean surface energy flux from wind u_*^3 ($\times 10^{-8} \text{ m}^3/\text{s}^3$)	Buoyancy flux J_b^0 ($\times 10^{-8} \text{ m}^2/\text{s}^3$)	ε ($\times 10^{-8} \text{ m}^3/\text{s}^3$)	ε / J_b^0	$u_f^3 / (u_f^3 + u_*^3)$
19:00 11/07- 07:00 12/07	5.35	1.79	6.78	4.65 (4.48 4.92)	0.69 (0.66 0.73)	0.75
19:00 18/07- 07:00 19/07	3.11	0.0620	6.18	4.59 (4.13 5.21)	0.74 (0.67 0.84)	0.98
19:00 25/07- 07:00 26/07	5.36	0.267	7.92	4.65 (4.30 5.01)	0.58 (0.54 0.63)	0.95

Table 4.1 Surface energy fluxes during the nighttimes of the three SCAMP measurements

Mixing efficiency Γ	Daytime SML (11:00-18:00)		Nighttime SML (19:00-07:00)	
	direct measurement	Ivey & Imberger (1991)	direct measurement	Ivey & Imberger (1991)
11/07-12/07	0.17 (0.15,0.19)	0.12 (0.10,0.14)	0.17 (0.13,0.18)	0.09 (0.08,0.11)
18/07-19/07	0.08 (0.05 0.11)	0.10 (0.07,0.15)	0.14 (0.13 0.16)	0.14 (0.11 0.15)
25/07-26/07	0.04 (0.02,0.05)	0.06 (0.04,0.08)	0.13 (0.12,0.14)	0.12 (0.11,0.13)

Table 4.2 the mixing efficiency within the SML

4.3 Vertical mixing below the surface mixed layer

The vertical mixing below the surface mixed layer is discussed with regards to the flow structure in this section. There was a positive current near the water bottom during the daytime of 11/07, the nighttime of 18/07-19/07 and the whole SCAMP period of 25/07-26/07 as shown in the panel b of Figure 4.7, Figure 4.11, Figure 4.9 and Figure 4.12. A basin-scale underflow induced by the differential cooling later shown in Chapter 6, created a turbulent boundary layer in which large gradient Richardson numbers $Ri_g > 0.25$ were observed as a typical evidence of the

subsurface flow. The active turbulence denoted by $\frac{\varepsilon}{\nu N^2} > 100$ was also observed in the panel a of Figure 4.13, Figure 4.17, Figure 4.15 and Figure 4.18 under the regime $Ri_g > 0.25$, which is consistent with Imberger's (1987) observation. Such active mixing could be caused by the internal wave breaking at the bottom boundary layer, in which the dissipation rate of TKE is proportional to the buoyancy frequency as described by equation $\varepsilon = r \left(\frac{3\alpha^2 N}{4\pi} \sin 4\beta_0 \right)$ from Ivey et al. (1998) for the idealized case of the dissipation ε in the turbulent benthic boundary layer with all the TKE being absorbed at the boundary.

The maximum of shear stress over the water column was observed above the underflow layer. Correspondingly, there was a layer of $Ri_g < 0.25$ corresponding to

the maximum shear stress layer with energetic turbulence (70% of $\frac{\varepsilon}{\nu N^2} > 100$) during 11:00 11/07 to 01:00 12/07 and during 25/07 -26/07 (whole 24-hr period) as shown in Figure 4.21 and 4.22 for these periods.

Figure 4.21 Turbulent intensity parameter $\varepsilon/\nu N^2$, gradient Richardson number $Ri_g / 0.25$ and shear stress on log scale with water temperature and alongshore velocity in the lower water column during 11:00 11/07- 01:00 12/07. The black boxes denote the depth where the maximum shear stress layer occurred

Figure 4.22 Turbulent intensity parameter $\varepsilon/\nu N^2$, gradient Richardson number $Ri_g / 0.25$ and shear stress on log scale with water temperature and alongshore

velocity in the lower water column during 25/07-26/07. The black lines denote the depth where the maximum shear stress layer occurred

The Fr_T and Re_T values of this layer were mostly in the range of 0.8-3 for Fr_T and $10-4 \times 10^3$ for Re_T , 61% for 11:00 11/07- 01:00 12/07 and 57% for 25/07-26/07 whole 24-hr period as denoted by black circles and black crosses respectively in Figure 4.23. Thus similar to the strong shear at the base of the surface mixed layer, there was a strong shear layer with active turbulence ($\frac{\varepsilon}{\nu N^2} > 100$) at the top of the submerged flow layer. The shear billows entrained the subsurface flow into the upper layer and potentially enhanced the mass exchange between the underflow and the water column.

Figure 4.23 Distribution of turbulent Froude number and turbulent Reynolds number of the maximum shear stress layers above the underflow with the circles denote the values during 11:00 11/07-01:00 12/07 and the crosses for the time period during the whole 24hr SCAMP measurement from 25/07-26/07; the box gives the range of Fr_T and Re_T for the turbulence induced by the shear ($0.8 \leq Fr_T \leq 3$ and $10 \leq Re_T \leq 2 \times 10^3$)

Above the underflow and below the thermocline where the velocity along the channel was negative towards the junction (Figure 4.7 to 4.9 panel b), low ε values was observed for this region as seen in Figure 4.7 to 4.9 panel d during for the daytime period before 13:00 11/07, the daytime period of 14:00-18:00 18/07 and the whole daytime period of 25/07. This is an evidence of the low production of TKE. The value of $Fr_T < 1$ in Figure 4.13 to 4.15 panel b indicated the available potential energy dominated over the production of TKE. Internal waves are thus expected to be the dominant motions in the region where the flow towards the junction between the SML and the undercurrent layer.

4.4 Vertical transport at K2

In the previous sections, the vertical mixing and its mechanism over the water column was discussed using dimensionless turbulence parameters. In this section, the magnitude of vertical heat transport in different parts of the water column and the importance of vertical transport under different physical forcing is investigated based on field data including SCAMP data and the meteorological data collected at K2.

4.4.1 Magnitude of the vertical heat transport at K2

The results for the vertical diffusivity K_z deducted from χ_T via equation (4.1.13) are presented in Figure 4.24 for the 3 days of SCAMP data. The daytime vertical pattern of K_z shows a three- layer structure: larger values in the surface mixed layer and in the lower hypolimnion where the underflow occurred, while smaller values in the layer between the SML and the underflow layer during 11/07 and 25/07. Due to the inflow during the daytime of 18/07 (see Figure 4.4 for the corresponding rainfall), the vertical eddy diffusivities was larger in the lower water column (below the SML denoted by the circled red lines in Figure 4.24) than in the upper water column (above the SML). During the night of 11/07-12/07, the vertical turbulent transport was active throughout the water column when compared to the same period of 18/07-19/07 and 25/07-26/07 as seen in Figure 4.24. Furthermore during the nights of 18/07-19/07 and 25/07-26/07 but before ~03:00, the K_z distribution

was three-layer similar to the daytime structure of 11/07 and 25/07: (i) active turbulent transport within the surface convective patches (denoted by the depth where $T=T_{\text{surf}}-0.02$ °C) and within the underflow regions and (ii) relatively weak turbulent transport in the depth region in-between. After 03:00 of these two nights, the water column was well mixed via the active vertical turbulence shown in Figure 4.24. The nocturnal vertical eddy diffusivities within the SML in Figure 4.24 were larger than the daytime vertical eddy diffusivities while the nocturnal vertical eddy diffusivities within the underflow layer were continuously large over both daytime and night time.

Figure 4.24 Vertical eddy diffusivities in log scales; in the plot dashed lines denote the temperature contour with a contour level of 0.1 °C; circled red lines denote the depth where the maximum vertical temperature gradient occurred and circled black lines denote the depth where the temperature was 0.02 °C lower than the surface temperature.

The vertical transport was further investigated by estimating the vertical heat flux via Fick's law:

$$J_T = -K_z (\rho_0 c_p) \frac{\partial T}{\partial z} \quad (4.4.1)$$

where the temperature was measured by the SCAMP temperature sensor. The results were presented in Figure 4.25 and the depth-averaged vertical heat fluxes

within each layer were summarized in Table 4.5 to Table 4.7, with the layers defined in Table 4.3 and Table 4.4. For the night times, two tables (Table 4.6 and 4.7) are used to cover two durations from 19:00-03:00 and 03:00-07:00 respectively. These two durations were the two stages of SML deepening process in Kranji Reservoir. The first was the time period of the SML deepening by TKE while the second period from 03:00-07:00 for both 12/07 and 26/07, the SML by $\partial T / \partial z$ was deepened to the water bottom and covered the full water column. The temperature gradient during the second period was smaller than the first period from 19:00-03:00 especially on 12/07 and 26/07.

The vertical heat fluxes within the SML were lower than the fluxes in the underflow layer but stronger than that within the middle layer during the daytimes on 11/07 and 25/07. The strongest vertical heat fluxes of $>50 \text{ W} / \text{m}^2$ was observed within the underflow layer, several times larger than the vertical heat fluxes in the other parts of the water column as shown in Table 4.5. This arose both from the large K_z values and from large vertical temperature gradient within the underflow layer.

During the daytimes of 18/07, the strongest vertical heat flux was observed within the SML. During the daytimes of 18/07, the layer below the SML was well mixed affected by the inflow and the vertical temperature within this layer was weakened by the inflow (temperature difference of about $\sim 0.2 \text{ }^\circ\text{C}$ as also shown in Figure 4.5). Thus, although the K_z values were larger below the SML than in the SML during the daytime of 18/07 in Figure 4.24, the larger temperature gradient within the SML made the vertical heat flux still stronger as shown in Table 4.5.

Consistent with the vertical eddy diffusivities, the nighttime vertical turbulent transport in Table 4.7 was more active overall than during the daytime in Table 4.5. The most active nighttime vertical transport still occurred in the underflow layer with K_z of $10^{-3} \text{ m}^2/\text{s}$ and J_T of $\sim O(10^2) \text{ W} / \text{m}^2$, which was evidence of the enhanced mass exchange between the bottom water and the water column. Similar to the K_z plots in Figure 4.24 during the 19:00-03:00 of 18/07-19/07 and 25/07-26/07, vertical heat fluxes within the convective patches were higher than in the other parts of the SML as shown in Table 4.6, which indicated that the convection

spread the heat downwards at J_T of $\sim O(10^2) \text{ W/m}^2$. However after 03:00, the vertical heat fluxes via convection as well as within the whole SML became smaller of $\sim O(10) \text{ W/m}^2$. This was because the water column had been well mixed and less heat can be transported or redistributed via vertical mixing.

Figure 4.25 Vertical fluxes of heat (W/m^2) on log scales; in the plot dashed lines denote the temperature contour with a contour level of 0.1°C ; circled red lines denote the depth where the maximum temperature gradient occurred and circled black lines denote the depth where the temperature was 0.02°C lower than the surface temperature.

Layer definition / period	Surface mixed layer (denoted by the black thick lines in Figure 4.5)	The middle layer between the SML and the underflow layer	The underflow layer
11/07	SML by $\partial T / \partial z$	Region below SML by $\partial T / \partial z$ and above the isotherm of 29.27 °C in Figure 4.5a	Region between the isotherm of 29.27 °C and the water bottom in Figure 4.5a
18/07		Region between the SML by $\partial T / \partial z$ in Figure 4.5b	
25/07		Region below the SML by $\partial T / \partial z$ and above the isotherm of 29.87 °C in Figure 4.5c	Region between the isotherm of 29.87 °C and the water bottom in Figure 4.5c

Table 4.3 Definition of each layer of the water column at K2 during the daytime heating period of 11:00-19:00

Layer definition / period	Surface mixed layer (denoted by the black thick lines in Figure 4.5)	Surface mixing layer (between the surface and the depth at which $T=T_{\text{surf}}-0.02$)	The underflow layer
11/07-12/07; 18/07-19/07 and 25/07-26/07	SML by maximum $\partial T / \partial z$	denoted by the blue thick lines in Figure 4.5	Region below the SML by $\partial T / \partial z$

Table 4.4 Definition of each layer of the water column at K2 during the cooling period 19:00-07:00

	Surface mixed layer	The middle layer	The underflow layer
11/07	32	14	82
18/07	104	25	
25/07	21	8.9	64

Table 4.5 Vertical fluxes of Heat (W / m^2) within each layer of the water column at K2 during the daytime heating period of 11:00-18:00

	Surface mixed layer	Surface mixing layer	the underflow layer
11/07-12/07	42	25	198
18/07-19/07	91	93	68
25/07-26/07	46	212	292

Table 4.6 Vertical fluxes of Heat within each layer of the water column at K2 during the cooling period 19:00-03:00

	Surface mixed layer	Surface mixing layer	Hypolimnion (underflow region)
11/07-12/07	19	16	74
18/07-19/07	19	15	67
25/07-26/07	78	23	437

Table 4.7 Vertical fluxes of Heat within each layer of the water column at K2 during the cooling period 03:00-07:00

4.4.2 Importance of the vertical transport under different meteorological conditions

In the previous section, the computed heat flux was calculated based on the one-dimensional Fick's law. In this section, we would like to investigate under what circumstance the vertical mixing dominates the heat transport in Kranji reservoir. The importance of the vertical transport under different meteorological conditions is evaluated by calculating the temperature vertical distribution arising from the vertical mixing via the one-dimensional heat/mass transport and comparing with the field measurement:

$$\frac{\partial T}{\partial t} = \frac{\partial}{\partial z} \left(-K_z \frac{\partial T}{\partial z} \right) + \frac{\partial}{\partial z} \left(\frac{R(z)}{\rho_0 c_p} \right) \quad (4.4.2)$$

Here the depth varying measured K_z was used, $R(z)$ is the heat source (W/m^2), ρ_0 denotes the reference water density ($=997 \text{ kg/m}^3$) and c_p the specific heat capacity of water ($=4179 \text{ J/(kg}^\circ\text{C)}$). The horizontal convection terms are neglected in the above equation so only temperature changes resulting from the vertical mixing and the heating were evaluated. By comparing with the field temperature measurements, one could then assess the importance of the vertical mixing under different meteorological conditions. Following the Fick's law in equation (4.4.1), the diffusion term can be re-written as

$$\frac{\partial}{\partial z} \left(-K_z \frac{\partial T}{\partial z} \right) = \frac{\partial}{\partial z} \left(\frac{J_T}{\rho_0 c_p} \right) \quad (4.4.3)$$

In the calculation, staggered grids were used (shown in Figure 4.26) in which the scalars such as temperature, chlorophyll a concentration and the vertical eddy diffusivities are defined in the cell center and the vectors such as the velocity and the fluxes are defined in the cell boundaries.

