Conceptual models of dissolved carbon fluxes considering interannual typhoon responses under extreme climates in a two-layer stratified lake

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Abstract

Extreme climates affect the seasonal and interannual patterns of carbon (C) distribution due to the regimes of river inflow and thermal stratification within lentic ecosystems. Typhoons rapidly load substantial amounts of terrestrial C into small subtropical lakes (i.e., Yuan-Yang Lake, YYL), renewing and mixing the water column. We developed conceptual dissolved C model and hypothesized that allochthonous C loading and river inflow intrusion may affect the dissolved inorganic C (DIC) and dissolved organic C (DOC) distributions in a small subtropical lake under these extreme climates. A two-layer conceptual C model was developed to explore how the DIC and DOC fluxes respond to typhoon disturbances on seasonal and interannual time scales in YYL while simultaneously considering autochthonous processes such as algal photosynthesis, remineralization, and vertical transportation. Monthly field samplings were conducted to measure the DIC, DOC, and chlorophyll a concentrations to compare the temporal patterns of fluxes between typhoon years (2015–2016) and non-typhoon years (2017–2018). The results demonstrated that net ecosystem production was 3.14 times higher in the typhoon years than in the non-typhoon years in YYL. The results suggested that a load of allochthonous C was the most crucial factor affecting the temporal variation of C fluxes in the typhoon years because of changes in the physical and biochemical processes, such as photosynthesis, mineralization, and vertical transportation. However, the lowered vertical transportation rate shaped the seasonal C in the non-typhoon years due to thermal stratification within this small subtropical lake.
1. Introduction

The Intergovernmental Panel for Environmental Changes Sixth Assessment Report (IPCC AR6) (2021) suggested that, by 2050, not only is air temperature going to increase by at least about 1.5°C but high-intensity storms and drought events will become more frequent as a result of global warming and climate change. In freshwater ecosystems, extreme climates may change the mixing regimes of water columns (Kraemer et al., 2021; Maberly et al., 2020; Woolway et al., 2020), heat wave events (Woolway et al., 2021a; Woolway et al., 2021b), droughts (Marcé et al., 2019), and floods (Woolway et al., 2018). Freshwater ecosystems store inorganic C of CO within Dissolved of Woolway et al., 2018) on the fate of C fluxes in small stratified lakes (Aufdenkampe et al., 2011; Cole et al., 2007; Raymond et al., 2013; Tranvik et al., 2009; Vachon et al., 2021; Woolway et al., 2020), heat wave events (Woolway et al., 2021a; Woolway et al., 2021b), droughts (Marcé et al., 2019), and floods (Woolway et al., 2018). Freshwater ecosystems store around 0.32 to 1.8 Pg C yr⁻¹, which is approximately equivalent to shallow coastal areas; these ecosystems provide important services for human sustainability, such as acting as processing hotspots in regional carbon (C) cycling (Aufdenkampe et al., 2011; Cole et al., 2007; Engel et al., 2018; Lauerwald et al., 2015; Raymond et al., 2013). Extreme weather events might induce stronger seasonal thermal stratification from spring to summer and longer overturns from autumn to winter, thereby changing C distribution and transportation within water bodies (Kraemer et al., 2021; Olsson et al., 2022a; Woolway et al., 2020). The responses of C fluxes in small lakes (lake area < 1 km²) are sensitive to climate change due to the increase with which C mixes with water columns (Doubek et al., 2021; MacIntyre et al., 2021; Winslow et al., 2015). Moreover, storms induce dramatic changes in thermal stratification and water inflows (Lin et al., 2022; Olsson et al., 2022b; Vachon and Del Giorgio, 2014; Woolway et al., 2018). River inflows and wind turbulence mix allochthonous C from sediments into the water column after storm events in small stratified lakes (Bartosiewicz et al., 2015; Czikowsky et al., 2018; Vachon and Del Giorgio, 2014). However, small lakes account for 25% to 35% of the total area of the earth's surface lakes (Cole et al., 2007; Downing et al., 2006; Raymond et al., 2013). Compared to the case in larger lakes, C fluxes in small lakes remain uncertain because small lakes have usually been ignored in calculations of C flux on a global scale (Cole et al., 2007; Raymond et al., 2013). Thus, elucidation of the C fluxes in small lakes in extreme climates is key to optimizing estimations of global C fluxes in extreme climates.

Understanding the influences of physical, hydrological, and biogeochemical processes on the fate of C fluxes in smaller lake ecosystems is challenging work (Aufdenkampe et al., 2011; Cole et al., 2007; Raymond et al., 2013; Tranvik et al., 2009; Vachon et al., 2021; Woolway et al., 2018). This is not only because of difficulties in measurement but also because of the dynamics and interactions between factors and processes associated with C fluxes. Dissolved inorganic carbon (DIC) concentration is an important factor in estimating CO₂ fluxes within lake ecosystems (Smith, 1985). Among C fluxes in a freshwater body, the partial pressure of CO₂ (pCO₂), defined as CO₂ emission across the air–water interface, is affected by dissolved inorganic C (DIC), water temperature, wind speed, and pH (Jähne et al., 1987; Smith, 1985).
River inflows, sediment, and respiration contribute to DIC loading into lakes (Hope et al., 2004; Vachon et al., 2021); simultaneously, autotrophic organisms, such as planktons and submerged vegetation, capture DIC via photosynthesis (Amaral et al., 2022; Nakayama et al., 2020; Nakayama et al., 2022). Moreover, calcification and mineralization may consume dissolved oxygen within water, inducing uncertainty in pCO₂ estimation (Hanson et al., 2015; Lin et al., 2022; Nakayama et al., 2022). Dissolved organic carbon (DOC) might contribute to CO₂ emission from lake water to the atmosphere through mineralization and remineralization within lake ecosystems (Hanson et al., 2015; Sobek et al., 2005). In subtropical freshwater ecosystems, DOC concentration is a vital factor in describing variances in mineralization and remineralization rates for dissolved C (Lin et al., 2022; Shih et al., 2019). Kossin et al. (2013) investigated global storm events with an accumulated rainfall of about 50 mm, which is approximately 10 %–40% of precipitation in a subtropical typhoon event. Other studies have found that typhoon disturbances quickly mix, renew, or dilute the water in small subtropical lakes (Kimura et al., 2012; Kimura et al., 2017; Lin et al., 2022). Therefore, investigating the magnitudes of DIC and DOC during typhoon disturbances is understanding the seasonal regimes and to estimating C fluxes in small subtropical lakes.

