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

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Abstract

Extreme climates affect the seasonal and interannual patterns of carbon (C) distribution in lentic ecosystems due to the regimes of river inflow and thermal stratification. Typhoons rapidly load substantial amounts of terrestrial C into smaller subtropical lakes (i.e., Yuan-Yang Lake, YYL, Taiwan), renewing and mixing the water column. We developed a 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 transformation. To compare the temporal patterns of fluxes between typhoon years (2015–2016) and non-typhoon years (2017–2018), monthly field samples were obtained and their DIC, DOC, and chlorophyll a concentrations measured. The results demonstrated that net ecosystem production was 3.14 times higher in the typhoon years than in the non-typhoon years. This results suggested that a loading of allochthonous C was the most crucial driver of the temporal variation of C fluxes in the typhoon years because of changes in 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 freshwater 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 around 0.32 to 1.8 Pg C yr⁻¹, which is approximately equivalent to shallow coastal areas, providing availability to food webs that support human resources—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 ease with which C mixes within 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 released 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, our understanding of C fluxes in small lakes remain uncertain because small lakes have usually been ignored in calculations of C fluxes on a global scale (Cole et al., 2007; Raymond et al., 2013). Thus, elucidation of the C fluxes in small lakes in extreme weathers conditions 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). 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). Not only is measurement difficult, but so in the elucidation of the dynamics and interactions between...
Dissolved inorganic carbon (DIC) concentration is an important factor in estimating CO$_2$ fluxes within lake ecosystems (Smith, 1985). Among C fluxes in a freshwater body, the partial pressure of CO$_2$ ($p$CO$_2$), defined as CO$_2$ emission across the air–water interface, is affected by DIC, water temperature, wind speed, and pH (Jähne et al., 1987; Smith, 1985). River inflows, sediment C burial, and heterotrophic respiration in the water column contribute to DIC dynamics in lakes (Hope et al., 2004; Vachon et al., 2021); simultaneously, autotrophic organisms, such as plankton 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 $p$CO$_2$ estimation (Hanson et al., 2015; Lin et al., 2022; Nakayama et al., 2022). Dissolved organic carbon (DOC) might contribute to CO$_2$ 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).

Typhoons might significantly impact C distributions within the water columns in subtropical regions (Chiu et al., 2020; Lin et al., 2022). Kossin et al. (2013) investigated global storm events with an accumulated rainfall of about 50 mm, which accounts for approximately 10%–40% of precipitation in a subtropical typhoon event. Some studies found not only that extreme rainstorms would impact the dissolved carbon in large lakes and catchments due to weathering (Sun et al., 2021; Zhou et al., 2023) but also 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). However, the complex interactions between biogeochemical and physical regimes for autochthonous and allochthonous C introduce uncertainty in elucidating the complete patterns between typhoons and dissolved C concentrations in small subtropical lakes. This uncertainty hinders our understanding of the seasonal and interannual variations in DIC and DOC concentrations (Lin et al., 2022). Thus, to understand the seasonal regimes and to estimate C fluxes in subtropical lakes, we investigated the variations in of DIC and DOC due to typhoon disturbances.

Typhoons’ effects on C fluxes were previously studied in a small, two-layer stratified, subtropical lake, Yuan–Yang Lake (YYL) in Taiwan (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 CO$_2$ emission in YYL during typhoon disturbances that occurred in October 2004: 2.2 to 2.7 g C m$^{-2}$ d$^{-1}$ of CO$_2$ was released into the atmosphere. CO$_2$ emissions into the atmosphere were recorded at around 3.0 to 3.7 g C m$^{-2}$ d$^{-1}$ in summer and autumn 2015 because of substantial loads of terrestrial C via river inflows after strong typhoons in YYL (Chiu et al.,
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). The results of 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 controls not only 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). Therefore, we hypothesized that allochthonous C loading and river inflow intrusion might affect DIC and DOC distributions. Further, autochthonous processes in small subtropical lakes, such as algal photosynthesis, remineralization, and vertical transportation, must also be considered. 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) (Fig. 1). Its water is brown because of the humic acid content (colored dissolved organic matter: 20–50 ppb QSE; with specific ultraviolet absorbance at 254 nm 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). Precipitation is over 3,000 mm yr\(^{-1}\), and typhoon precipitation contributes up to half of the total precipitation in YYL annually (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 over turns in September (Kimura et al., 2012; Kimura et al., 2017; Lin et al., 2021). The concentrations of DIC, DOC (Lin et al., 2021), total nitrogen, total phosphate (Chiu et al., 2020; Tsai et al., 2008) and bacteria compositions (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 it 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. 1). From January 2015 to December 2018, we measured the water surfaces for six river inflows and one outflow each month using a horizontal van Dorn bottle (2.20 L, acrylic) (Fig. 1). These liquid 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 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 refrigerated 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), with 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 using a portable fluorometer (model 10-AU-005-CE; Turner Designs, Sunnyvale, CA, USA), with specific wavelengths were 430 nm (blue) and 662 nm (red). All analysis was completed within 72 hours of exposure to light to reduce the degradation. 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 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, to simulate the DIC and DOC concentration under extreme weather scenarios in YYL, Next, we developed a conceptual equations model to generate continuous DIC and DOC data at the upper and lower layers, as shown in Sect. 2.3.2. This also helped us understand the transportation, photosynthesis, and remineralization rates between seasons and between typhoon and non-typhoon years.