Figure 4.26 Grids for the calculation of the one-dimensional diffusion equation (4.4.3)

By assuming the flux between the neighbouring cells is determined by the average K_z of the two adjacent cells (here SCAMP measured K_z values are used), the vertical flux of the cell j at time step $n+1$ is given as

$$\frac{J_{Tj}^{n+1}}{\rho_0 c_p} = \left(\frac{K_{zj-1}^{n+1} + K_{zj}^{n+1}}{2} \right) \left(\frac{T_j^{n+1} - T_{j-1}^{n+1}}{\Delta z} \right) \quad (4.4.4)$$

$$\frac{J_{Tj+1}^{n+1}}{\rho_0 c_p} = \left(\frac{K_{zj+1}^{n+1} + K_{zj}^{n+1}}{2} \right) \left(\frac{T_{j+1}^{n+1} - T_j^{n+1}}{\Delta z} \right) \quad (4.4.5)$$

and thus the diffusion term is

$$\frac{\partial}{\partial z}(J_T) = \left(\frac{J_{Tj+1}^{n+1} - J_{Tj}^{n+1}}{\Delta z} \right) \quad (4.4.6)$$

For the numerical solution of equation (4.4.2), implicit backward in time and central difference in space were applied and the equation (4.4.2) can be re-arranged as:

$$aT_{j-1}^{n+1} + bT_j^{n+1} + cT_{j+1}^{n+1} = T_j^n + \left(\frac{-k_s}{\rho_0 c_p} \right) R_j^n \Delta t \quad (4.4.7)$$

for $2 \leq j \leq N-1$:

$$a_j = -\frac{\Delta t}{\Delta z^2} \left(\frac{K_{zj}^{n+1} + K_{zj-1}^{n+1}}{2} \right); \quad (4.4.7a)$$

$$b_j = \left[1 + \frac{\Delta t}{\Delta z^2} \left(\frac{K_{zj+1}^{n+1} + K_{zj}^{n+1}}{2} \right) + \frac{\Delta t}{\Delta z^2} \left(\frac{K_{zj}^{n+1} + K_{zj-1}^{n+1}}{2} \right) \right]; \quad (4.4.7b)$$

$$c_j = -\frac{\Delta t}{\Delta z^2} \left(\frac{K_{zj+1}^{n+1} + K_{zj}^{n+1}}{2} \right); \quad (4.4.7c)$$

For the upper boundary (water surface), the heat flux into the system was taken as the net heat flux. The expressions for a, b, c and the RHS (right hand side) of the equation were given as (j=1):

$$a_1 T_0^{n+1} + b_1 T_1^{n+1} + c_1 T_2^{n+1} = T_1^n + \frac{H \Delta t}{\rho_0 c_p \Delta z} + \left(\frac{-k_s}{\rho_0 c_p} \right) R_1^n \Delta t; \quad (4.4.8)$$

$$a_1 = 0; \quad (4.4.8a)$$

$$b_1 = \left[1 + \frac{\Delta t}{\Delta z^2} \left(\frac{K_{z2}^{n+1} + K_{z1}^{n+1}}{2} \right) \right]; \quad (4.4.8b)$$

$$c_1 = -\frac{\Delta t}{\Delta z^2} \left(\frac{K_{z2}^{n+1} + K_{z1}^{n+1}}{2} \right); \quad (4.4.8c)$$

For the lower boundary (sediment), zero heat flux exchange between the sediment and the water column was assumed. The expressions for a, b, c and the RHS (right hand side) of the equation were given as (j=N):

$$a_N T_{N-1}^{n+1} + b_N T_N^{n+1} + c_N T_{N+1}^{n+1} = T_N^n + \left(\frac{-k_s}{\rho_0 c_p} \right) R_N^n \Delta t; \quad (4.4.9)$$

$$a_N = \left[1 + \frac{\Delta t}{\Delta z^2} \left(\frac{K_{zN}^{n+1} + K_{zN-1}^{n+1}}{2} \right) \right]; \quad (4.4.9a)$$

$$b_N = \left[1 + \frac{\Delta t}{\Delta z^2} \left(\frac{K_{zN}^{n+1} + K_{zN-1}^{n+1}}{2} \right) \right]; \quad (4.4.9b)$$

$$c_N = 0; \quad (4.4.9c)$$

The depth profile of temperature at time step $n+1$ was obtained by performing a matrix inversion on the equation (4.4.10) below:

$$\begin{pmatrix} b_1 & c_1 & & \\ a_j & b_j & c_j & \\ & \ddots & \ddots & \ddots \\ & & a_N & b_N \end{pmatrix} \begin{pmatrix} T_1^{n+1} \\ T_j^{n+1} \\ \vdots \\ T_N^{n+1} \end{pmatrix} = \begin{pmatrix} T_1^n + \frac{H\Delta t}{\rho_0 c_p \Delta z} + \frac{-k_s R_1^n \Delta t}{\rho_0 c_p} \\ T_j^n + \frac{-k_s R_j^n \Delta t}{\rho_0 c_p} \\ \vdots \\ T_N^n + \frac{-k_s R_N^n \Delta t}{\rho_0 c_p} \end{pmatrix} \quad (4.4.10)$$

In the calculation the grid size Δz was set as 0.5m and the time step Δt was 15 minutes. The heat source of R_j^n during the heating period was modelled follows (Imerito 2007):

$$R_j^n = (1-0.09)R_{short-wave}^n \cdot (1-0.55) \cdot \exp(-k_s z) \quad (4.4.11)$$

where $R_{short-wave}$ is the measured incident short wave solar radiation in W/m^2 ; z (positive) is the depth in m into the water (equal to $z = j\Delta z$); k_s is the attenuation coefficient, about 1.3 m^{-1} in Kranji Reservoir in 2012. During the cooling period, the heat loss is assumed on the surface and the following equation (4.4.12) is given as when $2 \leq j \leq N$

$$aT_{j-1}^{n+1} + bT_j^{n+1} + cT_{j+1}^{n+1} = T_j^n \quad (4.4.12)$$

when $j=1$

$$a_1 T_0^{n+1} + b_1 T_1^{n+1} + c_1 T_2^{n+1} = T_1^n + \frac{H\Delta t}{\rho_0 c_p \Delta z}; \quad (4.4.13)$$

The results were plotted in Figure 4.27 along with the field measurements. The temperature calculation was in reasonable agreement with the field measurement during the daytimes of 11/07-12/07 and 18/07-19/07, indicating that one dimensional transport was dominating during these time periods. The surface layer thickness calculated with K_z (direct measurement) evolved from -0.5 m to about -3 m on 11/07 and from -0.5 m to about -2 m on 18/07, which was very close to that

observed. The temperature calculation on 25/07 deviated noticeably from the field observation. The water column was more stratified and more heated in the 1-D model. The calculated SML was about 3 m while the observed SML was less than 2 m. This was an evidence of the non-one-dimensionality in the Kranji thermal system. During the nighttime, although slightly affected by the accumulated numerical diffusion, the calculation gave good estimates of the temperature distribution during the nighttime for the first two measurement periods. From the above results, it seemed that the temperature distribution was determined by the vertical mixing and vertical heating during the first two measurement periods. During the last measurement period, the thermal structure was not only dependent on the vertical mixing but also the horizontal dispersion/advection. Thus the computed heat flux here is valid for the first two days as the processes in these days are 1-D. Whether the processes are 1-D or 3-D are further investigated via the physical forcing regime of Xing et al. (2014).

Figure 4.27 Calculation of water temperature using one-dimensional transport equation with the K_z derived from the SCAMP measurements

We first however note that the turbulent diffusivity could be calculated by two methods

$$1) \quad K_z = \frac{1}{2} \chi_T \left/ \left(\frac{\partial \bar{T}}{\partial z} \right)^2 \right. \text{ direct measurement via SCAMP and}$$

$$2) \quad K_z = \Gamma \frac{\varepsilon}{\nu N^2} \quad \text{where} \quad \Gamma = \frac{R_f}{1 - R_f} \quad \text{and the } R_f \text{ follows Ivey and Imberger (1991).}$$

The SCAMP day of 11/07-12/07 was selected to evaluate the sensitivity to the choice of K_z as the dynamics was dimensional. As shown in Figure 4.28, calculation of temperature (panel b) with the turbulent diffusivity directly measured via χ_T is closer to the field observations (panel a). The water column calculated with turbulent diffusivity by Imberger & Ivey (1991) (panel c) is more stratified. Thus, the heat transport using the turbulent diffusivity of Imberger & Ivey (1991) is less efficient than in the field and by using the directly measured $K_{z\chi}$. This could be because the parameterization by Ivey and Imberger (1991) was not able to calculate mixing efficiency in the weakly stratified water column due to a large occurrence of small Fr_T values ($Fr_T < 0.63$) (Dunckley et al. 2012).

Figure 4.28 temperature calculation using different values of K_z

The dynamics in Kranji Reservoir was earlier characterized by Xing et al. (2014) in terms of the physical forcing regimes based on two dimensionless numbers I_{IW} and I_{WB} . I_{IW} evaluates the relative importance of the buoyancy flux from inflow to the wind mixing forcing while I_{WB} that is similar to the Monin-Obukhov length, evaluates the competing effects of the heating and wind mixing. Xing et al. (2014) defined (i) a solar radiation dominated regime with $I_{WB}<1$ and $I_{IW}<0.4$, (ii) a windy regime with $I_{WB}>1$ and $I_{IW}<0.4$ and (iii) a cold inflow regime with $I_{WB}<1$ and $I_{IW}>0.4$. Here the 3 days of SCAMP measurements in 2012 were evaluated based on this classification.

Following Xing et al. (2014),

$$I_{IW} = \frac{BH}{u_*^3 A_s} \quad (4.4.14)$$

where u_* is the wind-induced shear velocity calculated following equation (4.1.25), A_s is the surface area of Kranji Reservoir, about $3 \times 10^6 m^2$ and B denotes the buoyancy flux from the tributary inflow and would be calculated by

$$B = \alpha \Delta T_0 g Q \quad (4.4.15)$$

in which α is the thermal expansion coefficient of water ($=2.57 \times 10^{-4} \text{ } ^\circ\text{C}^{-1}$), ΔT_0 is the measured surface temperature difference between the tributaries and the reservoir ($\Delta T_0 = T_{\text{surface at K2}} - T_{\text{tributary}}$) and Q is the sum of the inflows from the catchments into the reservoir as simulated by XP-SWMM using rainfall data from the catchment. A weighted average as dependent on both the inflow volume and tributary temperatures was further performed to get a better representative tributary temperature:

$$T_{\text{tributary}} = \frac{Q_{M1} \times T_{M1} + Q_{M2} \times T_{M2} + Q_{M3} \times T_{M3} + Q_{M8} \times T_{M8}}{Q} \quad (4.4.16)$$

As no rainfall/inflow data was available at M2 for the field experiment, we substituted it with the inflow at M1. The other dimensionless number I_{WB} is:

$$I_{WB} = -\frac{u_*}{u_f} \quad (4.4.17)$$

and was directly calculated using equations (4.1.25) and (4.1.26). It is noted here the duration of I_{WB} calculation only covers the heating period from 06:00-18:00 when the reservoir is being stratified due to incoming solar radiation while the duration of I_{IW} was over the whole 24hr period from 0:00 to 23:59, as following Xing et al. (2014).

The resulting 15 minute time series of the I_{WB} and I_{IW} are plotted in Figure 4.29. During the first 24hr SCAMP measurement from 11:00 11/07 - 11:00 12/07, the dynamics of the reservoir was under the effect of the wind mixing. During the second 24hr SCAMP measurement from 11:00 18/07-11:00 19/07, the stratification induced by the inflow was important while the wind induced mixing effect was the smallest among the three 24hr periods. During the SCAMP period from 11:00 25/07- 11:00 26/07, the influences coming from the inflow and the wind were small and the solar radiation was the dominant forcing. The physical regimes of these days are listed in Table 4.8. It should be noted here that I_{IW} was averaged over the time period of 0:00-23:59 and I_{WB} was averaged over the heating period of 06:00-18:00 of the corresponding days in Table 4.8. From the above physical forcing regime calculation, the vertical mixing in Kranji Reservoir in the cold inflow regime could be described as one-dimensional system as the cold inflow uniformly mixed the water column in the whole reservoir and made the vertical variations dominant in the dynamics of Kranji reservoir. The reservoir could not be treated as a strictly one-dimensional system on the solar radiation dominated day of 25/07-26/07 as the cold undercurrent that was caused by differential cooling and varied along the channel made the horizontal variation also important; thus the reservoir became three dimensional. The three dimensional ELCOM-CAEDYM model is thus later applied to give a more comprehensive picture to the thermal structure, flow structure as well as chlorophyll a distribution in Kranji Reservoir as is discussed in Chapter 6 and Chapter 7.

Figure 4.29 Fifteen minute time series of I_{IW} and I_{WB} during the three SCAMP days in July 2012

	I_{IW}	I_{WB}	Physical forcing regime
11/07	0.68	0.44	Cold inflow regime
12/07	0.52	0.16	Cold inflow regime
18/07	1.00	0.31	Cold inflow regime
19/07	2.92	0.33	Cold inflow regime
25/07	0.31	0.14	Solar radiation regime
26/07	0.10	0.51	Solar radiation regime

Table 4.8 Time averaged I_{IW} and I_{WB} values during the three 24hr SCAMP periods in July 2012 and their correspondent physical forcing regimes; I_{IW} was averaged over the time period of 0:00-23:59 and I_{WB} was averaged over the heating period of 06:00-18:00 of the corresponding days.

4.5 Summary

The maximum temperature gradient method for the definition of the surface mixed layer (SML) gave an accurate estimation of the surface mixed layer depth, as it

matched the evolution of SML predicted from the TKE budget model as well as the features displayed by the observed turbulence parameters and the mean flow structure. The temperature difference of 0.02°C from the surface temperature was a good indicator of the surface mixing layer (the surface regions mixed intermittently and actively by large overturns) as it was visually consistent with the shallowest maximum in the Thorpe displacement profiles. These results of TKE budgets and the L_t profiles support the criterion suitable for the definitions of surface mixed layer and surface mixing layer in Kranji reservoir.

The driving mixing mechanism and the properties of the turbulence had been examined by TKE budget analysis, the $Fr_T - Re_T$ plot of Imberger and Ivey (1991), the turbulence activity parameter $\frac{\epsilon}{\nu N^2}$ and the mixing efficiency Γ . The $Fr_T - Re_T$ turbulent flow behaviour as delineated by Imberger and Ivey (1991) was found to apply here in that it was consistent with the TKE budget analysis and could explain the observed vertical structures as developing diurnally.

The daytime SML was newly built up and thin (<2m) as the water column in Kranji Reservoir was well mixed during the night time. Along with the strong wind always accompanied with the strong heating, the thin daytime SML layer was very much affected by the wind shear generated billows and exhibited a low mixing efficiency $\Gamma \sim 0.1$. Such billow dominant daytime SML might be expected under similar meteorological conditions (strong heating with strong wind) in tropical shallow lakes where the complete mixing occurs on a daily scale. The nighttime SML was dominated by penetrative cooling with a ratio of ϵ / J_b^0 of ~ 0.6 typical for nighttime convections. The nighttime mixing efficiency (Γ about 0.2 from direct SCAMP measurement) as arising from the penetrative convection was much more efficient than those from daytime wind. Correspondingly, the turbulence also varied diurnally within the SML. The sizes of the overturns were small of ~ 0.1 m during the daytime and increased to ~ 1 m during the night time.

The mixing was also strong and active near the bottom waters throughout the SCAMP observation period of three days. The velocity profiles indicated underflows near the water bottom towards the dam due to the horizontal convection.

The large $Ri_g > 0.25$ values and the increasing ε with the buoyancy frequency N within the bottom turbulent layer supported the turbulence induced by the underflow intrusion (Imberger 1987). The analysis on vertical eddy diffusivities and the vertical transport showed the strongest vertical transport was in the underflow regions, which potentially enhanced the mass (nutrient) exchange between the water column and the sediment. This role of the underflow to the Kranji ecosystem deserves further investigation.

In this work we have observed that the SML encompassed the whole water column on two SCAMP nights (11/07-12/07 and 25/07-26/07). Two thermoclines, i.e. diurnal thermocline and seasonal thermocline are known to exist in temperate shallow lakes (Read and Rose 2013) but only the diurnal thermocline exists in Kranji reservoir. More active mass exchanges over the water column in tropical shallow lakes are expected. Furthermore, the turbulent parameters observed in the SML including the mixing length, the buoyancy frequency and the vertical eddy diffusivities would be useful to develop better estimates of the time scale of phytoplankton transport via mixing in the SML. This should contribute to the future development of more realistic biological models in tropical lakes.

The temperature calculation results based on the one-dimension vertical transport equation coupled with the forcing regimes classified by Xing et al. (2014) showed that the thermal structure in Kranji was one-dimensional as driven by vertical mixing and vertical heating during the first two SCAMP days which under the influence of the cold inflows. However horizontal transport still affected the water column at K2 as indicated by the undercurrent near the bottom. During the last SCAMP measurement period, the one-dimensional calculation gives a more stratified water column in terms of temperature, which indicated that the vertical heat distribution for the strong three-layer structure in the solar radiation dominated regime was not only dependent on the vertical mixing but also the horizontal dispersion/advection. As shown by the classification results of the three SCAMP measurements, alternative forcing become dominant within a short time period (on a weekly scale for the three SCAMP measurements shown here in Kranji Reservoir), probably due to its delicate dynamic system of Kranji Reservoir. It

seems that the thermal structure in tropical shallow lakes could behave differently. For example, the strong differential cooling/heating occurring on the solar radiation dominated days makes the horizontal advection significant and thus the system is three-dimensional. In contrast, inflow flushing into the reservoir mixes the reservoir horizontally and thus makes the system become one-dimensional.

5. Numerical Study of Kranji Reservoir Flow Pattern using ELCOM

A numerical study was carried out using the Estuary, Lake, and Coastal Ocean Model (ELCOM) in order to develop a comprehensive picture of the flow and thermal structure, and as well as the basin scale circulation patterns in Kranji reservoir. The field measurements carried out during July 2012 provided temperature flow data that was used to assess the ELCOM results.