Typhoons’ effects on C fluxes were previously studied in a small, two-layer stratified, subtropical lake, Yuan–Yang Lake (YYL) (Chiu et al., 2020; Jones et al., 2009; Lin et al., 2021; Lin et al., 2022). Jones et al. (2009) used the conceptual hydrology model and sensor data to estimate CO₂ emission in YYL during typhoon disturbances that occurred in October 2004: 2.2 to 2.7 g C m⁻² d⁻¹ of CO₂ was released into the atmosphere. CO₂ emissions into the atmosphere were recorded at around 3.0 to 3.7 g C m⁻² d⁻¹ because of substantial loads of terrestrial C via river inflows after strong typhoons in YYL (Chiu et al., 2020). In particular, vertical mixing, thermal stratification, and river retention regimes were found to be essential physical processes in the C fluxes in YYL (Lin et al., 2021; Lin et al., 2022). These studies suggest that river intrusion and thermal stratification are key factors shaping the seasonal and interannual patterns of C fluxes during typhoon disturbances. River intrusion not only controls the C fluxes, algal biomass, and nutrient loading, but also influences the length of stratification and hydraulic retention times (Lin et al., 2021; Lin et al., 2022; Maranger et al., 2018; Nakayama et al., 2020; Olsson et al., 2022a; Olsson et al., 2022b; Zwart et al., 2017; Vachon and Del Giorgio, 2014). We hypothesized that allochthonous C loading and river inflow intrusion may affect DIC and DOC distributions (Fig. 1). Further, allochthonous processes in small subtropical lakes, such as algal photosynthesis, remineralization, and vertical transportation, must also be considered (Fig. 1). Here, we tested our hypothesis developing two-layer conceptual C models to assess C flux responses to typhoon disturbances in small subtropical lakes.
2. Materials and methods

2.1 Study site

YYL is a shallow (mean water depth: 4.3 m) and oligotrophic (total phosphorous: 10-20 μg-P L⁻¹; total nitrogen: 20-60 μg-N L⁻¹) subtropical mountain lake (Chou et al., 2000; Tsai et al., 2008; Wu et al., 2001) on Chi-Lan Mountain at around 1,640 asl in north-central Taiwan (24.58° N, 121.40° E) (Figure 2). Its water is brown because of its humic acid content (colored assessed by a portable fluorometer (model C3; Turner Designs, Sunnyvale, CA, USA); mean pH: 5.4). YYL is surrounded by old-growth trees such as Chamaecyparis formosensis, Chamaecyparis obtusa var. formosana, and Rhododendron formosanum Heiml (Chou et al., 2000). Annual precipitation is over 3,000 mm yr⁻¹, and typhoon precipitation contributes up to half of the total precipitation in YYL (Chang et al., 2007; Lai et al., 2006). Due to the rapid renewal of the water body, the water retention time (or residence time) was around 4.4 days in typhoon Megi from 27 September to 1 October 2016 (Lin et al., 2022). The water surface temperature ranges from 15 to 25 °C during March to August, and the water column overturns in September (Kimura et al., 2012; Kimura et al., 2017; Lin et al., 2021). The concentrations of dissolved C (Lin et al., 2021), nutrients (Chiu et al., 2020; Tsai et al., 2008) and organisms (Shade et al., 2011) increase within YYL from autumn to winter. YYL has been registered as a long-term ecological study site by the Ministry of Science and Technology (MOST) of Taiwan since 1992 and became part of the Global Lake Ecological Observatory Network (GLEON) in 2004.

2.2 Water sampling and chemical analysis

We collected water quality samples (DOC, DIC, and Chl. a) at water depths of 0.04, 0.50, 1.00, 2.00, and 3.50 m at the buoy site (Fig. 2). We also measured the water surfaces of six river inflows and an outflow monthly using a horizontal van Dorn bottle (2.20 L, acrylic) from January 2015 to December 2018 (Fig. 2). These samples were collected using a portable hand pump and glass microfiber filter papers (47 mm GF/F, nominal pore size 0.70 μm; Whatman, Maidstone, Kent, UK) to obtain filtrate samples. Water samples were stored at around 4°C in a refrigerator until analysis. Samples were analyzed using an infrared gas detector to detect DIC and DOC concentrations with persulfate digestion (model 1088 Rotary TOC autosampler; OI Analytical, College Station, TX, USA). The filter papers were kept in opaque bottles at around -25 °C in a refrigerator until the samples were analyzed using a portable fluorometer (model 10-AU-005-CE; Turner Designs, Sunnyvale, CA, USA). The specific wavelengths were 430 nm (blue) and 662 nm (red). In the laboratory, the filter papers were extracted with methanol to
obtain Chl. a concentration. These samples were analyzed for less than 72 h to prevent light and
chemical degradation.

2.3 Data analysis and numerical modeling

Three water quality variables (DIC, DOC, and Chl a) were compared between different
layers (upper and lower layers), years (typhoon years and non-typhoon years), and seasons (spring,
summer, autumn, and winter). First, we separated our investigation data into typhoon years and
non-typhoon years as described in Sect. 2.3.1. Next, we developed a conceptual equations model
to generate continuous DIC and DOC data at the upper and lower layers as shown in Figure 1.
This helped us understand the transportation, photosynthesis, and remineralization rates between
seasons and between typhoon and non-typhoon years (see Sect. 2.3.2).

2.3.1 Typhoon and non-typhoon years

We collected meteorological data from a meteorological tower located about 1.0 km
from YYL (Lin et al., 2021; Lin et al., 2022). Rainfall (model N-68; Nippon Electric Instrument,
Tokyo, Japan) and wind speed (model 03001, R.M. Young, Traverse City, MI, USA) data were
stored in a datalogger (model CR1000; Campbell Scientific, Logan, UT, USA) for every 10 min.
River discharge ($Q_m$, m$^3$ d$^{-1}$) was estimated every 10 min using the rainfall data and a water depth
meter (model HOBO U20; Onset Computer, Bourne, MA, USA) at the end of a river inflow
(Figure 2) using the Manning formula. Transparency was estimated using Secchi disc data
measured at local times (GMT+08:00) from 10:00 to 14:00.

As Table 1 shows, four strong typhoons were recorded, contributing a total of 2,254
mm of precipitation in all 24 months of 2015 and 2016, accounting for 35.6% of the annual
precipitation. However, no typhoon rainfall was recorded at YYL in 2017 and 2018; the total
precipitation in that 2-year period was around 2,537 mm. There was no significant difference in
average water depth between 2017 and 2018 (Table 1). The average discharge was less than 774
m$^3$ d$^{-1}$ in 2017 and 2018. Thus, we considered 2015 and 2016 as typhoon years, and 2017 and
2018 as non-typhoon years.