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). Data on rainfall (model N-68; Nippon Electric Instrument, Tokyo, Japan) and wind speed (model 03001, R.M. Young, Traverse City, MI, USA) were stored in a datalogger (model CR1000; Campbell Scientific, Logan, UT, USA) every 10 min. River discharge (Q_in, m³ 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 (Fig. 1) using the Manning formula. Transparency was estimated using Secchi disc data measured at a certain interval in that time frame local times (GMT+08:00) from 10:00 to 14:00 (GMT+08:00).

As Table 1 shows, four strong typhoons were recorded by using wind speed and rainfall meteorological parameters, contributing a total of 2,254 mm of precipitation in all 24 months of 2015 and 2016. This accounted for 35.6% of across two years of typical rainfall (> 3000 mm yr⁻¹) 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 2,571,398 mm yr⁻¹, which were below the half of average years. The annual average wind speed during from 2015 to 2016 was higher by 0.09 m s⁻¹ than in 2017 and 2018 (Table 1). Despite there was no significant difference in average water depth between 2017 and 2018 (Table 1), the average discharge in 2015 and 2016 was less higher than 774 m³ d⁻¹ in 2017 and 2018 (Table 1). 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). The phytoplankton and remineralization effects on DIC and DOC fluxes (dDIC/dt and dDOC/dt, mg-C L⁻¹ d⁻¹) were considered in a conceptual two-layer equation model as shown in Equations 1–4. The fluxes in the upper layer (from the water surface to 2.5 m water depth) were calculated as follows:

\[
\frac{d \text{DIC}_U}{dt} = Q_U \Delta \text{DIC}_R - Q_{out} \Delta \text{DIC}_U - V_U \alpha_{PU} \text{Chl}_U + V_U \alpha_{MU} \text{DOC}_U + A_U \text{w}_i (\text{DIC}_L - \text{DIC}_U) \]

\[
+ Q_L \text{DIC}_L - \frac{A_F \text{CO}_2}{C_U} + P \text{a}_U \quad (1)
\]

\[
\frac{d \text{DOC}_U}{dt} = Q_U \Delta \text{DOC}_R - Q_{out} \Delta \text{DOC}_U - V_U \alpha_{MU} \text{DOC}_U + A_U \text{w}_i (\text{DOC}_L - \text{DOC}_U) \]

\[
+ Q_L \text{DOC}_L + P \text{b}_U \quad (2)
\]

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

\[
\frac{d \text{DIC}_L}{dt} = Q_L \Delta \text{DIC}_R - V_L \alpha_{PL} \text{Chl}_L + V_L \alpha_{ML} \text{DOC}_L + A_L \text{w}_i (\text{DIC}_U - \text{DIC}_L) - Q_L \text{DIC}_L \]

\[
+ \frac{A_P BF_{DIC}}{C_U} + P \text{a}_L \quad (3)
\]
\[ V_L \frac{d\text{DOC}_L}{dt} = Q_L \text{DOC}_R - V_L \alpha_m \text{DOC}_L + A_I w_I (\text{DOC}_U - \text{DOC}_L) - Q_L \text{DOC}_L + P b_L \]  

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

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

where, as shown in Table 2, total lake volume \((V_{total}, 53,544 \text{ m}^3)\) comprises 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 the lake surface area \((A_s)\) is 36,000 m² and the bottom of the 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 varied from 4.56 to 4.66 m during the typhoon period (Chiu et al., 2020; Lin et al., 2022). Therefore, we can assume that the changes in lake volumes and areas were negligible. The coefficient \(C_U\), with a value of 1000, was used to establish a standard unit for \(F_{\text{CO}_2}\) (mg-C m⁻² d⁻¹), considering the air–water CO₂ exchange by Fick’s law as follows:

\[ F_{\text{CO}_2} = k_{\text{CO}_2} \cdot K_H (p_{\text{CO}_2\text{water}} - p_{\text{CO}_2\text{air}}) \]  

where \(k_{\text{CO}_2}\) 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). \(p_{\text{CO}_2\text{air}}\) (μatm) is the CO₂ partial pressure in the atmosphere using air pressure data (Lin et al., 2021; Lin et al., 2022), and the atmospheric CO₂ concentration is assumed to be 400 ppm. \(p_{\text{CO}_2\text{water}}\) (μatm) is the CO₂ partial pressure at the water surface around 0.04 m water depth from water quality data (temperature, pH, and DIC concentrations at the water surface). The empirical equation (Cai and Wang, 1998) was also followed by Lin et al. (2021). \(F_{\text{CO}_2}\) 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 \(BF_{\text{DIC}}, BF_{\text{DO}_2}, \text{mg-C L}^{-1}\) in the lower layer was considered (Lin et al., 2022).