5.1 Setting up ELCOM for Kranji Reservoir

The steps listed below are followed for the setting up and running an ELCOM calculation for Kranji Reservoir following the ELCOM user manual (Hodges 2000):

- 1) prepare bathymetry.dat file that provides topography and grid information;
- 2) prepare bc.dat file that designates sets of cells for applying boundary conditions;
- 3) run the pre-processor to produce sparsedata.unf and usedata.unf files;
- 4) prepare input files including initial conditions files, temporal boundary condition files (*.dat) for data applied to the boundary condition cells, a datablock file (*.db) for designating data output operations;
- 5) configure the ELCOM run elcom file;
- 6) run the executable, directing the screen output to a file;
- 7) post-process data;

The bathymetry.dat file was set up earlier by Xing (2007) based on a survey conducted in the 1970s and complemented by data extracted from satellite images in 2006. The Kranji bathymetry consists of a horizontal orthogonal grid of 20x20 m wide cells and a vertical grid of 0.2 m with a total of 100 layers to represent a water depth of 22 m. The 3-D view of Kranji Reservoir and its bathymetry are shown in Figures 5.1 and Figure 5.2.

Figure 5.1 3-D view of Kranji Reservoir with a horizontal grid of 20x20m and a vertical grid of 0.2m

Figure 5.2 Bathymetry of Kranji Reservoir with the colorbar showing the water depth in meters

The bathymetry of Kranji Reservoir also provided the spatial information on the boundaries at which the inflow and the outflow occurred. The cell coordinates of the boundaries were used to prepare the boundary file, bc.dat. The sparsedata.unf and usedata.unf files were then produced by running the pre-processor. The input files included initial conditions files, temporal boundary condition files (*.dat) for data applied to the boundary condition cells. The initial conditions and the temporal boundary conditions were obtained from the field measurements as follows:

- a) the meteorological data including net radiation, short-wave/long wave radiations, air temperature, relative humidity, wind speed and wind direction were measured from NTU-Stanford met station as described in Section 3.2;
- b) the inflow data was simulated by XP-SWMM based on the rainfall data collected by the rain gauges in the Kranji catchments;
- c) the water temperature was measured by thermistors SBE 56 at K2 and at tributaries;

The configuration of the ELCOM run was as follows:

- a) ELCOM run for 17 days from 10/07/2012 to 27/07/2012 and the time step was set as 120 seconds;
- b) grid size used was 20 m x 20 m (horizontal) x 0.2 m (vertical)
- c) turbulent benthic boundary layer was selected as boundary conditions for the ELCOM run;
- d) the diffuse attenuation coefficient k_d was 1.3 m^{-1} for 2012 as measured by LiCor LI192.
- e) default and adjusted mixing coefficients were used and the values are listed in Table 5.1.

Mixing coefficients	Default values in ELCOM	Adjusted values for ELCOM (Sherman et al. 1978)
Shear coefficient	0.15	0.5
Wind stirring coefficient	1.13	2.2
Dissipation coefficient	1.15	1.15
Convection coefficient	0.2	0.2
Benthic coefficient	0.21	0.21

Table 5.1 Default mixing coefficients in ELCOM (Hodges and Dallimore 2006) and adjusted from Sherman et al. (1978)

The ELCOM output was the .nc file and further processed by Matlab. To evaluate model performance, ELCOM output was first compared with field measurements, focusing on the general qualitative reproduction of observed thermal structure and flow structure in the horizontal (along-channel), transverse (cross-channel) and vertical directions. These field measurements are described earlier in Section 3.2 and Section 3.4.

5.2 ELCOM calculations of thermal structure

ELCOM calculation of the temperature at K2 is plotted in Figure 5.3 along with the temperature measured by the SBE 56 thermistors of the K2 thermister chain for the July 2012 field exercise. This covered a 17-day period from 10/07/2012 to 27/07/2012. ELCOM results had the same temperature trend as the measurement: the temperature stayed constant the first period over 10/07~13/07 (the thermal behaviour of these periods is further described in Section 6.1), decreased during the second period (14/7-19/07) and significantly increased during the last period (20/7-27/07). Both calculations and measurements showed that the water column was more stratified during the daytime and less stratified during the night time. However, the ELCOM predicted a more stratified temperature field than the observed; this being similar to the temperature predictions using the 1-D heat transport equation described in Section 4.4. The larger mixing coefficients of shear and wind stirring given by Sherman et al. (1978) (see Table 5.1) were also used for

the further ELCOM runs. However, the result was still more stratified than the field measurement as shown in Figure 5.4. It was apparent that heat was transported more efficiently downwards in the reservoir than modelled in ELCOM.

Figure 5.3 a. Temperature calculated by ELCOM and b. temperature measured by SBE 56 temperature loggers at K2 thermister chain

Figure 5.4 a. Temperature calculated by ELCOM with shear coefficient equal to 0.5 and stirring coefficient equal to 2.2 and b. temperature measured by SBE 56 temperature loggers at K2 thermister chain

ELCOM performance was evaluated in greater detail by comparing the calculated temperatures at various depths with the field measurements. For this comparison depths at -0.5 m, -2.8 m, -5.8 m and -8.3 m were used as corresponding to actual thermistor locations shown by the T-chain configuration in Figure 3.6. These depths also represented the temperatures at the surface, the upper water column, the middle water column and the water bottom. The calculated water temperatures at the surface and at -2.8 m depth were close to the observations with high R-square values close to 0.8 when compared to a 45° line (line of perfect agreement) as shown in Figure 5.5a and b. The calculated water temperature at -5.8 m was well correlated with the field measurement as shown in Figure 5.5c but did not fit the 45° line of perfect agreement. Rather ELCOM significantly underestimated the temperature when the temperature value was below 29.2 °C. There was no good correlation between the ELCOM calculation and the thermistor measurements at the depth of -8.3 m (Figure 5.5d) over the whole fieldwork period. This was attributed primarily to the lack of good inflow data as arising from a large gap in the rain gauge data at CP4 (Sg. Tengah in Figure 3.5) over 11/06 to 31/07 2012 which encompassed the entire field work period. When the days with predicted large inflow events, specifically 15/07~22/07 (indicated as black dots in Figure 5.5d) were excluded, the performance of ELCOM improved significantly and the calculation fitted the measurement with a R^2 of 0.65 as shown by the stars in Figure 5.5d. In terms of the thermal structure at K2, ELCOM performed well for the upper water column (-0.5 m up to -2.8 m) but deteriorated for the lower water column.

Figure 5.5 Comparison of temperature between ELCOM results and measurement at K2 over July 2012 field exercise. Stars in (d) indicate temperature values during 10/07~14/07 and 23/07~27/07 without large inflow events

The surface horizontal temperature difference between K2 and the tributaries (M1 to M3 and M8) calculated by ELCOM and shown by the black solid lines in Figure 5.6 were compared with the measured values. The ELCOM results were well correlated to the field observation (correlation coefficient > 0.85) as shown in Table 5.2 with the best being obtained between M8 and K2. The agreement between M3 and K2 shown in Figure 5.6d was reasonably good except that the temperature gradient fluctuated less than observed during the last week of the 3-week field exercise period. Predicted temperature difference between M2 and K2 shown in Figure 5.6c was also in reasonable agreement with field observation though the fluctuation level in the horizontal temperature gradient was also underpredicted. The significant temperature drop observed during 19/07~20/07 shown in Figure 5.6c was missed in the prediction because of unavailable rain gauge (and thus inflow) data at CP4 (Sg. Tengah catchment). According to the field measurements, the water at M1 (Sg. Kangkar) was warmer than in the main water body for the most of the third period (denoted by the blue line shown in Figure 5.6b). In contrast the ELCOM showed the temperature gradient between M1 and K2 reserved on the morning of 22/07, 23/07, 24/07 and 26/07 (denoted by the black line in Figure 5.6b).

Predicted horizontal temperature difference between M1 and K2 also tracked the field measurement for the first two days but the temperature difference became underestimated by ELCOM.

Figure 5.6 a. Surface temperatures from ELCOM and by thermistors at K2; b to e: Surface temperature difference between K2 and tributaries obtained by ELCOM and by thermistors.

Horizontal temperature difference	Correlation
M1-K2	0.91
M2-K2	0.85
M3-K2	0.86
M8-K2	0.96

Table 5.2 Correlation coefficients of horizontal temperature difference between ELCOM calculation and field measurements

5.3 ELCOM prediction of flow patterns

In this section, ELCOM calculation of velocity at two locations K2 and K1 are compared with the corresponding ADCP measurements at these locations. The alongshore velocity pattern obtained via ELCOM and ADCP measurement at K2 respectively, as well as its EOF1 (Empirical Orthogonal Function, mode 1)

component is shown in Figure 5.7. The upper panels show the ELCOM velocity at K2 and its EOF1; while the lower panels the ADCP measured velocity at K2 and its EOF1 component. The EOF contributions to the total variance of the velocity are listed further in Table 5.3. The EOF1 contributed 75% of the total variance of the ELCOM alongshore velocity at K2 and 70% of that of the measured velocity. It is observed in Figure 5.7 that the EOF1 velocities were in reasonable qualitative agreement. The EOF2 component contributes 14% of the total variance of ELCOM velocity and 17.5% of ADCP velocity as indicated in Table 5.3. By comparing Figure 5.8b to Figure 5.8d, the EOF2 component of the ELCOM prediction was different from that of ADCP measurement and the magnitude was smaller, especially in the lower water column. The EOF3 components are not discussed as the contribution of EOF3 was small, less than 10%.

The EOF1 patterns both of ADCP measurement and ELCOM indicated a wind-generated baroclinic flow pattern positive surface current toward the tributary junction over the surface to -2 m depth and negative flow in the middle water column. Such flow pattern was clear during the first period (i.e. over 10/07~13/07) and the third period (20/7~26/07) of the July 2012 field work: During the 2nd period (from 14/07~19/07) there were several rainfall events along with inflows and gate discharges. The baroclinic flow pattern of the 2nd period was therefore not as readily observed as during the other two periods. The EOF1 components also contained the night time differential cooling induced flow pattern: negative surface flow towards the junction and the positive flow towards the dam in the lower water column. Similar to the baroclinic pattern, the night time flow pattern was clear during the first period (i.e. over 10/07~13/07) and the third period (20/7~26/07) of the July 2012 field work. The EOF2 of ADCP measurement also contained underflow during the daytime, which was superimposed to the EOF 1 to give a three-layer flow structure as shown in Figure 5.7c. However, EOF2 of ELCOM did not capture the daytime underflow. Thus the ELCOM could pick up the major features of the alongshore velocity field, i.e. wind-driven baroclinic flow pattern and the night time differential cooling driven flow pattern but it gave a much weaker underflow than that observed.

Figure 5.7 ELCOM calculated velocity along the reservoir main channel (alongshore) and ADCP measurements at K2 and their EOF 1 components

Figure 5.8 ELCOM calculated velocity along the reservoir main channel (alongshore) and ADCP measurements at K2 and their EOF 2 components

K2	ADCP	ELCOM
EOF1	70%	75%
EOF2	17.5%	14%
EOF3	7%	6.6%

Table 5.3 EOF contributions to the total variance of the alongshore velocities at K2 by ELCOM results and field measurement

The hourly averaged alongshore velocities at different depths at K2 are compared. A second order low-pass Butterworth filter with cut-off of 3 hr has been applied on the K2 velocities sampled at 15 min before taking hourly average. The correlation of the velocities at K2 was also calculated and shown in Table 5.4. As seen, ELCOM provided a good prediction of velocity at K2 for the 3-week July 2012 field exercise period and especially during the third week when it was hot and windy with no inflow/rainfall. The correlation during that time period was better than 0.7 for the water column from the surface to -6 m.

Water depth (m)	Correlation coefficients at K2	
	Between the velocities for the entire 3-week July 2012 field exercise period	Between the velocities during the 3 rd week period when the wind was the major forcing
-1.6	0.51	0.79
-4.2	0.66	0.70
-5.2	0.73	0.82
-6.2	0.68	0.83
-8.2	0.20	0.53
-9.2	0.24	0.57

Table 5.4 Correlation coefficients between ELCOM and ADCP velocities along the channel at different depths at K2

Next various linear fits between ELCOM predictions and measurements was calculated for the hourly-averaged alongshore velocity at different depths as shown in Figure 5.9 with perfect agreement being a 45° line. Overall ELCOM gave a velocity field with smaller amplitudes. There were two reasons hypothesized for this under-prediction:

- a) the weakened horizontal temperature gradient in the model generated a smaller underflow during the night-time, which was confirmed by a decreasing slope in the fitted line with depth;

- b) less wind energy/momentum was transported vertically downwards as implied by the low efficiency of the vertical heat transport in ELCOM.

Figure 5.9 Alongshore velocity of ELCOM and ADCP at K2 at the depths of -1.6, -3.6, -6.6 and -8.6 m

Figure 5.10 shows the velocity at K1 predicted by ELCOM and measured by the ADCP. The upper panels show the ELCOM velocity and its EOF1 component; while the lower panels the ADCP velocity and its EOF1 component. It is noted in Table 5.5 that EOF1 component contributed 80.7% of the total variance of ELCOM alongshore velocity and up to 98% of the ADCP measurement at K1. Reasonable agreement was achieved between ELCOM and measurements at K1 but the agreement was not as good as at K2 as seen by the correlation coefficients in Table 5.6. In the discussion below in Chapter 6, it is shown that K1 being located at the overlap zone where two postulated flow circulation loops met, not unexpectedly increased the level of difficulty for modelling the flow. However, both ELCOM and ADCP showed smaller velocities at K1 than at K2, which indicated the ELCOM was able to qualitatively reproduce velocity variation along the reservoir channel.

Figure 5.10 ELCOM calculated velocity along the reservoir main channel and ADCP measurements at K1

K1	ADCP	ELCOM
EOF1	98.3%	80.7%
EOF2	1%	13%
EOF3	0.3%	3.6%

Table 5.5 EOF contributions to the total variance of the alongshore velocities at K1 by ELCOM results and field measurement

Water depth (m)	Correlation coefficients at K1	
	between the velocities during the whole 3-week period	between the velocities during the 3 rd week period when the wind is the major forcing
-1.0	0.45	0.44
-3.0	0.43	0.45
-4.5	0.53	0.63

Table 5.6 Correlation coefficients between ELCOM and ADCP velocities along the channel at different depths at K1

Lastly the hourly averaged cross-sectional velocity profile calculated by ELCOM was compared with the ADCP transect taken across the channel of the reservoir passing through K2. This comparison is shown in Figure 5.11. The ADCP transect was carried out from 12:00~14:00 on 17/07 2012. Figure 5.11d is a composite spanning the corresponding 2 hours of ELCOM calculation for a more realistic comparison. As mention in Section 3.4, 8 points were picked up across the width of the reservoir's main channel intersecting K2. Figure 5.11a presents a velocity field based on interpolation of the measurement taken at these points. Figure 5.11d is the composite spanning 2 hours of ELCOM results. The ELCOM results captured the same cross-sectional velocity pattern as measured by the ADCP transect as shown in Figure 5.11 a & d. The depth-averaged velocity in the shallower water was positive flowing with the wind direction and against the wind in the deeper part.

Figure 5.11 Velocity profiles across channel passing through K2 as obtained from ELCOM calculations and ADCP measurements during 12:00~14:00 on 17/07/2012

5.4 Impact of the grid size on the ELCOM calculation

The effect of the ELCOM grid size on the calculation is discussed in this section. Two comparisons were conducted using grids of 40 m x 40 m (horizontal) x 0.5 m (vertical), 20 m x 20 m x 0.2 m (used in discussion above) and a finer 10 m x 10 m x 0.2 m. One comparison was conducted using computed alongshore velocities at

locations over the range 1400m to 1480 m from the west (denoted by two vertical lines in Figure 5.12) as this part of the channel is deep and it is also where K2 is located. The objective was on resolving the bottom layer with different grid sizes. The other comparison was conducted using the ADCP transect profile. The objective was on ensuring that the grid size adequately reflected the velocity distribution across the channel.

Figure 5.12 Bathymetry over the width of Kranji Reservoir's main channel intersecting K2 with different grid sizes; line with circles denotes grid of 40x40x0.5, line with crosses denotes grid of 20x20x0.2 and line with squares denotes the grid of 10x10x0.2; the vertical lines mark the location of 1400 m and 1480 m from the westmost bank of the reservoir.