2.3.2 Conceptual two-layer DIC and DOC model

Nakayama et al. (2010) successfully developed a conceptual two-layer dissolved oxygen
model based on strong wind turbulence at Tokyo Bay. Lin et al. (2021) pointed out that thermal
stratification that inhibits vertical C flux between the upper and lower layers in shallow stratified
lakes makes it possible to develop conceptual two-layer C models (Lin et al., 2022; Nakayama et
al., 2022), and the phytoplankton and remineralization effects on DIC and DOC fluxes ($d$DIC/$dt$
and $d$DOC/$dt$, mg-C L$^{-1}$ d$^{-1}$) were considered in a conceptual two-layer equation model as shown
The fluxes in the upper layer (from the water surface to 2.5 m water depth) were calculated as follows:

\[ V_U \frac{dDIC_U}{dt} = Q_U DIC_R - Q_{out} DIC_U - V_U \alpha_{PU} Chl_U + V_U \alpha_{MU} DOC_U + A_i w_i (DIC_L - DIC_U) \]

\[ + Q_L DIC_L - \frac{A_i F_{CO2}}{C_U} + Pa_U \]  

\[ V_U \frac{dDOC_U}{dt} = Q_U DOC_R - Q_{out} DOC_U - V_U \alpha_{MU} DOC_U + A_i w_i (DOC_L - DOC_U) \]

\[ + Q_L DOC_L + Pb_U \]  

Those in the lower layer (from 2.5 to 4.0 m water depth) were calculated as follows:

\[ V_L \frac{dDIC_L}{dt} = Q_L DIC_R - V_L \alpha_{PL} Chl_L + V_L \alpha_{ML} DOC_L + A_i w_i (DIC_U - DIC_L) - Q_L DIC_L \]

\[ + \frac{A_i BF_{DIC}}{C_U} + Pa_L \]  

\[ V_L \frac{dDOC_L}{dt} = Q_L DOC_R - V_L \alpha_{ML} DOC_L + A_i w_i (DOC_U - DOC_L) - Q_L DOC_L + Pb_L \]  

\[ V_{total} = V_U + V_L \]  

\[ Q_{in} = Q_U + Q_L \]  

where, as shown in Table 2, total lake volume \((V_{total}, 53,544 \text{ m}^3)\) departs to the upper layer \((V_U, 45,456 \text{ m}^3)\) and to the lower layer \((V_L, 8,808 \text{ m}^3)\) (Equation 5), and where lake surface area \((A_s)\) is 36,000 m² and the bottom of lake area \((A_B)\) is 3,520 m². The interface is 2.5 m vertically, and the interface area \((A_I)\) is 7,264 m² in YYL. The water depth is not only steady but also changes. However, the change in water depth ranged from 4.56 to 4.66 m during the typhoon period. Therefore, we can assume that the changes in lake volumes and areas were negligible. The \(C_U\) is a coefficient value (= 1,000) to establishing a standard unit for \(F_{CO2}\) (mg-C m\(^{-2}\) d\(^{-1}\)), considering the air–water CO\(_2\) exchange by Fick’s law as follows:

\[ F_{CO2} = k_{CO2} \cdot K_H (pCO2_{water} - pCO2_{air}) \]  

where \(k_{CO2}\) is the gas transfer velocity from empirical wind speed equations (Cole and Caraco, 1998; Jähne et al., 1987; Smith, 1985; Wanninkhof, 1992). \(K_H\) is Henry’s coefficient calculated by water temperature empirical equations (Plummer and Busenberg, 1982). \(pCO2_{air}\) (μatm) is the CO\(_2\) partial pressure in the atmosphere using air pressure data (Lin et al., 2021; Lin et al., 2022), and the atmospheric CO\(_2\) concentration is assumed to be 400 ppm. \(pCO2_{water}\) (μatm) is the CO\(_2\) partial pressure at the water surface around 0.04 m water depth from water quality data.
chlorophyll \( \alpha \), \( s \alpha \). Wang, 1998 was also followed by Lin et al. (2021). \( F_{CO2} \) contributed approximately half of the net ecosystem production (NEP) across the water surface to the atmosphere in YYL (Lin et al., 2021). Further, because sediment carbon may be an important flux into shallow subtropical lakes, the sediment C flux (\( BF_{DIC} \), \( BF_{DOC} \), mg-C L\(^{-1}\)) in the lower layer was considered (Lin et al., 2022).

We assumed that the river discharge and outflow discharge \( (Q_{out}, m^3 d^{-1}) \) are in a quasi–steady state \( (Q_{in} = Q_{out}) \), dividing into upper discharge \( (Q_U, m^3 d^{-1}) \) and lower discharge \( (Q_L, m^3 d^{-1}) \) (Equation 6). Lin et al. (2021) showed that the buoyancy frequencies in YYL were 0.011 ± 0.004 s\(^{-1}\), 0.013 ± 0.004 s\(^{-1}\), 0.006 ± 0.003 s\(^{-1}\), and 0.007 ± 0.004 s\(^{-1}\) from spring (March–May), summer (June–August), autumn (September–November), and winter (December–February), respectively, inhibiting the vertical profile of DIC mixed due to stratification. We estimated the percentages of \( Q_U \) and \( Q_L \) based on the buoyancy frequency following Lin et al. (2020 and 2022). \( Q_U \) values were 75%, 80%, 45%, and 50% of \( Q_{in} \) for spring to winter, and \( Q_L \) values were 25%, 20%, 55%, and 50% of \( Q_{in} \). The physical and biogeochemical regimes under climate change remain uncertain, such as biological compositions, mixing regimes, morphometric characteristics, and air–water energy fluxes (evaporation and transpiration) (Woolway et al., 2020). To simulate extreme climate scenarios, we shifted the ratio of \( Q_{in} \) for each season and tested the river intrusion hypothesis (Fig. 1). We established two extreme conditions: Level 1 and Level 2. Level 2 is the more extreme condition: \( Q_U \) is 80% (spring), 85% (summer), 50% (autumn), and 50% (winter) of \( Q_{in} \); \( Q_L \) is 20% (spring), 15% (summer), 50% (autumn), and 50% (winter) of \( Q_{in} \). Level 1 is the condition between the present and the Level 2 condition: \( Q_U \) is 77% (spring), 82% (summer), 47% (autumn), and 50% (winter) of \( Q_{in} \); \( Q_L \) is 23% (spring), 18% (summer), 53% (autumn), 50% (winter) of \( Q_{in} \).

The contributions of photosynthesis production depended on the chlorophyll \( a \) concentration (\( Chl_U, Chl_L, mg \ L^{-1} \)) and on the absorption coefficients in the upper layer \( (\alpha_{PU}, d^{-1}) \) and lower layer \( (\alpha_{PL}, d^{-1}) \). The coefficients of DOC remineralization rates in the upper layer \( (\alpha_{MU}, d^{-1}) \) and lower layer \( (\alpha_{ML}, d^{-1}) \) were also considered in the conceptual models. The \( Pa_U, Pa_L, Pb_U, and Pb_L \) are constants in the conceptual models. To obtain unknown values \( (\alpha_{PU}, \alpha_{MU}, \alpha_{PL}, \alpha_{ML}, w_1, BF_{DIC}, BF_{DOC}, Pa_U, Pa_L, Pb_U, and Pb_L) \), we applied multiple linear regression analysis. Further, these unknown values were tested by trial and error to obtain the parameters of the best-fit condition (Nakayama et al., 2022), dividing the seasonal and nonseasonal serranoids to learn the seasonal differences. The same parameters of the best-fit condition were used to obtain the extreme conditions for Level 1 and Level 2. We used the coefficient of determination \( (R^2) \) and the Nash–Sutcliffe model efficiency coefficient \( (NSE) \).
to quantify the performance of the equation model with DIC and DOC sampling data (observation data) for each simulation.