We assumed that the river discharge and outflow discharge \((Q_{\text{in}}, \text{m}^3 \text{ d}^{-1})\) are in a quasi-steady state \((Q_{\text{in}} = Q_{\text{out}})\), divided into upper discharge \((Q_U, \text{m}^3 \text{ d}^{-1})\) and lower discharge \((Q_L, \text{m}^3 \text{ d}^{-1})\) (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 $Q_{in}$ (Table 2).

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). Thus, we attempted to simulate extreme climate scenarios; we shifted the ratio of $Q_{in}$ for each season and tested the river intrusion hypothesis. 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), and 50% (winter) of $Q_{in}$ (Table 2).

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 $P_{aU}$, $P_{aL}$, $P_{bU}$, and $P_{bL}$ are constants in the conceptual models. To obtain unknown values ($\alpha_{PU}$, $\alpha_{MU}$, $\alpha_{PL}$, $\alpha_{ML}$, $w_i$, $BF_{DIC}$, $BF_{DOC}$, $P_{aU}$, $P_{aL}$, $P_{bU}$, and $P_{bL}$), 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). 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) (Nash and Sutcliffe, 1970) to quantify the performance of the equation model with DIC and DOC sampling data (observation data) for each simulation as follows.

$$NSE = 1 - \frac{\sum_{i=1}^{n} (Obs_i - Sim_i)^2}{\sum_{i=1}^{n} (Obs_i - \bar{Obs})^2}$$  \hspace{1cm} (8)

where $Obs$ is observation data of DIC and DOC concentrations, and $Sim$ is best-fit data for conceptual model.

**2.3.3 DIC and DOC fluxes**

Net 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. 1), 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 Equations 9–12. The upper layer NEP of DIC flux (mg C d⁻¹) was obtained from Equation 1 as follows:

Upper flux \(_{\text{DIC}}\) = \(C_U \alpha_{PU} Chl_U - C_U \alpha_{MU} DOC_U - C_U A_{lwf} (\text{DIC}_L - \text{DIC}_U) - C_U \frac{Pa_U}{V_U} - C_U \frac{A_{lwf} DIC_U}{V_U} \)

\(= C_U \frac{Q_U DIC_R + Q_L DIC_L - Q_{out} DIC_U - A_s F_{CO_2}}{V_U} \)

\(= C_U \frac{1}{t_{ru}} \left( \frac{Q_U}{Q_{in}} DIC_R + \frac{Q_L}{Q_{in}} DIC_L - DIC_U \right) - F_c \)

\(\frac{t_{ru}}{t_{ru}} = \frac{V_U}{Q_{in}} \)

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

Upper flux \(_{\text{DOC}}\) = \(C_U \alpha_{MU} DOC_U - C_U A_{lwf} (DOC_L - DOC_U) - C_U \frac{Pb_U}{V_U} \)

\(= C_U \frac{Q_U DOC_R + Q_L DOC_L - Q_{out} DOC_U}{V_U} \)

\(= C_U \frac{1}{t_{ru}} \left( \frac{Q_U}{Q_{in}} DIC_R + \frac{Q_L}{Q_{in}} DIC_L - DOC_U \right) \)

The lower layer flux of DIC flux (mg C m⁻³ d⁻¹) was estimated from Equation 3:

Lower flux \(_{\text{DIC}}\) = \(C_U \alpha_{PL} Chl_L - C_U \alpha_{ML} DOC_L - C_U A_{lwf} (\text{DIC}_U - \text{DIC}_L) - \frac{A_{BF_{DIC}}}{V_L} \)

\(- C_U \frac{Pa_L}{V_L} = C_U \frac{Q_L (\text{DIC}_R - \text{DIC}_L)}{V_L} = C_U \frac{1}{t_{RL}} Q_{in} (\text{DIC}_R - \text{DIC}_L) \)

\(t_{RL} = \frac{V_L}{Q_{in}} \)

The lower layer flux of DOC flux (mg C m⁻³ d⁻¹) was estimated from Equation 4:

Lower flux \(_{\text{DOC}}\) = \(C_U \alpha_{ML} DOC_L - C_U A_{lwf} (\text{DOC}_U - \text{DOC}_L) - \frac{A_{BF_{DOC}}}{V_L} - C_U \frac{Pb_L}{V_L} \)