The time series of the hourly averaged alongshore velocities were shown in Figure 5.13 for the grid of 40x40x0.5, Figure 5.14 for 20x20x0.2 and Figure 5.15 for 10x10x0.2. It is seen from Figure 5.13 that the alongshore velocity was equal to zero in the vertical range from the bottom to 2 m above the bottom at point 1480 m from the west. At points 1400m and 1440 m from the west, the velocity was zero in a vertical range of ~1 m above the water bottom. The artificial zero-velocity layer became thinner in the grids of 20x20x0.2 and 10x10x0.2; only at points of 1400m and 1420 m from the west was a thin layer (<1 m) of velocity equal to zero near the water bottom. The bottom layer thus was better resolved by the finer grid sizes (20x20x0.2 or 10x10x0.2) than the coarse grid size (40x40x0.5). It was because the

zero-flow layer at the bottom boundaries arose from the nature of the bottom-following z -coordinate system used in ELCOM, as the bottom boundary may have significant changes in elevation for steep lake bottoms (Edwards et al. 2004). The finer grid smoothed the abrupt local topography introduced by the stair-step representation and caused less calculation error in bottom boundaries.

Figure 5.13 Time series of the hourly averaged alongshore velocities with a grid size of 40 m over the cross section intersecting K2 in the range of 1400 m to 1480 m from the west

Figure 5.14 Time series of the hourly averaged alongshore velocities with a grid of 20x20x0.2 over the cross section intersecting K2 in the range of 1400 to 1480 m from the west

Figure 5.15 Time series of the hourly averaged alongshore velocities with a grid of $10 \times 10 \times 0.2$ over the cross section intersecting K2 in the range of 1400 to 1480 m from the west

The other comparison carried out between the ADCP transect and the ELCOM calculations with different grid sizes was shown in Figure 5.16. Overall, the ELCOM results of the three grid sizes showed similar flow patterns across the channel: the positive flow at the western side of the channel and the negative flow at the centre part and the eastern side of the channel. However, the computed results of both the $20 \times 20 \times 0.2$ and $10 \times 10 \times 0.2$ were more resolved than the results of $40 \times 40 \times 0.5$ and reflected more details of the ADCP transect e.g. tight red patches at the west side of the channel. Both the $20 \times 20 \times 0.2$ and $10 \times 10 \times 0.2$ were able to show

sufficiently good computational results and the correlation of the ELCOM results to ADCP measurement for the finer 10x10x0.2 could not be proved to be better than with 20x20x0.2 as the resolution of the ADCP measurement was limited to the 0.5 m bin size. Thus, the 20x20x0.2 grid size was deemed to provide a good compromise between computational time and the accuracy.

Figure 5.16 ADCP transect and ELCOM calculation with different grid sizes

5.5 Summary

In summary, ELCOM calculations are qualitatively in agreement with temperature field in the reservoir and qualitatively captured the important observed features, provided that good inflow data was also available. The agreement however deteriorated in the lower water column probably because the transport model used in ELCOM was less efficient for modelling transport in Kranji reservoir. In the mixing layer model used in ELCOM as reviewed in section 2.4.2 (p. 23 in the revised thesis), the mixing is determined by the turbulence kinetic energy TKE and the local energy required for mixing ηE_{req} . η is the mixing fraction as given by the time step divided by the shorter of the timescales of convective overturns and turbulence mixing as $\eta = \frac{\Delta t}{\min(T_{convection}, T_{turbulence})}$. Mixing occurs when $TKE > \eta E_{req}$.

We have used different coefficient values for TKE calculation but the results were still more stratified than the field measurement. As the ELCOM model was initially developed for temperate lakes that are more stratified than the tropical lakes, the

mixing layer model could be less efficient in tropical shallow lakes by assuming that the mixing occur when $TKE > \eta E_{req}$. The predicted horizontal temperature difference by ELCOM was well correlated with observed though the fluctuation levels were smaller. The difference in these details could be attributed to the lack of precise inflow information or the low precision bathymetry data for both reservoir body and the tributaries. As a response to the weaker temperature gradient in the calculations, the calculated velocity was smaller than the field measurement, particularly during the night time.

For the flow structure, the overall ELCOM results were generally in reasonably good agreement with the ADCP measurements (hourly averaged velocities) for the July 2012 field exercise and especially for the third period of the field exercise. The flow and thermal pattern of the second period was not reproduced due to the lack of the accurate rain and thus inflow data at CP4 (Sg. Tengah in Figure 3.5). The influence of the grid size on the ELCOM results was preliminary discussed in terms of the velocity profiles across the deeper part of the main channel. It is observed that the finer grid can reduce the calculation error at the bottom boundary. However the finer grid required more computational time. The grid of 20 m x20 m x0.2 m was thus selected as it provides a good compromise between computational effort and the calculation accuracy.

While ELCOM have been applied in lakes to calculate the basin-scale motions, e.g. the stratified Sau reservoir, Spain (Vidal and Casamitjana 2008), stratified Upper Lake Constance, Germany (Appt et al. 2004) and a salt stratified large tropical lake, Lake Maracaibo, Venezuela (Laval et al. 2005), few studies report such detailed comparisons of velocity fields with observations in terms of point comparisons, longitudinal velocity profiles with depth and cross-reservoir transverse velocity profile, as shown in this work. Thus the agreement achieved is deemed sufficient to form an accurate qualitative representation of the flow field that is discussed in the next Chapter. In addition, ELCOM have been mostly used in stratified lakes and little in diurnally stratified lakes such as Kranji Reservoir. The results presented here indicate that ELCOM also adequate for modelling tropical shallow lake systems where the forcing is weak and variable.

6. Circulation Patterns in Kranji Reservoir

The flow circulation patterns in Kranji Reservoir are investigated in this chapter. Field observations of the thermal and flow structures from the collected data, particularly from the 10/07-26/07 2012 field exercise, together with numerical calculations of the flow via ELCOM are used to investigate the flow patterns in Kranji reservoir and to address Question 2 proposed in Section 2.5. Specifically the qualitative flow patterns are investigated but now focused on the period of the Southwest Monsoon in Kranji Reservoir since the July 2012 field work was in that monsoonal period with southerly winds prevailing. This chapter looks into the following for the Southwest Monsoon period.

- What are the possible thermal structure and the flow structure in Kranji reservoir?
- How do the meteorological parameters affect the thermal structure in Kranji Reservoir?
- How the reservoir controlling parameters such as inflow affects the hydrodynamics in Kranji reservoir?

6.1 Assembled data

The assembled data used for discussion of the thermal and flow structures in Kranji reservoir included meteorological data collected by the weather station at K2, inflow data, temperature measured at K2 and at the tributaries, and velocity measured by ADCPs at K2 and K1. The K2 weather station measured wind speed, wind direction, rainfall on the reservoir, net radiation, shortwave radiation, air temperature and relative humidity from 10/07-27/07 2012 as described earlier in Chapter 3. The sampling frequency of the data collection was set as 30 seconds and the logging frequency was five minutes. The calibrated XP-SWMM model in Xing et al. (2012) was used to provide the tributary runoff calculated based on the rainfall data recorded by the rain gauges in the five catchments (see Figure 3.5 for the rain gauge locations). It is noted that there was a malfunction of the CP4 rain gauge at the Sg. Tengah catchment from June to July 2012. This necessitated the use of CP4

data for the Sg. Tengah thus introducing inaccuracies for the predicted tributary flow into M2 from Sg. Tengah. Lastly fifteen-minute averages were performed on the measured meteorological data and the XP-SWMM calculated inflow, and the results are depicted in Figure 6.1.

Figure 6.1 Meteorological variables during July 2012 field work

Based on the meteorological data shown in Figure 6.1, the calculation of the net surface heat flux was carried out following Section 4.1.4. The net surface heat flux during the entire July 2012 field work period is plotted in Figure 6.2. The net surface heat flux is used to calculate the physical forcing regimes in Kranji Reservoir following Xing et al (2014).

Figure 6.2 Net heat flux at K2 during the 2012 field work from 10/07-27/07

The temperature was measured by the thermistors at K2 and the tributaries at a frequency of 30 seconds as described in Section 3.1 and Section 3.2 and 15 minute averages are taken. The possibly questionable ADCP measured velocities near the surface, above -1.15 m at K2 and above -1.1 m at K1, were removed as described in Section 3.4. The horizontal velocities U (alongshore), V (cross-shore) were then obtained by aligning the ADCP measured velocities with the principal axes and 15 minute averages performed. The temperature and velocity data are linked to the net heat flux to explain the diurnal variation in thermal and flow structures, which is discussed in next few sections.

6.2 Meteorological data

Observations of the meteorological variables during the July 2012 field exercise (10/07-26/07 2012) are first described for a better understanding of the flow field driving mechanisms. The meteorological conditions shown in Figure 6.1 could be categorized into three periods with each period and its daily averaged wind speed and solar radiation listed in Table 6.1. The first period over 10/07 to 13/07 2012 had total incoming shortwave radiation ranging from 500 W/m^2 to 800 W/m^2 during the daytimes. The smallest short wave radiation during this period was on 11/07 because of a small rainfall event on the afternoon of 11/07. The wind speed, wind direction and its distribution over the entire July 2012 field work is shown in Figure 6.3. Wind mainly came from southeast towards northwest during the first period (10/07-13/07 2012 in Figure 6.3) with average wind speed of 1.8 m/s . Via

comparison, the largest daily averaged wind over the entire 3-week (from 10/07-26/07 2012) field exercise was 2.4 m/s on 12/07 while the daily average wind was about 1.7 m/s. Two moderate inflow events occurred in the evening of 11/07 and in the morning of 13/07. The air temperature exhibited a diurnal pattern: heated up to 30°C during the daytime and cooled down to 26°C during the night time. The moderate short wave radiation caused the air temperature to increase slightly with time during this first period. Overall this first period was characterized as mild solar heating with occasional light rain, i.e. mixed, in contrast to the two other periods as described below.

	Daily averaged wind speed (m/s)	Daily averaged shortwave solar radiation (W/m ²)
First Period: 10/07-13/07	1.8	154
Second Period: 14/07-20/07	1.4	122
Third Period: 21/07-26/07	2.0	231
The whole period:10/07-	1.7	172

Table 6.1 Daily averaged wind speed and short wave radiation during the July 2012 field work

The second period over 14/07 to 20/07 2012 was characterized by frequent rainfall causing large inflows and by relatively mild wind with an average of 1.4 m/s. The largest inflow during the field experiment occurred on 15/07. Accompanied with the rainfall over 14/07-17/07 the wind direction changed as shown in Figure 6.3. As with the first period, the wind mainly came from the south but occasionally from the north when it rained. The daily averaged shortwave radiation value was 122 W/m² for the second period and smaller than the other two periods. The smallest shortwave radiation over the whole period was observed on 14/07 and 16/07. The combined effects of frequent rainfall and the small shortwave radiation made the second period cooler than the other two.

The third period over 21/07 to 26/07 2012 was an interval of hot and windy days with an average wind speed of 2 m/s. A diurnal pattern of wind was observed during this period: strong southeasterly wind, typically 3-5 m/s, from the noon to

the evening and small westerly wind (typically <1 m/s) from the night to the early morning. There was no rainfall and inflow during this third period. The daily averaged total incoming shortwave radiation was 231 W/m^2 largest of the three periods. The strong heating during the daytimes induced a remarkable increase in air temperature that was the strongest of the three periods. The air temperature on the afternoon was increased by about 2°C from 30°C to 32°C .

The days over these 3 periods were classified into the physical forcing regimes defined by Xing et al. 2014 and briefly described in Section 4.4.2: solar radiation dominated regime with $I_{WB}<1$ and $I_{IW}<0.4$; windy regime with $I_{WB}>1$ and $I_{IW}<0.4$ and cold inflow regime with $I_{WB}<1$ and $I_{IW}>0.4$. The daily averaged I_{IW} and I_{WB} together with the daily average solar radiation are plotted in Figure 6.4. During the four days of the first period, there were two days (11/07 and 12/07) which were cold inflow days and the two other days (10/07 and 13/07) being solar radiation. The cold inflow regime dominated over the second period (14/07 to 20/07) while the solar radiation was the dominant physical forcing with its daily average all above 200 W/m^2 during the third period (21/07 to 26/07 2012). Overall, 8 days out of 17 days during the July 2012 field experiment were the cold inflow days and the rest 9 days were the solar days. There was no windy regime days predicted during this three-week deployment period. The classification of the forcing regimes facilitates the next discussion of how the thermal structure and the flow structure respond to different meteorological conditions in Kranji Reservoir.

Figure 6.3 Wind during July 2012 field work: wind direction and wind speed

Figure 6.4 Daily average I_W (0:00-23:59), I_{WB} (06:00-18:00) and solar radiation during the July 2012 field work with the black line denoting $I_W=0.4$

6.3 Thermal structure and flow structure of the reservoir

6.3.1 Thermal structure in the reservoir

The measured water temperatures at K2 and at the four tributaries during the July 2012 field are shown in Figure 6.5 and Figure 6.6. The time evolution of the water temperature was similar to the air temperature. The water temperature remained almost constant for the first period with its milder incoming radiation and light rain. The water temperature dropped during the second period as affected by the small incoming radiation and frequent rainfall/inflow. The water temperature was heated up during the third period in response to the strong solar heating.

Figure 6.5 Temperature profiles at K2 during the July 2012 field work with the thick black lines demarcating the three periods

Figure 6.6 Measured surface temperatures at K2 and at the four tributaries M1-M3 and M8 with the thick lines demarcating the three time periods

The observed surface mixed layer during the daytime stratification is thin, of less than 2 m. During the night time, the convective cooling was the major forcing to deepen the surface mixed layer to the water bottom, as demonstrated by TKE budget analysis over the three SCAMP days described in Section 4.2. On the other hand, the temperature contour in Figure 6.5 showed the water column was also cooled from the bottom after midnight (e.g. 23/07 as marked in Figure 6.5) which shifted the isotherms upwards during the third period (21/07-27/07). This is later shown in Section 6.4 to originate from a density driven cold current that caused the reservoir to be stratified from the bottom during the night time. The thickness of the surface mixed layer thus arose from a combined effect of convective cooling and bottom cold flow.

Water temperature of tributary inflows

The water surface temperatures at the tributaries and at the main body of the reservoir (K2) showed a daily horizontal surface temperature difference cycle in response to differential heating and differential cooling as shown in Figure 6.6 for the July 2012 field exercise. This was due to the depth difference between shallow tributaries and the deeper main body. The water depths at the individual tributaries

are also different and thus the details of horizontal temperature difference varied across the tributaries. M1 at Sg. Kangkar is about 0.8m deep, the shallowest among the four measurement locations M1-M3 and M8. M1 thus tended to have a higher surface temperature than the other tributaries as well as the main reservoir body after 13/07 as shown in Figure 6.6a. Compared to the surface temperature at K2 in Figure 6.6b, the surface water at M1 was only cooler for 10/07 till the morning of 12/07 and for midnight to the early morning periods of 15/07, 17/07, 18/07, 20/07 and 25/07. It was warmer than at K2 for the most time of the third period (21/07-27/07). A maximum temperature difference of 2 °C occurred on 21/07 and 22/07 between M1 and K2.

The temperature was lower at M2 (Sg. Tengah) than at K2 for the first ten days (10/07-20/07), which indicated frequent inflows coming into the reservoir from M2. A large cold inflow created a temperature difference as large as -1.7 °C between M2 and K2 on the evening of 19/07 to the morning of 20/07. For the last week (21/07-27/07), the effect of differential cooling and heating arising from the very different depths became obvious. The water at M2 was warmer than that at K2 when Sg. Tengah was heated up and became cooler when it was cooled down. The temperature difference oscillated between -1°C and 1°C.

M8 (Sg. Pangsua) surface water was cooler than K2 for the most of time before 21/07, except for the afternoon of 10/07. The water column heated up rapidly at M8 on that afternoon of 10/07 and became warmer than at K2. For the last week, temperature difference varied similarly to that between K2 and M2 but with smaller amplitude.

The horizontal temperature difference between M3 (Sg. Pengsiang) and K2 reversed daily for the first period and the third period: temperature increasing during the daytime and decreasing during the night time from M3 to K2. During the second period, the tributary M3 was cooled down by inflows and its surface temperature was about ~1 °C lower than K2 surface temperature.