2.3.3 DIC and DOC fluxes

VNt ecosystem production was defined as the difference between primary production and ecological respiration (NEP = GPP - ER) due to photosynthesis and respiration via biota (Dodds and Whiles, 2020). Given that we assumed that the C fluxes were dependent on the river inflows in YYL (Fig. 2), we estimated the NEP by end-member analysis using the C concentration of the river inflow and outflow (Lin et al., 2021; Nakayama et al., 2020) by following Equation 8–Equation 11. The upper layer NEP of DIC flux (mg C d⁻¹) was obtained from Equation 1 as follows:

\[
\text{Upper flux}_{\text{DIC}} = C_U \alpha_{P} Chl_U - C_U \alpha_{MU} DOC_U - C_U \frac{A_I w_I (DIC_L - DIC_U)}{V_U} - C_U \frac{P a_U}{V_U} \\
= C_U \frac{Q_U DIC_R + Q_L DIC_L - Q_{out} DIC_U}{V_U} - \frac{A_S}{V_U} F_{CO2} \\
= C_U \frac{1}{\tau_{ru}} \left( \frac{Q_U}{Q_{in}} DIC_R + \frac{Q_L}{Q_{in}} DIC_L - DIC_U \right) - F_C \\
\tau_{ru} = \frac{V_U}{Q_{in}} 
\] (8)

The upper layer flux of DOC flux (mg C m⁻³ d⁻¹) was estimated from Equation 2:

\[
\text{Upper flux}_{\text{DOC}} = C_U \alpha_{MU} DOC_U - C_U \frac{A_I w_I (DOC_L - DOC_U)}{V_U} - C_U \frac{P b_U}{V_U} \\
= C_U \frac{Q_U DOC_R + Q_L DOC_L - Q_{out} DOC_U}{V_U} \\
= C_U \frac{1}{\tau_{ru}} \left( \frac{Q_U}{Q_{in}} DOC_R + \frac{Q_L}{Q_{in}} DOC_L - DOC_U \right) 
\] (9)
The lower layer flux of DIC flux (mg C m$^{-3}$ d$^{-1}$) was estimated from Equation 3:

\[
\text{Lower flux}_{\text{DIC}} = C_U \alpha_{PL} Ch_L - C_U \alpha_{ML} DOC_L - C_U \frac{A_I w_I (\text{DIC}_U - \text{DIC}_L)}{V_L} - \frac{A_B BF_{\text{DIC}}}{V_L} - C_U \frac{P_a L}{V_L} = C_U \frac{1}{t_{rL}} \frac{Q_L}{Q_{in}} (\text{DIC}_R - \text{DIC}_L) \]

\[t_{rL} = \frac{V_L}{Q_{in}} \tag{10}\]

The lower layer flux of DOC flux (mg C m$^{-3}$ d$^{-1}$) was estimated from Equation 4:

\[
\text{Lower NEP}_{\text{DOC}} = C_U \alpha_{ML} DOC_L - C_U \frac{A_I w_I (\text{DOC}_U - \text{DOC}_L)}{V_L} - \frac{A_B BF_{\text{DOC}}}{V_L} - C_U \frac{P b L}{V_L} = C_U \frac{Q_L (\text{DOC}_R - \text{DOC}_L)}{V_L} = C_U \frac{1}{t_{rL}} \frac{Q_L}{Q_{in}} (\text{DOC}_R - \text{DOC}_L) \]

\[t_{rL} = \frac{V_L}{Q_{in}} \tag{11}\]

Thus, the total flux of DIC and DOC are:

\[\text{Flux}_{\text{DIC}} = \frac{V_U \text{Upper flux}_{\text{DIC}} + V_L \text{Lower flux}_{\text{DIC}}}{V_{total}} \tag{12}\]

\[\text{Flux}_{\text{DOC}} = \frac{V_U \text{Upper flux}_{\text{DOC}} + V_L \text{Lower flux}_{\text{DOC}}}{V_{total}} \tag{13}\]

where, $F_C$ is $\frac{A_S}{V_U} F_{CO2}$ and $t_{rU}$, $t_{rL}$ are residence times (d) in the upper and lower layers, respectively. These parameters were used for the best-fit condition as shown in Table 2.
3. Results

3.1 DIC, DOC, and Chl. a concentrations in the typhoon and non-typhoon years

The results of the comparisons between the two periods of typhoon years demonstrated that there were no significant differences in DIC, DOC, and Chl a concentration between the layers in the typhoon years 2015 and 2016; however, all these parameters differed significantly between the layers in the typhoon years 2017 and 2018 (Fig. 3). The average DIC$_U$ was 1.23 mg-C L$^{-1}$, and DIC$_L$ was 3.66 mg-C L$^{-1}$; the average DOC$_U$ was 5.87 mg-C L$^{-1}$, and DOC$_L$ was 8.02 mg-C L$^{-1}$; and the Chl$_U$ and Chl$_L$ were 18.5 µg-C L$^{-1}$ and 2.13 µg-C L$^{-1}$, respectively (Fig. 3). However, the t-test results showed no significant differences in DIC, DOC, and Chl. a concentrations ($p$-values $\geq 0.05$) in the typhoon years (Fig. 4 a). In the non-typhoon years, the upper layer DIC$_L$ was the only variable negatively correlated with DOC in the upper and lower layers (Fig. 5b). The DIC of the lower layer was positively correlated with the Chl$_L$ due to the abundant respiration in the lower layer (Fig. 5).
### 3.2 Performance of conceptual two-layer DIC and DOC models

The results for the typhoon years demonstrated that most of the seasonal scenarios were better fitting than the nonseasonal scenarios (Fig. 6). Under the seasonal scenarios, the DIC$_U$ was around 1.5 to 5.0 mg-C L$^{-1}$ (Fig. 6a-b) and DIC$_L$ was around 5.0 mg-C L$^{-1}$ stably (Fig. 6d). However, the NSE of DIC$_L$ was 0.73 under the nonseasonal scenarios, which was higher than seasonal scenarios (NSE = 0.71) (Table 2), because DIC$_L$ was elevated dramatically, by 40 mg-C L$^{-1}$, under the nonseasonal scenarios during the 2016 typhoon period (Fig. 6c). In the non-typhoon years (2017–2018), the best-fit values of DIC$_U$ and DIC$_L$ did not differ significantly between the seasonal and nonseasonal scenarios ($R^2$ and NSE were around 0.40 and 0.70, respectively). These results demonstrated that DIC$_U$ and DIC$_L$ in the typhoon years must use the seasonal scenarios, whereas in the non-typhoon years they should use the nonseasonal scenarios. On the other hand, the DOC under the seasonal scenarios fit our observation data ($R^2$ = 0.91, 0.46 and NSE = 0.95, 0.73 for DOC$_U$, DOC$_L$, respectively) (Fig. 6e-h, Table 3). Thus, the results suggested that the DOC$_U$ and DOC$_L$ must use the seasonal scenarios in both the typhoon and non-typhoon years.