These observed temperature difference between the main water body (K2) and four tributaries indicated that a convective circulation was probably set in motion via horizontal temperature differences as caused by the differential cooling/heating.

The vertical thermal structure and the inflow

The temperature difference between the surface and bottom waters (i.e. the stratification) as measured by the K2 thermister chain was small in Kranji reservoir, especially when compared to its temperate counterparts (see Figure 6.7). The amplitude of the stratification was above 1 °C during the day time and below 0.5 °C during the night time with a peak value of 1.2 °C occurring during 20/07-22/07. The temperature contour in Figure 6.5 and the tributary surface temperature in Figure 6.6c indicated a large cold inflow that plunged into the reservoir from M2 (Sg. Tengah) from the evening of 19/07 to the morning of 20/07. The inflow being ~2°C (see Figure 6.6c) colder than the main water body enhanced the stratification and made the lake stratified for 3 days (see $\Delta T > 0.5$ from the afternoon of 20/07-22/07 in Figure 6.7). On the other hand, the minimum temperature difference occurred on the nights of 16/07-17/07 during which inflows occurred inflows on 16/07 as shown in Figure 6.1. Thus, it is observed that the inflow can also remove vertical stratification in the Kranji Reservoir by significantly mixing the reservoir horizontally.

Figure 6.7 Temperature difference between surface and bottom at K2 during July 2012 field work

To further investigate the mixing effect of the inflow on the vertical thermal structure, averages were taken on normalized inflow-induced buoyancy B_{norm} and the surface-bottom K2 temperature difference denoted as ΔT_{K2} , over the cooling

periods i.e. the night time with negative net surface heat flux (over 19:00-07:00) of the cold inflow days, as shown in Figure 6.8. These cold inflow days are defined with $I_{WB} < 1$ and $I_{IW} > 0.4$ by Xing et al. (2014) and thus excluded other major impacts such as the wind stirring and large varying solar radiation heating on the vertical thermal structure. The normalized inflow-induced buoyancy is defined as below as

$$B_{norm} = \frac{\Delta T_0 V_{inflow}}{T_{lake} V_{lake}} = \frac{\Delta T_0 Q \Delta t}{T_{lake} V_{lake}} \quad (6.3.1)$$

where ΔT_0 is the temperature difference between the inflow temperature and the surface temperature at K2, Δt is the duration of the cooling period equal to 12 hours (43200s), and Q is the mean inflow rate (m^3/s) over the duration. T_{lake} is the average lake temperature, assumed to be the average temperature over the water column at K2. The water volume in the Kranji Reservoir is estimated by equation (6.3.2) :

$$V_{lake} = A_s * H \quad (6.3.2)$$

where A_s is the surface area of the Kranji Reservoir equal to $3 \times 10^6 m^2$ and H is the mean depth of 6.7m. The lake temperature and volume are introduced in equation (6.3.1) in order to assess the change introduced by the inflow buoyancy to the lake's heat content. The values are replotted in Figure 6.9. The surface-bottom K2 temperature difference ΔT_{K2} is plotted against the normalized inflow induced $B_{normalized}$ and against the net inflow flow rate ΔQ calculated in equation (6.3.3) and represents the net flow coming into the reservoir.

$$\Delta Q = \frac{A_s \Delta H}{\Delta t} \quad (6.3.3)$$

where ΔH denotes the depth change within the cooling periods measured by the RBR depth sensor. It is observed from Figure 6.9a that for $B_{norm} < 0.2 \sim 0.3 \times 10^{-4}$, ΔT_{K2} increased with B_{norm} , i.e. the inflow enhanced the vertical stratification by creating a colder underflow layer. Conversely ΔT_{K2} decreased with B_{norm} when $B_{norm} > 0.2 \sim 0.3 \times 10^{-4}$ i.e. the inflow with large buoyancy now augmented the mixing. It is noted that the negative net inflow also occurred as seen from negative ΔQ in

Figure 6.9 b. This was due to the gate discharge operations. Such gate discharge likely further reduced ΔT_{K2} because it further caused horizontal motions and horizontal mixing. However the impact of the large inflow buoyancy in reducing ΔT_{K2} was still observed (open circles in Figure 6.9a). Thus, the normalized buoyancy B_{norm} provided a qualitative indicator of the effect on increasing/reducing the stratification and more cold inflow day data are required to obtain a more accurate description.

In summary, the cold inflow could either enhance the vertical stratification or make the temperature in the reservoir uniform if the buoyancy flux from the inflow was sufficiently large. The inflow's influence on the thermal structure was a combined effect of the inflow rate and the temperature difference between the tributaries and the main water body. Additionally, when there were gate operations, the created horizontal motions also reduced the vertical stratification.

Figure 6.8 Inflow-induced buoyancy B and ΔT_{K2} averaged over the cooling period (19:00-07:00) during the 2012 field work with the closed circles denoting the cold inflow days

Figure 6.9 a. K2 Surface-bottom temperature difference ΔT_{K2} versus normalized inflow-induced buoyancy $B_{normalized}$ and b. ΔT_{K2} versus the net inflow rate (m^3/s) during the cooling periods (19:00-07:00) of the cold inflow days with the closed circles denoting the occurrence of the gate opening

6.3.2 Flow structure in the reservoir

The most significant feature captured by the ADCP velocity profiles at K2 was the diurnal cycle of velocity structure during the solar radiation dominant days (and especially during the third period over 21/07-26/07) as shown in Figure 6.10 . This was observed to be closely related to the net surface heat flux also plotted in the figure, specifically a three-layer velocity structure during the heating period (defined as the daytime hours of 08:00-18:00) and two-layer structure during the cooling period (defined as the night time hours of 19:00-07:00). The three-layer structure during the heating period represented a typical 2nd mode V2 internal wave mode (Munnich 1996) with two vertical nodes (zero velocity) in the vertical velocity profile as shown in Figure 6.11a. This arose from the southerly wind that pulled the surface flow in the wind direction from the three-arm tributary junction (south) towards the dam (north). For the surface to remain nearly horizontal, the water in the middle water column moved against the wind direction from the dam (north) to the junction (south). A thin layer of water near the bottom as arising from

the underflow⁸ that was shown later to be generated by the differential cooling on the previous night flowed towards the dam again thus forming the three-layer structure.

Figure 6.10 ADCP velocity measurements at K2 and at K1 together with profiles of calculated net surface heat flux

⁸ In order to differentiate from the inflow caused by the rainfall events in the catchments, ‘underflow/undercurrent’ was used to refer to the submerged flow induced by the differential cooling in the lower water column.

Figure 6.11 Vertical velocity profiles at K2 and K1 at 13:00,15:00, 23:00 of 24/07 and 01:00 of 25/07 ; both days were solar days.

For the cooling periods (negative net surface heat flux during night time), the vertical V1 mode became dominant at K2 (Figure 6.11c) with a weak surface current towards the junction and a strong underflow towards the dam. This was triggered by the differential cooling between the tributaries and the main water body (see Figure 6.6 showing the cooler tributary temperatures during the night time). From Figure 6.10 it is seen that the underflow became strong right after the surface cooling (negative net surface heat flux) started. As the V2 mode occurred consistently during the daytime (07:00-19:00) and the V1 mode mostly during night time (19:00-07:00), the terms daytime flow and the night time flow at K2 are used in the next paragraphs.

The flow at K1 was much weaker than the flow at K2 and it also varied similarly diurnally as shown in Figure 6.10c. During the daytime, the flow structure was two-layered (see velocity profiles of 13:00 and 15:00 on 24/07 in Figure 6.11b). The surface current flowed with the wind towards the north and the water in the lower water column moved southwards against wind. However, no undercurrent was observed near the bottom. Indeed during the night time, the water column at K1 was calm and velocity was close to zero and with direction changing. For example, before midnight of 24/07, the water moved towards the junction but it changed its

direction and moved towards the dam after the midnight of 24/07 (Figure 6.11d). There was no underflow at K1 throughout the days.

During the cold inflow days defined in Section 6.2 according to Xing et al. (2014)'s classification, the diurnal cycle described above was also observed at K2 (see Figure 6.10 for e.g. 14/07, 15/07 and 16/07) but the pattern was not as distinct as during the last week which were solar radiation dominated days. The vertical velocity structure exhibited a V1 mode during the daytime of some cold inflow days, e.g. the daytime of 12/07 and the daytime of 18/07. Due to lack of good inflow data and especially arising from the lack of reliable rainfall data at CP4 (Sg. Tengah catchment), the velocity structure in the cold inflow days will not be further discussed. Rather the remaining discussion is focused on the velocity and the flow pattern during the solar radiation dominated days.

6.4 Circulation pattern in Kranji reservoir

Different flow structures were displayed during daytime and night time by both ADCP measurements and ELCOM calculations, indicating different driving mechanisms for circulations in Kranji reservoir. These flow patterns are discussed in terms of the daytime circulation and the night time circulation below. As mentioned in Section 5.2, the lack of inflow data, specifically rainfall data at CP4 (Sg. Tengah), limited the performance of the ELCOM model. Thus the circulation pattern is discussed during the solar radiation dominated days of the third period (from 21/07-27/07), when there was no inflow. It should be noted that the third period are wholly solar dominated days but during these days the wind was the major forcing driving the basin-scale circulation patterns. This is because I_{WB} in Xing et al.'s (2014) classification only represents the competing effects of the heating and wind mixing in the vertical direction and it does not evaluate the impact of the wind on driving the basin-scale flow patterns.

6.4.1 Daytime circulation pattern: along channel upwelling and downwelling

Figure 6.12b shows the ELCOM temperature calculation along the channel during 12:00 on 24/07 2012. It shows a surface horizontal temperature difference of about

0.6°C established between the junction and the dam. This temperature difference was also qualitatively seen on the other days during the third period of the July 2012 field exercise (e.g. on 23/07 as shown in Figure 6.13). The cause of the horizontal temperature difference is likely the southerly wind (wind from south). The southerly wind pushed the surface water towards the dam (north). The surface water converged and sank near the windward side of the reservoir (downwelling). To maintain the surface at nearly horizontal, the water in the middle water column came up near the upstream tributary junction end (upwelling). The southerly wind was therefore judged to be the most important driving forcing to the daytime circulation pattern in Kranji reservoir over the July 2012 field exercise.

Figure 6.12 a. Velocity along the channel; b. Temperature along the channel at 12:00 24-Jul-2012

Figure 6.13 a. Velocity along the channel; b. Temperature along the channel at 12:00 23-Jul-2012

More evidences of this pattern show up via the time series of the buoyancy frequency and the velocity. Figure 6.14 is the colour plot of the buoyancy frequency in log scale superimposed with temperature contour during 00:00 21/07 - 23/07 2012. Although weak, the thermocline produced by the cold undercurrent of the previous night time differential cooling persisted in the lower water column throughout the daytime. It was then gradually suppressed downwards when the daytime wind forcing became dominant over the horizontal temperature difference. In the wind-affected water column, a new stratification denoted by a strong thermocline was established from the surface heating. The water column was

therefore divided into three layers. The toward-dam submerged flow was closely related to the weak thermocline. It decayed with the suppression of the weak thermocline and result in the two-layer flow structure being developed along with the new stratification in the wind-affected upper water column. Thus the daytime circulation in Kranji was a combined result of the wind-induced circulation and the density-induced circulation. Figure 6.15a depicts schematically the two circulations in Kranji reservoir during the daytime (07:00-19:00): a strong wind-induced circulation in the upper two layers and a weak density-induced circulation in the lower water column.

Figure 6.14 a. Time series of the buoyancy frequency N at K2 during 21/07-24/07 superimposed with temperature contours; b. Time series of the velocity along the channel at K2 during 21/07-24/07 superimposed with temperature contours

Figure 6.15 Schematic showing the circulation patterns in Kranji reservoir during the daytime and the night time

A mean basin-scale circulation is defined in terms of the depth-integrated magnitude of the alongshore velocity.

$$q = \int_{-h}^0 U dz \quad (6.4.1)$$

which was readily computed from ELCOM. Figure 6.16a depicts the basin-scale circulation during the daytime (07:00-19:00) on 23/07 2012. Such a flow pattern was observed every day during the third period covering 6 days and the averaged daytime flow pattern over this third period is shown in Figure 6.17a. In Figure 6.17a, positive velocity is towards the dam and negative towards the junction. The circulation loops, as predicted by ELCOM were shown by blue streamlines in Figure 6.16 and Figure 6.17.

Referring to Figure 6.17a for the daytime circulation, one large loop is seen to occur along the main channel in the range from 0-3500m from the dam, although there was a break in the loop over 2500-2800m from the dam. The mean flow moved towards the dam on both sides of the channel, where the water is shallower (as also seen from the traverse plots of alongshore velocity in Figure 6.18a). It moved towards the junction in the deep water in the deeper central part of the channel as seen in Figures 6.17a and Figure 6.18a. Thus if the major source of nutrients into the reservoir was assumed to come from the tributaries, both shallow sides of the

channel could be the major nutrient transport path into the reservoir during the daytime and the deep central part of the channel could be the path of the mass transported out of the reservoir.

Figure 6.16 Circulation pattern during the daytime and the nighttime from 07:00 23/07-07:00 24/07

Figure 6.17 Circulation pattern averaged over the daytime periods and the nighttime periods respectively for the third period (21/07-26/07 2012)

Figure 6.18 Traverse plots of the alongshore velocity across the channel passing through K2 averaged over the daytime and the nighttime from 19:00 22/07~19:00 23/07

6.4.2 Nighttime circulation pattern: along channel differential cooling

In Kranji reservoir, the night time circulation was mainly driven by differential cooling. The ADCP measurement at K2 indicated there was a strong undercurrent towards the dam (the north) from the junction (south) and a return flow at the surface (see earlier Figure 6.11c). It is also observed from Figure 6.10 that the occurrence of the strong undercurrent was coincident with the cooling period. Figure 6.19 shows the ELCOM result along the channel at 03:00 23/07/2012. The upper panel is the alongshore velocity and the positive value denotes the velocity towards the dam. The lower panel shows the calculated temperature along the channel. As the dam end is deeper than the junction, the differential cooling made the water at the junction cooler than that at the dam. The ELCOM temperature along the channel confirmed the existence of the differential cooling. For example, the horizontal temperature gradient was approximately 0.8°C between the dam end and the junction during the night of 23/07 as seen in Figure 6.19. Similar observations were also made on the other nights during the third period (e.g. the night of 24/07 in Figure 6.20). Hence, there occurred a horizontal density driven circulation in the distance between about 600 m and 2000 m away from the dam: a weak warm surface flow (denoted by blue patches in Figure 6.19a and Figure 6.20a) from the dam towards the junction, and a strong submerged current (denoted by red patches in Figure 6.19a and Figure 6.20a) from the junction towards the dam. These features are in qualitative agreement with the field observation show in Figure 6.10b.

Figure 6.19 a) Velocity along the channel b) The temperature along the channel at 03:00 23/07 2012

Figure 6.20 a) Velocity along the channel b) The temperature along the channel at 03:00 24/07 2012

Both the field measurement (Figure 6.10c) and the ELCOM results (Figure 6.19a and Figure 6.20a) indicated there was no strong night-time undercurrent in the velocity profile at K1, which is about 2500 m away from the dam as measured along the main channel. The temperature contour in Figure 6.19b and Figure 6.20b shows complete mixing (almost vertical isotherms) occurring at K1. It was the region where the surface return flow changed its direction and the cold undercurrent originated. The vertical isotherms also suggested a small horizontal pressure gradient. As a result, a much smaller along channel velocity is observed at K1 than at K2.

Lastly, the calculated water temperature at the tributaries was higher than at K1 in Figure 6.19b and Figure 6.20b, which was consistent with the field observation that the tributary surface temperature was larger than at K2 (seen in Figure 6.6 during the 3rd period). Some blue patches were observed in the lower water column in the range between 2500 m and 3500 m from the dam, which was evidence of a second but rather weak circulation between the K1 and tributaries.

The mean basin-scale flow pattern during the night time (19:00-07:00) in Kranji reservoir was simpler than during the daytime as the major forcing was only differential cooling. As shown in Figure 6.17b, two major circulations were found in the main channel. The bigger one started from the dam and extended up to ~2500 m from the dam. The undercurrent usually with a lower temperature tended to sink and flowed in the lower part of the cross section. The water is shallower at two sides and deeper in the middle over the cross section in the main channel. It is thus noted from the traverse plot of the alongshore velocity (Figure 6.18 for location K2 across channel) that the positive flow occurred in the middle of the channel, which is the deepest part of the channel. Due to the zero volume mass transport over the cross section, the flows at the two sides of the main channel tended to be negative towards the junction to balance with the flow in the middle. Thus during the night time, within the big circulation loop, the shallow sides of the channel could be the transport path of the mass out of the reservoir and the deep central part of the channel could be the transport path of mass into the reservoir.