As shown in Table 3, the parameters for the conceptual two-layer DIC and DOC models showed different regimes between the typhoon and non-typhoon years. In the typhoon years, the photosynthesis absorption rates coefficients ($\alpha_{PU}$, $\alpha_{PL}$) were negative (photosynthesis < respiration) for each season. YYL was a C source due to a large allochthonous C loading during typhoons; the respiration was elevated by around 30- to 150-fold from summer to autumn. However, the transportation coefficients ($w_I$) were higher in autumn than in the other seasons (Table 3) due to weak stratification and large C loading during typhoons. Further, the higher remineralization rates during typhoon disturbances from summer to autumn resulted in positive $\alpha_{MU}$ and $\alpha_{ML}$. In the non-typhoon years, the remineralization rates were negative (Table 3). Thus, the results suggest that the conceptual two-layer C models may reasonably fit the observation data.
3.3 Interannual and seasonal NEP in YYL

The typhoon disturbances in summer and autumn played an important role in promoting the C released by YYL (Table 4). Overall, YYL released 245 mg C m\(^{-3}\) d\(^{-1}\) of DIC and 415 mg C m\(^{-3}\) d\(^{-1}\) of DOC during the typhoon years; during the non-typhoon years, it released 51.7 mg C m\(^{-3}\) d\(^{-1}\) of DIC and 22.8 mg C m\(^{-3}\) d\(^{-1}\) of DOC (Table 4). The average \(F_C\) was one to two times larger than \(\text{Flux}_{\text{DIC}}\), and 219 and 133 mg C m\(^{-3}\) d\(^{-1}\) were released from YYL into the atmosphere in the typhoon and non-typhoon years, respectively (Table 4). In summer, the upper layer DIC and DOC consumed approximately 3.7 times more DIC in the typhoon years than in the non-typhoon years (Table 4). In autumn, 216 mg C d\(^{-1}\) of upper layer DIC was released; however, 46.1 mg C m\(^{-3}\) d\(^{-1}\) of upper layer DOC was produced in the typhoon years. The upper layer \(\text{Flux}_{\text{DIC}}\) was negative in the autumn of the typhoon years, when 268 mg C m\(^{-3}\) d\(^{-1}\) more \(F_C\) was released compared to the non-typhoon years. In addition, the lower layer was most released of C into the outflow; however, the fluxes in the lower layer were more than twice as high in the summer as in the autumn of the typhoon years (Table 4). The average of total \(\text{Flux}_{\text{DIC}}\) was 3.14 times more released C in the typhoon than in the non-typhoon years; the average of total \(\text{NEP}_{\text{DOC}}\) was increased 62.3 mg C m\(^{-3}\) d\(^{-1}\) of DOC between the typhoon years and non-typhoon years due to the over ten-times higher flux in the upper layer (Table 4).

The ratios of DIC and DOC concentrations reveal the magnitudes of allochthonous DOC loading into YYL (Shih et al., 2019; Walvoord and Striegl, 2007), and the upper and lower layers show different patterns. In the typhoon years, the upper layer ratios decreased (higher DOC loading) from summer to autumn, whereas in the lower layer, the DIC:DOC decreased from autumn to winter. In the non-typhoon years, the autumn DIC:DOC ratio was the lowest, around 0.216 to 0.351 (Table 4).
3.4 Interannual responses of DIC and DOC to typhoons

We simulated the responses of DIC and DOC flux to typhoons using conceptual two-layer C models. The results showed that the DIC was more sensitive to typhoon disturbances than DOC under the scenarios of Level 1 and Level 2 (Fig. 7-9). Overall, the C level declined in the upper layers but increased in the lower layers (Fig. 7). DIC and DOC in the upper layer tended to decline from 1.0 (Level 1) to 2.0 mg-C L$^{-1}$ (Level 2) (Fig. 7a, c); however, they increased to 10.0 and 20.0 mg-C L$^{-1}$ in the lower layer under Level 1 and Level 2, respectively (Fig. 7b, d).

The DIC concentration in the upper layer was significantly lower in typhoon than in the non-typhoon years during spring and autumn under Level 2 (Fig. 8a-c). Under the best-fit and Level 1 conditions, the DIC concentrations decreased significantly from winter to spring (Fig. 8e-d). The lower layer DIC values under the best-fit and Level 1 conditions differed significantly between the typhoon and non-typhoon years (Fig. 8e-h). The DIC of the lower layer under Level 2 differed significantly from winter to spring only (Fig. 8e, h). However, the upper layer DOC showed significant typhoon responses for each condition from winter to spring (Fig. 9a, d). The DOC of the upper layer tended to differ more significantly under the extreme climates from summer to autumn (Fig. 9b-c). The DOC of the lower layer showed different typhoon responses between the spring and the other seasons (Fig. 9 e-h).
4. Discussion

Annual total precipitation was 40% higher in typhoon years than in non-typhoon years (Table 1). Water retention and typhoon-induced upwelling control the dynamics of DIC and DOC during the summer and autumn (Chiu et al., 2020; Jones et al., 2009; Tsai et al., 2008; Tsai et al., 2011). Typhoon-induced upwelling affected water quality data differently between the upper and lower layers (Fig. 3). DIC, DOC, and Chl. a concentrations differ significantly between upper and lower layers in the typhoon years (Fig. 3) due to thermal stratification (Chiu et al., 2020; Lin et al., 2022; Tsai et al., 2008; Tsai et al., 2011). Further, the abundance of organisms leads to intensive respirations in the lower layers during the non-typhoon period; for example, an anoxic condition at the hypolimnion may affect C mineralization and remineralization rates in non-typhoon years (Carey et al., 2022; Chiu et al., 2020; Lin et al., 2022; Shade et al., 2010; Shade et al., 2011). Therefore, these physical and biogeochemical processes might describe different patterns between the upper and lower layers, as revealed by the Pearson correlations (Fig. 5).

Thermal stratification and allochthonous C loading may drive the responses of fluxes to typhoons in YYL. In the typhoon years, the absolute values of fluxes were higher than in the non-typhoon years (Table 4). We found that precipitation from typhoons loaded large amounts of allochthonous C into YYL during summer and autumn, which might describe the higher fluxes in autumn compared to other seasons (Table 4). Typhoons dramatically changed the seasonal and interannual patterns of DIC fluxes due to river intrusion (Fig. 7a–b; Fig. 8), which corresponds to our hypothesis (Fig. 1) and to the results of previous studies (Chiu et al., 2020; Lin et al., 2021; Lin et al., 2022). In summer, the spatial differences in DIC and DOC between layers were inhibited due to strong thermal stratification, describing the positive upper net primary production and lower negative net primary production (Lin et al., 2021). The thermal stratification and anoxic condition may have been controlled by the seasonal and interannual patterns of DIC and DOC fluxes in the non-typhoon years (Tables 3–4; Fig. 6). Additionally, because of the absence of typhoon-induced mixing and allochthonous C loading, the absolute values of total fluxes in the non-typhoon years were less than those in the non-typhoon years (Table 4). These results suggested that the allochthonous C loading was the most crucial factor for DIC and DOC fluxes in the typhoon years; on the other hand, the transportation rate shaped the seasonal C due to thermal stratification in the non-typhoon years.

Water temperature might be a crucial driver in controlling C fluxes in YYL (Chiu et al., 2020; Lin et al., 2021; Lin et al., 2022). We found that the fluxes and $F_{CO2}$ in summer were usually higher than in winter (Tables 3–4) due to the higher levels of photosynthesis, remineralization, and strength of thermal stratification (Lin et al., 2021; Lin et al., 2022). With the conceptual two-layer C models (Table 3), photosynthesis absorption ($\alpha_{PU}$, $\alpha_{PL}$),
remineralization ($\alpha_{MU}$, $\alpha_{ML}$), and transportation ($w_t$) well represented the seasonal variations of DIC and DOC data. These parameters of the conceptual two-layer C models appeared in reasonable patterns (Table 3). The higher remineralization and photosynthesis rates resulted in higher absolute values of fluxes in the autumn of the typhoon years (Tables 3-4). In the non-typhoon years, the photosynthesis rates contributed to the total fluxes (Tables 3-4). Moreover, without the typhoon-induced mixing and refreshing of the water column, anoxic conditions may occur (Carey et al., 2022; Vachon et al., 2021), which could result in negative remineralization rates in non-typhoon years. Thus, the conceptual two-layer C models well characterizes the seasonal and interannual responses of DIC and DOC fluxes to typhoons in YYL.

Under extreme weather conditions, Level 2 usually shifted to different typhoon responses for each season (Fig. 8-9) due to extreme river intrusions. DIC changes more significantly than DOC under Level 1 and Level 2 (Fig. 7-9), because the photosynthesis, transportation, and remineralization rates may crucially affect the seasonal and interannual patterns of DOC as well (Fig. 1). Moreover, we compared the fluxes with different model conditions as shown in Fig. 10, demonstrating that the responses of FluxDIC to typhoons differed dramatically between Level 1 and Level 2 (Fig. 10a-c); especially, the Upper FluxDIC released more C in the typhoon years and absorbed more C in the non-typhoon years than Obs (Fig. 10a).

Not only were the absolute values of FluxDIC over 3 times higher in the typhoon than the non-typhoon years (Table 4), but SD was higher in the typhoon years as well (Fig. 10). Additionally, the $F_C$ was 43% higher (~83 mg C m$^{-3}$ d$^{-1}$) in typhoon years than in non-typhoon years (Table 4). Therefore, the typhoon disturbances control DIC loading and C emissions in YYL, which is consistent with our previous studies (Chiu et al., 2020; Lin et al., 2021; Lin et al., 2022).

However, DOC fluxes changed less under Level 1 and Level 2 (Fig. 10d-f), a finding that is consistent with our continuous DOC data (Fig. 7c-d). Processes such as respiration, mineralization, and sediment burial may impact DOC fluxes (Bartosiewicz et al., 2015; Hanson et al., 2015; Maranger et al., 2018). To our knowledge, bio-photochemical mineralization and degradation may play a key role in shaping C fluxes because colored DOC reduces ultraviolet radiation (UVR) and active photosynthetic radiation (PAR) (Allesson et al., 2021; Chiu et al., 2020; Schindler et al., 1996; Scully et al., 1996; Williamson et al., 1999). Ejarque et al. (2021) also successfully developed a conceptual one-layer model of DOC and DIC, considering bacterial respiration, photo-mineralization and degradation in a temperate mountain lake. In addition., Nagatomo et al. (2023) suggested that the DIC might be underestimated if submerged vegetation is ignored. Thus, we suggest that photo-biochemical processes (such as photo-mineralization) and submerged vegetation should be considered in the upper layer to clarify and validate the responses of the total C fluxes under extreme climates in a two-layer stratified lake.
5. Conclusions

Our conceptual two-layer C model revealed that allochthonous and autochthonous processes both accounted for C flux responses to typhoon disturbances on seasonal and interannual scales by applying our proposed two-layer conceptual C model. Without typhoons, the strength of thermal stratification was the primary determinants the seasonal and interannual patterns of DIC and DOC concentrations data and typhoon-induced fluxes upwelling and loading facilitated 102.2 mg-DIC m⁻³ d⁻¹ and 62.3 mg-DOC m⁻³ d⁻¹ in YYL, respectively (Table 4). We successfully developed two-layer conceptual C models to obtain continuous DIC and DOC data in YYL and to simulate extreme conditions. The changes in seasonal river intrusion regimes in YYL resulted in a 3-fold higher total FluxDIC in the typhoon years than in the non-typhoon years. However, our model should be improved under extreme climate scenarios by considering other autochthonous processes, such as sediment burial, photo-biochemical processes, and anoxic conditions. The present results suggest that physical processes (river intrusion and vertical transportation) and biogeochemical processes (mineralization, photosynthesis, and respiration) in a subtropical small lake account for the C flux responses to typhoons on seasonal and interannual time scales.
Competing interests

The authors have no conflicts of interest to report.

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Hao-Chi Lin: Conceptualization, Methodology, Investigation, Formal analysis, Writing – original draft. Keisuke Nakayama: Methodology, Supervision, Writing – review & editing, Conceptualization. Jeng-Wei Tsai: Investigation, Funding acquisition, Writing – review & editing. Chih-Yu Chiu: Funding acquisition, Writing – review & editing.

Data availability

The data that support the findings of this study are adopted from our previous works, including Chiu et al. (2020), Lin et al. (2021), and Lin et al. (2022).
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onset of and recovery from hypoxia in Tokyo Bay, Japan: Derived distribution analysis


Williamson, C. E., Morris, D. P., Pace, M. L., and Olson, O. G.: Dissolved organic carbon and nutrients as regulators of lake ecosystems: Resurrection of a more integrated paradigm,


Table 1. Comparison of Yuan-Yang Lake’s rainfall and hydrological records in typhoon and non-typhoon years.

<table>
<thead>
<tr>
<th>Records</th>
<th>Typhoon years</th>
<th>Non-typhoon years</th>
</tr>
</thead>
<tbody>
<tr>
<td>Time period (year)</td>
<td>2015-2016</td>
<td>2017-2018</td>
</tr>
<tr>
<td>Total precipitation (mm)</td>
<td>6,332</td>
<td>3,795</td>
</tr>
<tr>
<td>Total typhoon rainfall (mm)</td>
<td>2,254</td>
<td>0</td>
</tr>
<tr>
<td>Average water depth (m ± SD)</td>
<td>4.54 ± 1.7</td>
<td>4.51 ± 1.5</td>
</tr>
<tr>
<td>Average river discharge (m$^3$ d$^{-1}$)</td>
<td>3,717</td>
<td>2,943</td>
</tr>
<tr>
<td>Transparency (Secchi disc depth, m ± SD)</td>
<td>1.58 ± 0.45</td>
<td>1.38 ± 0.28</td>
</tr>
</tbody>
</table>
Table 2. Parameters of the two-layer conceptual model in Yuan-Yang Lake