Both the mean flow pattern in Figure 6.17 and the velocity profile along the channel in Figure 6.19 and Figure 6.20 confirmed that there was a weak second circulation between K1 and the tributaries during the night time. K1 was in the zone where the two circulations met and the surface flows changed its direction, which agreed with observations from the ADCP measurement at K1 in Figure 6.10c and Figure 6.11d, i.e. a flow with relative uniform and small velocity throughout the water column and changing direction from time to time. The small loop in the main channel is expected to be an important factor to nutrient transport to the lake. It could trap the mass from the catchment in this part of the lake (2500~3500 m from the dam) and hampered the exchange between the catchments of KangKar, Tengah, Peng Siang, and this has been supported by the high chlorophyll a concentration calculated by CAEDYM to later shown in Chapter 7.

6.5 Summary

The meteorological conditions during the July 2012 field exercise could be categorized into three periods: the first 4-day period characterized by moderate wind, moderate heating and slight rainfall with 2 solar days and 2 cold inflow days; the second 7-day period characterized by large rainfall events, small wind and large inflow events with all seven days being the cold inflow days and the third 6-day period characterized by strong heating, strong wind but no rainfall and inflow. All six days of the third period are solar radiation dominated days according to the forcing regime classification of Xing et al. (2014).

In the solar radiation dominated regime, both temperature and velocity measurements at K2 showed that a daytime three-layer structure and a night time two-layer structure were the typical thermal and flow structures in Kranji reservoir during the July 2012 field experiment which is in the Southwest Monsoon period. The daytime flow circulation was the result of the combined effect of the wind induced circulation and the horizontal temperature gradient induced circulation. Firstly, the wind played an essential role in the daytime flow circulation with the southerly wind causing the upwelling near the junction and the downwelling near the dam, and further established the daytime horizontal temperature gradient. The

underflow generated from differential cooling during the previous night did not deteriorate immediately when the heating began as the new horizontal temperature caused by the upwelling/downwelling supplied a pressure gradient to maintain the underflow. It also caused a weak thermocline in the lower water column. As the heating continued, stratification with a stronger thermocline developed from the surface and gradually suppressed the weak thermocline and the undercurrent in the lower water column. The night time flow pattern was driven by the differential cooling due to the varying bathymetry along the channel. The night time flow pattern also explained the behaviour of the flow structure at K1 with no underflow and a two-layer structure during the daytime, and a uniform velocity profile with direction changing during the night time. K1 was also the location where the ELCOM-predicted two night time circulation loops met. The second weak circulation loop between K1 and the tributaries was driven by the higher tributary temperatures and lower temperature at K1, which could be important to the Kranji ecosystem (later discussed in Chapter 7).

In this chapter, we have investigated the basin-scale circulations and found the circulation in Kranji reservoir was alternatively determined by two different mechanisms: daytime circulation by wind and nighttime circulation by differential cooling. Such circulation patterns are important for understanding the mass transport in the reservoir. In the next chapter, we will show how these circulation patterns influenced the phytoplankton distribution in Kranji reservoir. Furthermore, the analysis in this chapter reveals the short timescale variability of the driving forcing of the basin-scale circulation based on the three dimensional calculation results. In the earlier chapters, we investigated the short timescale variability of the vertical mixing of the local water column based on the data achieved at one point of Kranji reservoir. A more comprehensive dynamical picture of the vulnerable dynamics in Kranji reservoir thus is developed based on the one dimensional and three dimensional studies.

7. Role of Vertical Mixing and Circulation in Phytoplankton Distribution

In this Chapter the temporal and spatial distributions of chlorophyll a concentration are qualitatively linked to the vertical mixing and circulation patterns in Kranji reservoir presented in Chapter 6. Vertical chlorophyll a concentration profiles are available from the SCAMP measurements of the July 2012 field exercise (described in Section 3.5 and Section 4.1) and from a further sparse chlorophyll a dataset as made available by PUB via their periodic surface water quality sampling at 4 locations in Kranji reservoir.

7.1 Transport of chlorophyll-a via vertical mixing

The relationship between the SML behaviour and the vertical distribution of phytoplankton biomass at K2 is first compared in Figure 7.1 for the 3 days in July 2012 of available SCAMP data. Comparing the temperature (right panel) and chlorophyll a (left panel), it is observed that the phytoplankton biomass represented by the chlorophyll a concentration had a similar vertical structure as temperature. The highest concentration occurred near the surface and the concentration decreased with depth. Both the maximum temperature gradient depth (thick red solid lines in Figure 7.1) and the TKE budget modelled SML depth (circled red lines) tracked the max chlorophyll-a concentration gradient (dashed black lines with dots) well, suggesting that the phytoplankton biomass which concentrated near the surface during the day time would be transported downwards with the night time SML deepening and mixed over the water column. The convective cooling, the major forcing to the deepening, thus played an important role in shaping the vertical distribution of phytoplankton biomass.

Figure 7.1 Three 24 hour SCAMP measured temperature and chlorophyll a concentration in July 2012 with red solid lines denoting the SML defined by the maximum temperature gradient, circled red lines denoting the SML found by the TKE budget analysis and dashed black lines with dots denoting the depth where the maximum chlorophyll a concentration occurred

The vertical flux of chlorophyll a concentration was estimated by the Fick's law of diffusion:

$$J_{chla} = -K_z \frac{\partial Chla}{\partial z} \quad (6.4.2)$$

where the chlorophyll-a concentration was measured by the SCAMP fluorometer and positive flux indicates downwards and the K_z follows Equation 4.1.13 (Osborn and Cox 1972). Figure 7.2 shows the magnitude of vertical flux of chlorophyll a concentration on log scale. The chlorophyll a flux vertical distribution pattern in Figure 7.2 was similar to the vertical heat flux in Figure 4.25: strong chlorophyll a flux in the surface mixed layer and in the underflow layer and weak chlorophyll a flux in-between. A depth average was performed across the SML, the underflow layer and the layer between the SML and the underflow layer as for the heat flux (see Section 4.4, Tables 4.2 to 4.4), and the results are listed in Table 7.1 to Table 7.3. In terms of the temporal variation, the vertical flux of chlorophyll a concentration was suppressed during the daytime measurement periods 11:00-19:00

(Table 7.1) and became active during the nighttime and the early morning periods (from 19:00-07:00) (Table 7.2 and Table 7.3). It was because the measured vertical eddy diffusivities were smaller during the daytime than nighttime as shown in Figure 4.24, and overcame the larger daytime vertical gradient of chlorophyll a concentration as seen in Figure 7.3. Also seen in Figure 7.3 for the late night and early morning periods (03:00-07:00), the vertical gradient of chlorophyll a concentration became smaller and negative and thus there were more negative vertical fluxes of chlorophyll a concentration observed in Table 7.3, i.e. chla fluxes transported upwards (positive flux value is downwards here). This could arise from the night time phytoplankton migration over the background of well mixed water column. As the vertical distribution of chlorophyll a concentration was more uniform (via vertical mixing) during the late night period, the small vertical migration of the phytoplankton became more visible and more important to the vertical distribution of the phytoplankton mass.

Figure 7.2 Vertical fluxes of chlorophyll a concentration in log scales; dash lines denote the temperature contour with a contour level of 0.1 °C; circled red lines denote the depth where the maximum temperature gradient occurred and circled black lines denote the depth where the temperature was 0.02 °C lower than the surface temperature.

Figure 7.3 Vertical gradient of Chlorophyll a concentration (mg/m^4) during the three 24-hr SCAMP measurements

In terms of the chlorophyll a flux vertical structure, the most active chlorophyll a fluxes came from the underflow layers. The flux of phytoplankton biomass within the underflow (underflow layer in Table 7.1-Table 7.3) was larger than that in the SML for most of the time. Since sediments are usually a nutrient pool, biomass actively transported downwards could be re-introduced into the water column via mixing from the underflow (e.g. the negative value in Figure 7.3). In addition, the underflow brings the biomass from the shallower part of the reservoir, which will be later introduced in the local water column. The role of the underflow to the Kranji ecosystem deserves further investigation.

	Surface mixed layer	Layer between the SML and the underflow layer Upper hypolimnion	Underflow layer
11/07-12/07	1.4×10^{-4}	3.2×10^{-4}	6.4×10^{-4}
18/07-19/07	0.0010	0.0092	
25/07-26/07	8.9×10^{-6}	-4.0×10^{-6}	8.0×10^{-4}

Table 7.1 Vertical fluxes of chlorophyll a concentration ($\text{mg}/(\text{m}^2\text{s})$) within each layer of the water column at K2 during the daytime heating period of 11:00-18:00

	Surface mixed layer	Convective patches (between the surface and the depth at which $T=T_{\text{surf}}-0.02$)	Hypolimnion (underflow region)
11/07-12/07	0.0012	4.6×10^{-4}	0.048
18/07-19/07	0.0182	0.0116	0.026
25/07-26/07	0.0262	0.0115	-0.0316

Table 7.2 Vertical fluxes of chlorophyll a concentration ($\text{mg}/(\text{m}^2\text{s})$) within each layer of the water column at K2 during the daytime heating period of 19:00-03:00

	Surface	Convective patches	Hypolimnion
11/07-12/07	-0.0011	-0.0011	6.3×10^{-4}
18/07-19/07	0.003	0.0016	0.0020
25/07-26/07	0.0034	-0.0014	0.055

Table 7.3 Vertical fluxes of chlorophyll a concentration ($\text{mg}/(\text{m}^2\text{s})$) within each layer of the water column at K2 during the cooling period 03:00-07:00

7.2 Spatial distribution of chlorophyll a under the southerly wind typical of the southwest monsoon

ELCOM coupled with CAEDYM (Computational Aquatic Ecosystem Dynamics Model) was run for the third period (21/07-26/07) to investigate the relationship between spatial distribution of chlorophyll a and the circulation patterns. For the ELCOM-CAEDYM run, the 40x40x0.5 grid size was selected as the computational time was much reduced and in this work, only the qualitative relationship between the chlorophyll a concentration and the flow patterns was investigated. In the ELCD run, three phytoplankton species were modelled, Cyanobacteria, Nodularia and freshwater diatoms, and the default set of the CAEDYM constants was used (Hipsey et al. 2006). These constants included phytoplankton constants, light limitation and nutrient parameters. As water samples were only collected at K2 during the July 2014 field work period to calibrate the SCAMP chl a concentration measurements, no nutrient data were available for the tributaries. Thus nutrient inputs DO, NH₄, NO₃ came from fortnightly PUB sampling results and the other values were adapted from the CAEDYM run for the Kranji Project (Lo 2008). Furthermore, the nutrient inputs were set as the same for all the tributaries and set as constant during the whole run. In this way, the influences of different nutrient loading and phytoplankton biomass of the tributaries on the spatial distribution of chl a concentrations were excluded. The input nutrient concentrations (in mg/l) used are listed in Table 7.4

DO	NH ₄	NO ₃	PO ₄	DONL	DOPL	DOCL	POCL	SiO ₂	SSOL1
5	0.18	1.95	0.03	0.23	0.01	5.17	0.43	3.54	183.94

Table 7.4 Nutrient inputs (in mg/l/2min) set for the CAEDYM run

This third period (21/07-26/07) with no rainfall in the catchment was chosen for running ELCOM-CAEDYM (ELCD) as there was the least uncertainty in the inflow. The results are displayed in Figure 7.4 for one day (24/07) averaged over daytime 07:00-19:00, and in Figure 7.5 for the whole third period (21/07-26/07) averaged for daytime 07:00-19:00. The computed daytime results are presented here as later it is compared to the PUB sampling results where water sampling was

conducted during working (i.e. daytime) hours. The two black solid lines indicate the locations of K1 and K2. It is observed in Figure 7.5 that there is a decreasing surface chla concentration in the range from 0-2000 m followed by an increasing surface concentration in the range from 2000-3500 m. This spatial distribution of chla concentration was close to the spatial distribution of the two nighttime loops in the main channel (see Figure 6.17), and with the small night time loop circulating the nutrients from the catchments of Tengah, Pengsiang and Kangkar within the main channel from 2500-3500 m. Although the water temperature in this region (i.e. 2500-3500 m) may be lower than the dam, the phytoplankton was supplied with more nutrients via the circulation and thus more phytoplankton was observed in the main channel near the junction.

High chlorophyll a concentration also occurred at the dam end of the lake (Figure 7.5). As depicted in Figure 7.1, the vertical structure of chla concentration was similar to the vertical temperature structure: high chla concentration in the surface layer that decreased with depth. The southerly wind pushed the surface water carrying high concentration of chla towards the dam which accumulated over there. Furthermore, the dam water was warmer and more stratified than near the junction, which favoured cyanobacteria growth. This surface transport and the warm stratified water at the dam both increased the chla levels there.

Figure 7.4 ELCOM-CAEDYM results averaged over daytime (07:00-19:00) of 24/07 2012

Figure 7.5 ELCOM-CAEDYM results averaged over the day time periods (07:00-19:00) of 21/07-26/07/2012

Figure 7.6 shows the PUB sampling locations. The PUB sampling was usually conducted fortnightly. The complete chlorophyll a concentration dataset for comparison with ELCOM were only available at St4, St3, K1 and St1. The corresponding locations of St4, St3, K1 and St1 were also denoted by the black solid lines in Figure 7.4 and Figure 7.5. It is noted that K1 was very close to St 1 being just downstream of Station 1. The samples were taken from 0.5 m below the surface. St 3 and St 4 were located within the deeper northeast part of the reservoir. Figure 7.7 showed the chlorophyll a concentration at sampling locations over June to August for 5-years years (2008 to 2012) when PUB data was made available. The average was taken over the 3-month (Jun to Aug) period corresponding to the Southwest Monsoonal period. Location K1 always exhibited the lowest concentration among the four stations. Although K1 and St 1 were very close, the chlorophyll a concentration at St 1 was significantly higher than at K1, indicating the possibility of a sufficient nutrient availability at St 1. The concentration at St 3 and St 4 were also much higher than at K1, indicating the growth of phytoplankton

was favoured near the dam due to the warm water temperature and the surface transport of waters with high chlorophyll a concentration. It should be noted that the PUB data also showed that the chlorophyll a levels had dropped significantly over 2008~2012 and thus the above comparison could only not be done on a qualitatively basis.

The above observations are qualitatively consistent with the ELCOM-CAEDYM (ELCD) result as shown in Figure 7.7. The ELCD result showed less variable chlorophyll a concentration distribution. It is because at the beginning of the ELCD run, the chlorophyll a concentration was set as uniform in the reservoir and the ELCD run only for 7 days.

Figure 7.6 Sampling locations in Kranji reservoir.

Figure 7.7 Spatial variation of chlorophyll a concentration from PUB data: 3-month (Jun to Aug) average over years 2008~2012 and the ELCOM result averaged over the third period ; note no data is available for K1 during Jul-Aug 2010

7.3 A scenario test with the northerly wind

A simple hypothetical scenario was also run to investigate the behaviour of the phytoplankton under the northerly wind (winds from North) as typical during the Northeast Monsoon period so as to provide possible insights on the seasonal variation. For this ELCD input parameters such as meteorological data, inflow data and nutrients in the flow were kept the same as the 6-day July (21/07-27/07) 2012 ELCD run discussed above. The only change was to introduce a 180° rotation in the input wind direction to then yield northerly winds as shown in Figure 7.8.

Figure 7.8 The wind direction measured in the July 2012 field work vs the wind direction used for the ELCOM run in NE monsoon

The ELCD results are presented in Figure 7.9 for 24/07 and Figure 7.10 for the whole third period. Downwelling now occurred in the main channel. Thus the spatial temperature distribution is different from the calculation under the southerly wind: the horizontal temperature decreased from the junction towards the dam as shown in Figure 7.9b and Figure 7.10b. In addition, as the junction is the nutrient-rich area of the reservoir, the chlorophyll a concentration reduced along the channel from the junction towards the dam, though there was a small increase near the dam (over 0-1000 m) as seen in Figure 7.10c. This spatial variation was in qualitative agreement with the PUB sampling results during the NE Monsoon (3-month average over Dec-Feb, 2008-2012) as shown in Figure 7.11. The sampling results also showed a decrease from the south (St 1) towards the north (St 3 and St 4).