<table>
<thead>
<tr>
<th>Parameters</th>
<th>Value</th>
<th>Unit</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Measurements</strong></td>
<td></td>
<td></td>
</tr>
<tr>
<td>$Q_{\text{out}}$</td>
<td>Outflow discharge</td>
<td>Daily data</td>
</tr>
<tr>
<td>$Q_{\text{in}}$</td>
<td>Inflow discharge</td>
<td>Daily data</td>
</tr>
<tr>
<td>$Q_U$</td>
<td>Upper layer Discharge</td>
<td>Daily data</td>
</tr>
<tr>
<td>$Q_L$</td>
<td>Lower layer discharge</td>
<td>Daily data</td>
</tr>
<tr>
<td>DIC$R$</td>
<td>River inflow DIC</td>
<td>Monthly data</td>
</tr>
<tr>
<td>DIC$U$</td>
<td>Upper layer DIC</td>
<td>Monthly data</td>
</tr>
<tr>
<td>DIC$L$</td>
<td>Lower layer DIC</td>
<td>Monthly data</td>
</tr>
<tr>
<td>Chl$U$</td>
<td>Upper layer Chl $a$</td>
<td>Monthly data</td>
</tr>
<tr>
<td>Chl$L$</td>
<td>Lower layer Chl $a$</td>
<td>Monthly data</td>
</tr>
<tr>
<td>DOC$U$</td>
<td>Upper layer DOC</td>
<td>Monthly data</td>
</tr>
<tr>
<td>DOC$L$</td>
<td>Lower layer DOC</td>
<td>Monthly data</td>
</tr>
<tr>
<td>$F_{CO2}$</td>
<td>Carbon emission (equation 7)</td>
<td>Monthly data</td>
</tr>
<tr>
<td><strong>Constants</strong></td>
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<td></td>
</tr>
<tr>
<td>$V_{\text{total}}$</td>
<td>Total lake volume</td>
<td>53,544</td>
</tr>
<tr>
<td>$V_U$</td>
<td>Upper layer volume</td>
<td>45,456</td>
</tr>
<tr>
<td>$V_L$</td>
<td>Lower layer volume</td>
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<td>$A_s$</td>
<td>Lake surface area</td>
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<tr>
<td>$A_I$</td>
<td>The interface area</td>
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<tr>
<td>$A_B$</td>
<td>The bottom of lake area</td>
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</tr>
<tr>
<td>$C_U$</td>
<td>Coefficient of the unit uniform</td>
<td>1,000</td>
</tr>
<tr>
<td><strong>Unknown Constants</strong></td>
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<tr>
<td>$\alpha_{PU}$, $\alpha_{PL}$</td>
<td>Coefficients of photosynthesis</td>
<td>Constant</td>
</tr>
<tr>
<td>$\alpha_{MU}$, $\alpha_{ML}$</td>
<td>Coefficients of mineralization</td>
<td>Constant</td>
</tr>
<tr>
<td>$w_I$</td>
<td>Coefficient of vertical transportation</td>
<td>Constant</td>
</tr>
<tr>
<td>$BF_{DIC}$, $BF_{DOC}$</td>
<td>Sediment DIC and DOC emission</td>
<td>Constant</td>
</tr>
<tr>
<td>$Pa_U$, $Pb_L$</td>
<td>Equations constant at lower layer</td>
<td>Constant</td>
</tr>
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Table 3. Best-fit parameters of a two-layer conceptual model of DIC and DOC in Yuan-Yang Lake from 2015 to 2018.

<table>
<thead>
<tr>
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<tbody>
<tr>
<td></td>
<td>Typhoon years</td>
<td>Non-typhoon years</td>
<td>Typhoon years</td>
<td>Non-typhoon years</td>
</tr>
<tr>
<td></td>
<td>Spring Summer Autumn Winter</td>
<td>Spring Summer Autumn Winter</td>
<td>Inter-annual</td>
<td>Inter-annual</td>
</tr>
<tr>
<td>Upper layer</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>$F_{CO2}$ (mg-C m$^2$ d$^{-1}$)</td>
<td>291, 245, 422, 127</td>
<td>231, 143, 104, 175</td>
<td>276, 163</td>
<td></td>
</tr>
<tr>
<td>$\alpha_{PU}$ (d$^{-1}$)</td>
<td>-1.20, -33.1, -183.5, -29.1</td>
<td>8.0, 6.0, 30.0, 7.77</td>
<td>-22.0, 8.0</td>
<td></td>
</tr>
<tr>
<td>$\alpha_{MU}$ (d$^{-1}$)</td>
<td>-0.0227, 0.0203, 0.08, -0.031</td>
<td>-0.01, -0.039, -0.033, -0.195</td>
<td>-0.035, -0.0238</td>
<td></td>
</tr>
<tr>
<td>$w_I$ (d$^{-1}$)</td>
<td>0.230, 0.172, 1.38, 0.30</td>
<td>0.10, 0.0478, 0.120, 0.180</td>
<td>0.159, 0.107</td>
<td></td>
</tr>
<tr>
<td>$Pa_U$ (d$^{-1}$)</td>
<td>12560, -1317, -23750, 9597</td>
<td>9880, 14000, 17600, 10100</td>
<td>4457, 12420</td>
<td></td>
</tr>
<tr>
<td>$Pb_U$ (d$^{-1}$)</td>
<td>-21930, 9461, -42130, -17070</td>
<td>-3630, -1251, -20820, -9289</td>
<td>-12760, -9119</td>
<td></td>
</tr>
<tr>
<td>$dDIC_U$</td>
<td>0.305, 0.614</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>(R$^2$, NSE)</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>$dDOC_U$</td>
<td>0.909, 0.953</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>(R$^2$, NSE)</td>
<td></td>
<td></td>
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<tr>
<td>BF$_{DIC}$,</td>
<td>0.04,</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>BF$_{DOC}$</td>
<td>0.00</td>
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<tr>
<td>Lower layer</td>
<td></td>
<td></td>
<td></td>
<td></td>
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<tr>
<td>$\alpha_{PL}$ (d$^{-1}$)</td>
<td>-0.627, -22.1, 15.0, -0.878</td>
<td>1.49, -6.87, 6.0, -16.6</td>
<td>-21.11, 2.0</td>
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</tr>
<tr>
<td>$\alpha_{ML}$ (d$^{-1}$)</td>
<td>-0.025, 0.123, 0.0755, 0.00973</td>
<td>-0.010, -0.0376, -0.04, -0.048</td>
<td>0.123, -0.019</td>
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<tr>
<td>$Pa_L$ (d$^{-1}$)</td>
<td>100, -5662, -10500, -1013</td>
<td>151.6, 2032, 1216, 909</td>
<td>-5662, -40.5</td>
<td></td>
</tr>
<tr>
<td>$Pb_L$ (d$^{-1}$)</td>
<td>-6012, -7395, -53940, -9639</td>
<td>-1338, -6296, -19470, -8748</td>
<td>-12240, -9919</td>
<td></td>
</tr>
<tr>
<td>BV$_{DIC}$,</td>
<td>0.04,</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>BV$_{DOC}$</td>
<td>0.00</td>
<td></td>
<td></td>
<td></td>
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<tr>
<td>(mg-C L$^{-1}$)</td>
<td></td>
<td></td>
<td></td>
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</tr>
<tr>
<td>$dDIC_L$</td>
<td>0.452, 0.707</td>
<td>0.192, 0.440</td>
<td>0.306, 0.731</td>
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</tr>
<tr>
<td>(R$^2$, NSE)</td>
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<td></td>
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<td></td>
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<tr>
<td>$dDOC_L$</td>
<td>0.460, 0.728</td>
<td>0.234, 0.128</td>
<td>0.338, 0.525</td>
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</tbody>
</table>
Table 4. Seasonal averages of carbon fluxes (mg C m$^{-3}$ d$^{-1}$) for each season in Yuan-Yang Lake. Positive values are shown in the carbon sink (*black*), and negative ones show the values after carbon was released (*red*).