Figure 7.9 ELCOM-CAEDYM results averaged for the daytime (07:00-19:00) on 24/07 2012 under the northerly wind

Figure 7.10 ELCOM-CAEDYM results averaged over the daytimes (07:00-19:00) of the third period (21/07-26/07 2012) under the northerly wind

Figure 7.11 Spatial variation of chlorophyll a concentration: 3-month (Dec to Feb) average concentration in 2008~2012 and the ELCOM result averaged over the third period; note no data is available for K1 during Dec 2009-Feb 2010.

By comparing the chla concentration in Figure 7.5c and Figure 7.10c, it can be seen that the chla concentration was higher during the northeast monsoon than the southwest monsoon. As shown in Figure 7.10b, the downwelling area during the northeast monsoon was much larger than during the southwest monsoon in Figure 7.5b and it occurred in the nutrient-rich shallow junction area. The high temperature and the sufficient supply of the nutrient in a larger downwelling area might lead to the higher phytoplankton biomass during the Northeast Monsoon. However, it should be noted that this Northeast Monsoon scenario test was done with the meteorological data in the Southwest Monsoon; thus the air temperature used is likely higher than during the Northeast monsoon and favour higher chlorophyll a growth.

In summary, the major driving mechanisms of the spatial and temporal (seasonal) variations of the phytoplankton biomass were thus postulated via the ELCD modelling results as follows:

- a) during summer/ SW Monsoon, high phytoplankton biomass occurred near the dam due to the downwelling generated warm water and the surface transport towards the dam; the phytoplankton biomass was also high near the junction because the second weak circulation loop trapped the nutrients from the catchments near the junction area.
- b) during winter/ NE Monsoon, the nutrients from the catchments and the downwelling near the junction both favor the growth of phytoplankton near the junction area. The chlorophyll a concentration thus decreased from the junction towards the dam.

7.4 Summary

In this chapter the vertical distribution of phytoplankton biomass in terms of the chlorophyll a concentration was first investigated. It was found that the maximum chlorophyll a concentration gradient was consistent with the maximum temperature gradient: stratified during the daytime and moved downwards and became mixed over the water column with the SML deepening. The TKE budget analysis showed that convective cooling, the major forcing to the deepening, played an important role in shaping the vertical distribution of phytoplankton biomass. The computation of the vertical chlorophyll a flux demonstrated that the vertical transport of chlorophyll a was more active during the night and the vertical exchange of the phytoplankton biomass resulting from the underflow was stronger than within the SML. The underflow thus played an important role in the vertical mixing in Kranji reservoir.

The upwelling and downwelling as driven by the southerly wind (wind from south) established the daytime circulation pattern while the nighttime circulation arose from the differential cooling. The ELCD predicted flow pattern showed the presence of two separate circulation loops in the main channel, which met at K1. The small loop thus could circulate the nutrients from the catchments of Tengah, Pengsiang and Kangkar within the main channel over the region 2500~3500 m (distance from the dam along the reservoir channel), which would resulted in a high chlorophyll a concentration in that area. For the large circulation loop over the

range of 0~2500 m from the dam, high phytoplankton biomass occurred near the dam due to the downwelling generated warm water and the surface transport towards the dam.

The ELCD scenario carried out for the Northeast Monsoon where only the wind direction changed and the other factors remained the same predicted a decreasing chlorophyll a concentration along the channel from north towards south probably because of the downwelling at the junction. These ELCD results were in qualitative agreement with the trends seen in PUB chlorophyll a sampling results both under southerly (Southwest Monsoon) wind and northerly (Northeast Monsoon) wind seasons.

8. Conclusion and Future Work

8.1 Conclusion

This thesis has mainly addressed the vertical mixing and circulation patterns in a tropical shallow reservoir, that of Kranji reservoir. The work has investigated in detail the mechanisms driving vertical mixing in Kranji reservoir. Another main finding is the different circulation patterns as occurring in Kranji reservoir during daytimes and nighttimes for the southwest monsoon periods without the influence of inflow, this as deduced via field measurement and ELCOM numerical calculations. The spatial distribution of chlorophyll a concentration was further linked to the circulation patterns.

The major findings in each chapter are summarized below. Chapter 2 described a successful field data collection effort for Kranji Reservoir. The data comprised (i) short term and (ii) long term data. The short term data consisted of intensively sampled meteorological, temperature and velocity data along with microstructure SCAMP and CTD data, this via two short term (about 3 weeks in duration each) field deployment exercises in Sept 2011 and July 2012. The long-term (Aug 2011 to Dec 2012) data consisted of PUB rainfall data and data collected by SysEng for PUB at K1 and K2 stations and along with new thermister chain data in at K2 station and in the tributaries.

The collected data, particularly high precision meteorological and water temperature and velocity data, and SCAMP microstructure data over the July 2012 field deployment underpinned analysis which allowed an improved understanding of the thermal and flow structures in Kranji reservoir. In particular, the Surface Mixed Layer (SML) dynamics are elucidated based on SCAMP microstructure measurements and flow structures using both ADCP measurements and ELCOM modelling. This required high precision of intensively sampled field data and the importance of such data cannot be underestimated.

In Chapter 4, the SML analysis indicated that the SML depth is well defined by the maximum temperature gradient point while the depth where a 0.02 °C temperature

drop from the surface was a good indicator of surface mixing layer or the extent of large overturns. The diurnal variation of the SML thickness as well as its mixing mechanism was investigated. The result of the Fr_T - Re_T diagram indicates the dominant mixing forcing during the daytime is the billow-driven turbulence, which is consistent with the observation made from buoyancy frequency and the velocity shear stress. And it also shows that the dominant mixing forcing during the night time is the penetrative cooling, which is in agreement with the TKE budget analysis. Thus, the Fr_T - Re_T turbulent flow behaviour as delineated by Imberger and Ivey (1991) was found to apply here to explain the observed vertical structures as developing diurnally. A daytime thin SML (<2 m) such as under strong heating was very much affected by the wind shear generated billows resulting in a low mixing efficiency $\Gamma < 0.1$. The nighttime SML was dominant by penetrative cooling with a higher mixing efficiency. The deepening of SML extended to the water bottom on two SCAMP nights (11/07-12/07 and 25/07-26/07). Two thermoclines, i.e. diurnal thermocline and seasonal thermocline are known to exist in temperate shallow lakes (Read and Rose 2013) but only the diurnal thermocline exists in Kranji reservoir. More active mass exchanges over the water column in tropical shallow lakes are expected. Furthermore, the turbulent parameters observed in the SML including the mixing length, the buoyancy frequency and the vertical eddy diffusivities will find use in developing better estimates of the time scale of phytoplankton transport via mixing in the SML and the light history of the phytoplankton in tropical lakes and thus should contribute to the future development of more realistic biological models.

The temperature structure calculation based on the one-dimensional vertical transport equation showed that the thermal structure was determined by the vertical mixing and vertical heating during the first two 24-hr SCAMP measurement periods, although horizontal transport still affected the water column at K2, the deep part of the reservoir. For the last 24-hr SCAMP measurement period, the one-dimensional model gives a more stratified water column indicating that the vertical mass/heat distribution for the observed strong three-layer temperature structure was not only dependent on the vertical mixing but also on the horizontal dispersion/advection .

The results of Chapter 5 demonstrated that ELCOM results are qualitatively in agreement with the temperature and velocity field in the reservoir and qualitatively captured the important observed features of the thermal and flow structure. The agreement however deteriorated in the lower water column probably because of the less efficient transport of momentum and heat in ELCOM. In addition, the weakened horizontal temperature gradient in the model generated a smaller velocity field during the night.

In Chapter 6, the field measurement of the thermal and flow structure have been investigated in details. Both temperature and velocity measurement in the reservoir showed a daytime three-layer structure and a nighttime two-layer structure during the third week of the July 2012 field work when no inflow or rainfall occurred. The velocity at K1 showed a two-layer structure during the daytime and a uniform velocity profile with its direction changing during the nighttime. In addition, the temperature difference between the tributaries and the main water body suggested that a convective circulation was probably set in motion via the differential cooling/heating. The inflow may enhance or destruct the vertical stratification as dependent on the temperature and the volume of the inflow into the reservoir. In addition, the gate discharge operations could set up a strong horizontal motion and horizontal mixing.

The circulation pattern in Kranji reservoir as deduced from ELCOM results was driven by upwelling and the downwelling due to the southerly wind (wind from south) together with the horizontal temperature gradient during daytime, and by differential cooling during nighttime. During the daytime heating period, the circulation was dominated by a big loop along the main channel with the current towards the junction in the central of the channel and towards the dam at both sides of the channel. During the nighttime, there existed two loops in the main channel, a big one located in the range from 1000m to 2500 m from the dam and a small loop in the range from 2500m to 3500 m from the dam. Within the loops, the current in the centre was towards the dam and the currents at the two sides were towards the junction.

In the Chapter 7, the spatial distribution of chlorophyll a concentration was linked to the vertical mixing and the circulation pattern discussed in the previous chapters. In terms of the vertical distribution of phytoplankton biomass, it is found that the maximum chlorophyll a concentration gradient was consistent with the maximum temperature gradient and moved downwards with the SML deepening. The convective cooling, the major forcing to the deepening, thus played an important role in shaping the vertical distribution of phytoplankton biomass. The computation of the vertical chlorophyll a flux demonstrated that the vertical exchange of the phytoplankton biomass resulting from the underflow was not small and indicated nutrients from the sediment could be re-introduced into the water column via mixing from the underflow. The role of the underflow in the Kranji ecosystem deserves further investigation.

The ELCOM-CAEDYM results presented in this work is a good example of how hydrodynamics influences on the phytoplankton distribution in tropical lakes. The high chlorophyll a in the main channel of Kranji Reservoir near the tributary junction could be attributed to the small nighttime circulation over the distance (range) ~2500 to 3500m from the dam along the main reservoir channel. This flow pattern could circulate nutrients from the catchments of Tengah, Pengsiang and Kangkar within this small loop. For the large nighttime loop occurring over 0~2500 m from the dam, high phytoplankton biomass occurred near the dam due to downwelling generated warm water and the surface transport towards the dam.

The seasonal variation of chlorophyll a was explored via a further ELCOM-CAEDYM (ELCD) scenario for the Northeast Monsoon to complement the results for July 2012 which was during the Southwest Monsoon period. The spatial chlorophyll a distribution during Northeast Monsoon exhibited a decreasing chlorophyll a concentration along the channel from south to north. This was attributed to warmer temperature and sufficient nutrient supply at the downwelling side near the junction under the northerly wind during the Northeast Monsoon. Lastly, these trends as derived from ELCD results were in qualitative agreement with the PUB chlorophyll a sampling results over Southwest and Northeast Monsoon periods.

From this work, it can be seen that the dynamics of tropical shallow lakes is fragile and highly variable because of the lack of distinct seasonal forcing and the shallowness leads to a more significant diurnal variation than any seasonal variation. As a result, weak changes in physical forcing can be significant to the lake systems e.g. a 0.6 °C temperature difference induced underflow which induce active vertical mass transport over the water column or the thermal structure could behave differently within several days under the alternative influences of the cold inflow and the solar radiation.

Furthermore, this work gives an example of high level physical-biological coupling in tropical lakes. It has been shown that the distribution of the chlorophyll a concentration in Kranji reservoir was consistent with by the flow pattern as driven by the monsoonal wind. Such responses of the tropical shallow lakes to the monsoon seasons deserve more investigation and it may help the understanding of the climate changes on the ecosystem in tropical regions.

Finally, the findings in this study should contribute to a better understanding and management of the reservoir. The SML behaviour indicates that daytime surface concentration of chlorophyll a concentration will be mixed vertically by nighttime penetrative cooling. Therefore, it should not be necessary to artificially enhance vertical mixing via e.g. aeration devices, which then reduces the operation cost of the reservoir. The inferred mean flow circulation patterns and implied concentration of nutrients may also guide location of remedial actions e.g. floating wetland to remove excessive nutrient from the water system.

8.2 Suggestions for future work

The analysis of vertical mixing indicated that the bottom underflow as driven by the horizontal temperature gradient was important to the mass transport between the sediment and the water column. The role of the underflow and the vertical transport in the lower water column deserves more investigation.

In this work, predicted thermal structure and the flow structure by ELCOM matched well with the temperature and velocity data when there was no inflow into the

reservoir. The model did not work well when the effects of the inflows became important due to lack of good inflow data to the ELCOM model. The performance of the model can thus be improved by providing more accurate inflow data. The ELCOM results also implied that the flow patterns during the northeast monsoon could be very different from those in the southwest monsoon. These flow patterns need to be further verified by more extensive field measurements over longer durations and different monsoonal periods.

As noted in Chapter 7 the chlorophyll a concentration was related to the circulation pattern and the vertical mixing. In the CAEDYM model, the inputs of the nutrient concentration from the catchments were based on earlier measurements in 2006 as well as fortnightly PUB sampling events, which could result in large uncertainties in the calculation. High quality nutrient concentrations not only in the main water body but also in the catchments are required for a better understanding of the interaction between the circulation pattern and nutrient distribution and for a more precise estimate of chlorophyll a concentration in the reservoir. Lastly, the scenario test in the NE monsoon was carried out based on the data collected in July 2012. Field measurements and intensively sampled water quality data during the NE monsoon are needed to definitively verify the accuracy of the predictions.

Reference

- Anis, A. and J. Moum (1992). "The superadiabatic surface layer of the ocean during convection." Journal of Physical Oceanography **22**(10): 1221-1227
- Alford, M. H. and R. Pinkel (2000). "Observations of overturning in the thermocline: The context of ocean mixing." Journal of Physical Oceanography **30**(5): 805-832.
- Appt, J., J. Imberger and H. Kobus (2004). "Basin-scale motion in stratified Upper Lake Constance." Limnology and Oceanography **49**(4): 919-933.
- Batchelor, G. (1959). "Small-scale variation of convected quantities like temperature in turbulent fluid Part 1. General discussion and the case of small conductivity." Journal of Fluid Mechanics **5**(01): 113-133.
- Boegman, L., et al. (2005). "The degeneration of internal waves in lakes with sloping topography." Limnology and Oceanography **50**(5): 1620-1637.
- Boegman, L., et al. (2005). "The energetics of large-scale internal wave degeneration in lakes." Journal of Fluid Mechanics **531**: 159-180.
- Brainerd, K. E. and M. C. Gregg (1995). "Surface mixed and mixing layer depths." Deep Sea Research Part I: Oceanographic Research Papers **42**(9): 1521-1543.
- Caughey, S. J. and S. G. Palmer (1979). "Some aspects of turbulence structure through the depth of the convective boundary layer." Quarterly Journal of the Royal Meteorological Society **105**(446): 811-827.
- Copetti, D., G. Tartari, G. Morabito, A. Oggioni, E. Legnani and J. Imerger (2006). "A biogeochemical model of Lake Pusiano (North Italy) and its use in the predictability of phytoplankton blooms: first preliminary results." Journal of Limnology **65**(1): 59-64.
- Cushman-Roisin, B., A. J. Willmott and N. R. Biggs (2005). "Influence of stratification on decaying surface seiche modes." Continental shelf research **25**(2): 227-242.
- Dillon, T. (1982). "Vertical overturns: A comparison of Thorpe and Ozmidov length scales." Journal of Geophysical Research: Oceans (1978–2012) **87**(C12): 9601-9613.
- Dunkley, J., J. Koseff, J. Steinbuck, S. Monismith and A. Genin (2012). "Comparison of mixing efficiency and vertical diffusivity models from temperature microstructure." Journal of Geophysical Research: Oceans (1978–2012) **117**(C10).