<table>
<thead>
<tr>
<th></th>
<th>Flux (mg C m$^{-3}$ d$^{-1}$)</th>
<th>DIC$_U$</th>
<th>DIC$_L$</th>
<th>Total (mg C m$^{-3}$ d$^{-1}$)</th>
<th>Flux$_{DIC}$</th>
<th>Flux$_{DOC}$</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>$F_c$ Upper  Lower</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Typhoon years</td>
<td>Average -219 - -</td>
<td>-</td>
<td>-</td>
<td>-150</td>
<td>-9.69</td>
<td></td>
</tr>
<tr>
<td>Spring</td>
<td>DIC -231 -243 -45.2</td>
<td>0.658</td>
<td>0.568</td>
<td>-210</td>
<td>62.1</td>
<td></td>
</tr>
<tr>
<td></td>
<td>DOC - 70.8 17.2</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Summer</td>
<td>DIC -194 29.1 -313</td>
<td>0.193</td>
<td>0.511</td>
<td>-26.4</td>
<td>18.8</td>
<td></td>
</tr>
<tr>
<td></td>
<td>DOC - 118 -495</td>
<td></td>
<td></td>
<td></td>
<td></td>
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</tr>
<tr>
<td>Autumn</td>
<td>DIC -351 -216 -659</td>
<td>0.349</td>
<td>0.475</td>
<td>-288</td>
<td>-151</td>
<td></td>
</tr>
<tr>
<td></td>
<td>DOC - 46.1 -1167</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Winter</td>
<td>DIC -100 -96.4 36.5</td>
<td>0.442</td>
<td>0.372</td>
<td>-74.8</td>
<td>31.2</td>
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<tr>
<td></td>
<td>DOC - 40.5 -16.9</td>
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</tr>
<tr>
<td>Non-typhoon years</td>
<td>Average -133 - -</td>
<td>-</td>
<td>-</td>
<td>-47.8</td>
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<td>DOC - 34.0 32.1</td>
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**Fig. 1.** Conceptual diagram of river intrusion (*red arrows*) and thermal stratification (*red dashed line*) dominant responses of DIC and DOC in a subtropical two-layer stratified lake under extreme climates.
Fig. 2. Sampling locations and bathymetry maps of Yuan–Yang Lake (YYL). The dark green rectangle shows the buoy station, which is at the deepest site of the lake. The red points and white cross show the river mouths of the inflows and outflows, respectively. The gray triangle shows the location of the water depth sensor.
Fig. 3. Comparisons of (a) DIC, (b) DOC, and (c) Chl a between upper (DIC_U, DOC_U, Chl_U) and lower (DIC_L, DOC_L, Chl_L) layers, grouped by typhoon and non-typhoon years. The bullet points show the water sampling data. We used a t-test to obtain the p-values. The ns show the p-values ≥ 0.05, * show p-values from 0.05 to 0.01, and ** show p-values from 0.01 to 0.001 by t-test.
Fig. 4. Seasonal variations of (a) upper layer DIC ($DIC_U$), (b) lower layer DIC ($DIC_L$), (c) upper layer DOC ($DOC_U$), (d) lower layer DOC ($DOC_L$), (e) upper layer Chl. a ($Chl_U$), (f) lower layer Chl. a ($Chl_L$) grouped by typhoon and non-typhoon years. The bullet points show the water sampling data. To determine the seasonality, we used one-way ANOVA to obtain the $p$-values. The ns show $p$-values $\geq 0.05$, * show $p$-values from 0.05 to 0.01, and ** show $p$-values are from 0.01 to 0.001.
Fig. 5. Pearson correlation coefficients of DIC, DOC, and Chl. a concentration at upper layer and lower layer DIC ($DIC_U$, $DIC_L$), DOC ($DOC_U$, $DOC_L$), Chl. a ($Chl_U$, $Chl_L$) during (a) typhoon years and (b) non-typhoon years. The black-crosses show insignificant values ($p$-values are $> 0.05$).
Fig. 6. Continuous daily DIC and DOC data at (a, b, c, f) the upper layer (DIC$_U$, DOC$_U$) and (c, d, g, h) lower layer (DIC$_L$, DOC$_L$) by using conceptual equations models. The gray lines show the interannual data, the black lines show the best fit for DIC, the red lines show the best-fit for DOC (Table 3), and the empty dots show water sampling (observation) data for each month.
Fig. 7. Continuous daily DIC and DOC data at (a, c) upper layer (DIC$_U$, DOC$_U$) and (b, d) lower layer (DIC$_L$, DOC$_L$) by using the conceptual equation model under extreme climates from 2015 to 2018. Blue lines are original best-fit data as in Fig. 6, in which the parameters of the DIC model in non-typhoon years are under the nonseasonal scenario and the others are under the seasonal scenario as in Table 3. Orange regions show Level 1; pink regions show Level 2.
Figure 8. Seasonal responses of continuous (a-d) upper layer DIC and (e-h) lower layer DIC (mg-C L$^{-1}$) between typhoon (Typhoon) and non-typhoon (Non) years for each season as in Fig. 4. Fit (blue boxes) condition shows the best-fit data by using the conceptual two-layer C model; Lv1 (yellow boxes) and Lv2 (red boxes) show the extreme climates. The empty dots show the continuous DIC and DOC data. The ns show p-values $\geq 0.05$, * show p-values from 0.05 to 0.01, ** show p-values from 0.01 to 0.001; **** show p-values less than 0.0001 based on a t-test.
Fig. 9. Seasonal responses of (a-d) upper layer DOC and (e-h) lower layer DOC (mg-C L\(^{-1}\)) between typhoon (Typhoon) and non-typhoon (Non) years for each season as in Fig. 8. The Fit (blue boxes) condition shows the best-fit data by using the conceptual two-layer carbon model; Lvl (yellow boxes) and Lv2 (red boxes) show the extreme climates. Empty dots show the continuous DIC and DOC data. The ns show p-values ≥ 0.05, * show p-values from 0.05 to 0.01, ** show p-values from 0.01 to 0.001; **** show p-values less than 0.0001 based on a t-test.
Fig. 10. Interannual (a) Upper flux_{DIC}, (b) Lower flux_{DIC}, (c) Flux_{DIC}, (d) Upper flux_{DOC}, (e) Lower flux_{DOC}, and (f) Flux_{DOC} (mg C m^{-3} d^{-1}) grouped by typhoon and non-typhoon years. The Obs condition (black boxes) show the observation data as in Fig. 6; The Fit condition (blue- boxes) shows the best-fit data by using the conceptual two-layer carbon model as in Fig.6; Level 1 (yellow boxes) and Level 2 (red boxes) show the extreme scenarios as in Fig. 7. The definitions of fluxes please see Sect. 2.3.3.