- Edwards, K. A., P. MacCready, J. N. Moum, G. Pawlak, J. M. Klymak and A. Perlin (2004). "Form Drag and Mixing Due to Tidal Flow past a Sharp Point." Journal of Physical Oceanography **34**(6): 1297-1312.
- Efron, B. and G. Gong (1983). "A leisurely look at the bootstrap, the jackknife, and cross-validation." The American Statistician **37**(1): 36-48.
- Farrow, D. and J. Patterson (1993). "On the response of a reservoir sidearm to diurnal heating and cooling." Journal of Fluid Mechanics **246**: 143-161.
- Farrow, D. E. (2004). "Periodically forced natural convection over slowly varying topography." Journal of Fluid Mechanics **508**: 1-21.
- Fischer, H. B. (1979). Mixing in inland and coastal waters, Academic press.
- Fish, G. R. (1957). A seiche movement and its effect on the hydrology of Lake Victoria, HM Stationery Office.
- Galbraith, P. S. and D. E. Kelley (1996). "Identifying Overturns in CTD Profiles." Journal of Atmospheric and Oceanic Technology **13**(3): 688-702.
- Geyer, W. (1997). "Influence of wind on dynamics and flushing of shallow estuaries." Estuarine, Coastal and Shelf Science **44**(6): 713-722.
- Gibson, C. and W. Schwarz (1963). "The universal equilibrium spectra of turbulent velocity and scalar fields." Journal of Fluid Mechanics **16**(03): 365-384.
- Griffin, S. L., M. Herzfeld and D. P. Hamilton (2001). "Modelling the impact of zooplankton grazing on phytoplankton biomass during a dinoflagellate bloom in the Swan River Estuary, Western Australia." Ecological Engineering **16**(3): 373-394.
- Hicks, B. (1972). "Some evaluations of drag and bulk transfer coefficients over water bodies of different sizes." Boundary-layer meteorology **3**(2): 201-213.
- Hipsey, M., J. Romero, J. Antenucci and D. Hamilton (2006). "Computational aquatic ecosystem dynamics model: CAEDYM v2." Contract Research Group, Centre for Water Research, University of Western Australia: 90.
- Hodges, B. and C. Dallimore (2006). "Estuary, Lake and Coastal Ocean Model: ELCOM." Science Manual. Centre of Water Research. University of Western Australia.
- Hodges, B. R., J. Imberger, A. Saggio and K. B. Winters (2000). "Modeling basin-scale internal waves in a stratified lake." Limnology and Oceanography **45**(7): 1603-1620.

- Hodges, B. R., B. Laval and B. M. Wadzuk (2006). "Numerical error assessment and a temporal horizon for internal waves in a hydrostatic model." Ocean Modelling **13**(1): 44-64.
- Horn, D., et al. (2001). "The degeneration of large-scale interfacial gravity waves in lakes." Journal of Fluid Mechanics **434**: 181-207.
- Hughes, G. O. and R. W. Griffiths (2008). "Horizontal convection." Annu. Rev. Fluid Mech. **40**(1): 185-208.
- Huisman, J., P. v. Oostveen and F. J. Weissing (1999). "Critical Depth and Critical Turbulence: Two Different Mechanisms for the Development of Phytoplankton Blooms." Limnology and Oceanography **44**(7): 1781-1787.
- Hutter, K., Y. Wang and I. P. Chubarenko (2011). Physics of Lakes, Springer Berlin Heidelberg.
- Imberger, J. (1985). "The diurnal mixed layer." Limnol. Oceanogr **30**(4): 737-770.
- Imberger, J. (1987). "Mixing in River Underflow." Journal of Hydraulic Engineering **113**(6): 697-715.
- Imberger, J. (1998). Physical processes in lakes and oceans, American Geophysical Union.
- Imberger, J. and P. Hamblin (1982). "Dynamics of lakes, reservoirs, and cooling ponds." Annual Review of Fluid Mechanics **14**(1): 153-187.
- Imberger, J. and G. Ivey (1991). "On the nature of turbulence in a stratified fluid. Part II: Application to lakes." Journal of Physical Oceanography **21**(5): 659-680.
- Imberger, J. and J. C. Patterson (1989). "Physical limnology." Advances in applied mechanics **27**: 303-475.
- Imboden, D. M. and A. Wüest (1995). Mixing mechanisms in lakes. Physics and chemistry of lakes, Springer: 83-138.
- Imerito, A. (2007). Dynamic reservoir simulation model. DYRESM v 4.0 science manual, University of Western Australia.
- Ivey, G., P. De Silva and J. Imberger (1995). Internal waves, bottom slopes and boundary mixing. Aha Hulikoa, Hawaiian Winter Workshop on Topographic Effects in the Ocean. Honolulu, Hawaii (ed. P. Muller & D. Henderson), DTIC Document.
- Ivey, G. and J. Imberger (1991). "On the nature of turbulence in a stratified fluid. Part I: The energetics of mixing." Journal of Physical Oceanography **21**(5): 650-658.

- Ivey, G. and R. Nokes (1989). "Vertical mixing due to the breaking of critical internal waves on sloping boundaries." Journal of Fluid Mechanics **204**: 479-500.
- Ivey, G. N., K. B. Winters and J. R. Koseff (2008). "Density Stratification, Turbulence, but How Much Mixing?" Annual Review of Fluid Mechanics **40**(1): 169-184.
- Jonas, T., A. Stips, W. Eugster and A. Wüest (2003). "Observations of a quasi shear-free lacustrine convective boundary layer: Stratification and its implications on turbulence." Journal of Geophysical Research: Oceans **108**(C10): 3328.
- Kannan, V. and S. V. Job (1980). "Diurnal depth-wise and seasonal changes of physico-chemical factors in Sathiar reservoir." Hydrobiologia **70**(1-2): 103-117.
- Laval, B., J. Imberger and A. Findikakis (2005). "Dynamics of a large tropical lake: Lake Maracaibo." Aquatic Sciences **67**(3): 337-349.
- Laval, B., J. Imberger, B. R. Hodges and R. Stocker (2003). "Modeling circulation in lakes: Spatial and temporal variations." Limnology and Oceanography **48**(3): 983-994.
- Leon, L. F., D. Lam, W. Schertzer and D. Swayne (2005). "Lake and climate models linkage: a 3-D hydrodynamic contribution." Adv. Geosci. **4**: 57-62.
- Lewis, W. M. (1996). Tropical lakes: how latitude makes a difference. Perspectives in tropical Limnology. F. B. Schiemer, K T. Amsterdam, The Netherlands, SPB Academic Publishing 43-64.
- Lewis, W. M. (2000). "Basis for the protection and management of tropical lakes." Lakes & Reservoirs: Research & Management **5**(1): 35-48.
- Lo, E. Y. M. (2008). Water quality monitoring, modeling and management for Kranji catchment/reservoir system. Singapore, Public Utilities Board.
- Lombardo, C. and M. Gregg (1989). "Similarity scaling of viscous and thermal dissipation in a convecting surface boundary layer." Journal of Geophysical Research: Oceans (1978–2012) **94**(C5): 6273-6284.
- Luketina, D. A. and J. Imberger (1989). "Turbulence and entrainment in a buoyant surface plume." Journal of Geophysical Research: Oceans (1978–2012) **94**(C9): 12619-12636.
- MacIntyre, S. and J. M. Melack (1982). "Meromixis in an equatorial African soda lake'." LIMNOLOGY **27**.
- MacIntyre, S. (1993). "Vertical Mixing in a Shallow, Eutrophic Lake: Possible Consequences for the Light Climate of Phytoplankton." Limnology and Oceanography **38**(4): 798-817.

- MacIntyre, S., K. M. Flynn, R. Jellison and J. R. Romero (1999). "Boundary mixing and nutrient fluxes in Mono Lake, California." Limnology and Oceanography **44**(3): 512-529.
- MacIntyre, S., J. R. Romero and G. W. Kling (2002). "Spatial-temporal variability in surface layer deepening and lateral advection in an embayment of Lake Victoria, East Africa." Limnology and Oceanography **47**(3): 656-671.
- MacIntyre, S., J. F. Clark., R. Jellison and J. P. Fram (2009). "Turbulent mixing induced by nonlinear internal waves in Mono Lake, California." Limnology and Oceanography **54**(6): 2255-2272.
- MacIntyre, S. (2012). Stratification and Mixing in Tropical African Lakes. Encyclopedia of Lakes and Reservoirs. L. Bengtsson, R. W. Herschy and R. W. Fairbridge. Berlin, Springer Verlag. **1**: 737-743.
- MacIntyre, S., J. R. Romero, G. M. Silsbe and B. M. Emery (2014). "Stratification and horizontal exchange in Lake Victoria, East Africa." Limnology and Oceanography **59**(6): 1805-1838.
- Mans, C., S. Bramato, A. Baquerizo and M. A. Losada (2011). "Surface seiche formation on a shallow reservoir in complex terrain." Journal of Hydraulic Engineering **137**(5): 517-529.
- Mao, Y., C. Lei and J. C. Patterson (2009). "Unsteady natural convection in a triangular enclosure induced by absorption of radiation—a revisit by improved scaling analysis." Journal of Fluid Mechanics **622**: 75-102.
- Mao, Y., C. Lei and J. C. Patterson (2010). "Unsteady near-shore natural convection induced by surface cooling." Journal of Fluid Mechanics **642**: 213-233.
- Merian, J. (1828). "Über die Bewegung Tropbarer Flüssigkeiten in Gefassen." Basel, Switzerland.
- Monismith, S. G., A. Genin, M. A. Reidenbach, G. Yahel and J. R. Koseff (2006). "Thermally Driven Exchanges between a Coral Reef and the Adjoining Ocean." Journal of Physical Oceanography **36**(7).
- Monismith, S. G., J. Imberger and M. L. Morison (1990). "Convective motions in the sidearm of a small reservoir." Limnol. Oceanogr **35**(8): 1676-1702.
- Monismith, S. G. and S. MacIntyre (2009). The Surface Mixed Layer in Lakes and Reservoirs. Encyclopedia of Inland Waters. G. E. Likens. Oxford, Academic Press: 636-650.
- Moum, J. N. (1996a). "Efficiency of mixing in the main thermocline." Journal of Geophysical Research: Oceans **101**(C5): 12057-12069.
- Moum, J. N. (1996b). "Energy-containing scales of turbulence in the ocean thermocline." Journal of Geophysical Research: Oceans **101**(C6): 14095-14109.

- Munnich, M. (1996). "The influence of bottom topography on internal seiches in stratified media." Dynamics of atmospheres and oceans **23**(1): 257-266.
- Munnich, M., A. Wuest and D. Imboden (1992). "Observations of the second vertical mode of the internal seiche in an alpine lake." Limnol. Oceanogr **37**(8): 1705-1719.
- Niemann, H., C. Richter, H. M. Jonkers and M. I. Badran (2004). "Red Sea gravity currents cascade near-reef phytoplankton to the twilight zone." Marine Ecology Progress Series **269**: 91-99.
- Osborn, T. R. (1980). "Estimates of the Local Rate of Vertical Diffusion from Dissipation Measurements." Journal of Physical Oceanography **10**(1): 83-89.
- Osborn, T. R. and C. S. Cox (1972). "Oceanic fine structure." Geophysical & Astrophysical Fluid Dynamics **3**(1): 321-345.
- Pawlowicz, R., B. Beardsley, S. Lentz, E. Dever & A. Anis (2001) Software simplifies air-sea data estimates. Eos, Transactions American Geophysical Union, **82**, 2-2.
- Pernica, P., M. Wells & S. MacIntyre (2014) Persistent weak thermal stratification inhibits mixing in the epilimnion of north-temperate Lake Opeongo, Canada. Aquatic Sciences, **76**, 187-201.
- Peters, H. (1997). "Observations of stratified turbulent mixing in an estuary: Neap-to-spring variations during high river flow." Estuarine, Coastal and Shelf Science **45**(1): 69-88.
- Piera, J., E. Roget and J. Catalan (2002). "Turbulent patch identification in microstructure profiles: A method based on wavelet denoising and Thorpe displacement analysis." Journal of Atmospheric & Oceanic Technology **19**(9).
- Read, J. S. and K. C. Rose (2013). "Physical responses of small temperate lakes to variation in dissolved organic carbon concentrations." Limnology and Oceanography **58**(3): 921-931.
- Romero, J. R., J. P. Antenucci and J. Imberger (2004). "One- and three-dimensional biogeochemical simulations of two differing reservoirs." Ecological Modelling **174**(1-2): 143-160.
- Ruddick, B., A. Anis and K. Thompson (2000) Maximum likelihood spectral fitting: The Batchelor spectrum. Journal of Atmospheric and Oceanic Technology, **17**(11), 1541-1555.
- Rueda, F. J. and S. G. Schladow (2002). "Surface seiches in lakes of complex geometry." Limnology and Oceanography **47**(3): 906-910.

Saggio, A. and J. Imberger (1998) Internal wave weather in a stratified lake. Limnology and oceanography **43**(8), 1780-1795.

Shay, T. J. and M. Gregg (1986). "Convectively driven turbulent mixing in the upper ocean." Journal of Physical Oceanography **16**(11): 1777-1798.

Sherman, F. S., J. Imberger and G. Corcos (1978). "Turbulence and mixing in stably stratified waters." Annual Review of Fluid Mechanics **10**(1): 267-288.

Spigel, R. H. and J. Imberger (1980). "The classification of mixed-layer dynamics of lakes of small to medium size." Journal of Physical Oceanography **10**(7): 1104-1121.

Stevens, C., G. Lawrence, P. Hamblin and E. Carmack (1996) Wind forcing of internal waves in a long narrow stratified lake. Dynamics of Atmospheres and Oceans **24**(1-4), 41-50.

Talling, J. (1957) Diurnal changes of stratification and photosynthesis in some tropical African waters. Proceedings of the Royal Society of London. Series B-Biological Sciences, **147**(926), 57-83.

Talling, J. F. (1966) The Annual Cycle of Stratification and Phytoplankton Growth in Lake Victoria (East Africa). Internationale Revue der gesamten Hydrobiologie und Hydrographie, **51**(4), 545-621.

Talling, J. F. and J. Lemoalle. (1998). Ecological Dynamics of Tropical Inland Waters. Cambridge Univ. Press.

Talling, J. F. and J. Lemoalle (1999). "Ecological Dynamics of Tropical Inland Waters." Journal of Ecology **87**(4): 733-734.

Te, S. H. and K. Y.-H. Gin (2011). "The dynamics of cyanobacteria and microcystin production in a tropical reservoir of Singapore." Harmful Algae **10**(3): 319-329.

Tedford, E. W., S. MacIntyre, S. D. Miller and M. J. Czikowsky (2014) Similarity scaling of turbulence in a temperate lake during fall cooling. Journal of Geophysical Research: Oceans, **119**(8), 4689-4713.

Thompson, R. and J. Imberger (1980). Response of a numerical model of a stratified lake to wind stress. Second International Symposium on Stratified Flows, IAHR, Trondheim, Norway.

Thorpe, S. A., A. Hall and I. Crofts (1972) The Internal Surge in Loch Ness. Nature, **237**(5350), 96-98.

Thorpe, S. (1973). "Turbulence in stably stratified fluids: A review of laboratory experiments." Boundary-layer meteorology **5**(1-2): 95-119.

Vidal, J. and X. Casamitjana (2008). "Forced resonant oscillations as a response to periodic winds in a stratified reservoir." Journal of Hydraulic Engineering **134**(4): 416-425.

Vidal, J., X. Casamitjana, J. Colomer and T. Serra (2005). "The internal wave field in Sau reservoir: Observation and modeling of a third vertical mode." Limnology and Oceanography **50**(4): 1326-1333.

Vidal, J., F. J. Rueda and X. Casamitjana (2007). "The seasonal evolution of high vertical-mode internal waves in a deep reservoir." Limnology and Oceanography **52**(6): 2656.

Wong, K. T. M., J. H. W. Lee and I. J. Hodgkiss (2007). "A simple model for forecast of coastal algal blooms." Estuarine, Coastal and Shelf Science **74**(1-2): 175-196.

Wüest, A. and A. Lorke (2003). "Small-scale hydrodynamics in lakes." Annual Review of Fluid Mechanics **35**(1): 373-412.

Xing, Z. (2007). Hydrodynamics of a tropical reservoir and effects on eutrophication, Department of Civil and Environmental Engineering, Nanyang Technological University.

Xing, Z. (2011). Thermal Structure and Hydrodynamics of A Shallow Tropical Reservoir. School of Civil and Environmental Engineering. Singapore, Nanyang Technological University. **Doctor of Philosophy**: 160.

Xing, Z., D. A. Fong, E. Y.-M. Lo and S. G. Monismith (2014). "Thermal structure and variability of a shallow tropical reservoir." Limnol. Oceanogr **59**(1): 115-128.

Xing, Z., D. A. Fong, K. M. Tan, E. Y. M. Lo and S. G. Monismith (2012). "Water and heat budgets of a shallow tropical reservoir." Water Resources Research **48**(6).