\(= C_U \frac{Q_L (\text{DOC}_R - \text{DOC}_L)}{V_L} = C_U \frac{1}{t_{RL}} Q_{in} (\text{DOC}_R - \text{DOC}_L) \)

Thus, the total flux of DIC and that of DOC are:
\begin{align*}
\text{Flux}_{\text{DIC}} &= \frac{V_U \text{Upper flux}_{\text{DIC}} + V_L \text{Lower flux}_{\text{DIC}}}{V_{\text{total}}} \\
\text{Flux}_{\text{DOC}} &= \frac{V_U \text{Upper flux}_{\text{DOC}} + V_L \text{Lower flux}_{\text{DOC}}}{V_{\text{total}}}
\end{align*}

where, \( F_C \) is \( \frac{A}{V_U} F_{CO2} \) 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 Observation Measuring data (monthly DIC, DOC, and Chl a concentrations) in typhoon and non-typhoon years

The comparisons between the two typhoon years (2015 and 2016) revealed no significant differences in DIC, DOC, and Chl a concentrations between the upper and lower layers; however, all these parameters differed significantly between the layers in the non-typhoon years 2017 and 2018 (Fig. 2). This is because of typhoon-induced mixing and lower thermal stratification between upper and lower layer (Lin et al., 2021; Lin et al., 2022). Overall, the average DIC$_U$ was 2.06 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 Chl$_U$ and Chl$_L$ were 2.13 µg-C L$^{-1}$ and 18.5 µg-C L$^{-1}$, respectively. In typhoon years, the average DIC$_U$ was 2.34 mg-C L$^{-1}$, and DIC$_L$ was 4.07 mg-C L$^{-1}$; the average DOC$_U$ was 6.10 mg-C L$^{-1}$, and DOC$_L$ was 8.38 mg-C L$^{-1}$; and the Chl$_U$ and Chl$_L$ were 2.38 µg-C L$^{-1}$ 12.2 µg-C L$^{-1}$, respectively (Fig. 2); In non-typhoon years, the average DIC$_U$ was 1.81 mg-C L$^{-1}$, and DIC$_L$ was 3.28 mg-C L$^{-1}$; the average DOC$_U$ was 5.66 mg-C L$^{-1}$, and DOC$_L$ was 7.67 mg-C L$^{-1}$; and Chl$_U$ and Chl$_L$ were 1.89 µg-C L$^{-1}$ and 24.4 µg-C L$^{-1}$, respectively (Fig. 2).

ANOVA results indicated no significant differences in DIC concentrations among seasons during the typhoon years (p-values ≥ 0.05), suggesting a lack of statistically significant variation in DIC data across seasons (Fig. 3a–b). However, the DOC concentration showed significant differences between seasons in the typhoon years (Fig. 3c–d). No significant differences between Chl$_U$ and Chl$_L$ were observed among the seasons (Fig. 3e–f), whereas the standard deviations (SD) of DIC and DOC were higher in summer and autumn (Fig. 3) due to terrestrial C loading (Chiu et al., 2020). In summer, the SD values of DIC$_U$ and DOC$_U$ were 3.51 mg-C L$^{-1}$ and 3.69 mg-C L$^{-1}$, respectively (Fig. 3a, c, e). In autumn, DIC$_L$ and DOC$_L$ had the highest SD (4.06 and 4.17 mg-C L$^{-1}$, respectively) (Fig. 3b, d). Notably, the maximums of DIC$_U$ and DOC$_U$ were 7.06 and 15.6 mg-C L$^{-1}$ and those of DIC$_L$ and DOC$_L$ were 10.9 and 19.8 mg-C L$^{-1}$, respectively, in the typhoon years (Fig. 3a-d).

Positive Pearson correlations of 0.45 to 0.80 were observed between the DOC and DIC in the typhoon years (Fig. 4a). In the non-typhoon years, the upper layer DIC$_L$ was the only variable correlated negatively with DOC in the upper and lower layers (Fig. 4b). DIC in the lower layer was positively correlated with the Chl$_L$ (Fig. 4) due to the abundant respiration in the lower layer (Lin et al., 2021; Tsai et al., 2008).
3.2 Performance of simulation data in conceptual two-layer DIC and DOC models

The results for the typhoon years demonstrated that DIC$_U$ was around 1.5 to 5.0 mg-C L$^{-1}$ (Fig. 5a–b) and DIC$_L$ was around 5.0 mg-C L$^{-1}$ (Fig. 5d). In the non-typhoon years (2017–2018), the best-fit NSE values of DIC$_U$ and DIC$_L$ were 0.6149 and 0.70, respectively. On the other hand, the DIC fit our observation data (R$^2$ values are 0.91, and 0.46; the NSE coefficients from in equation (8) are 0.95 and 0.73) (Fig. 5ce–dh, Table 3). The parameters for the conceptual two-layer DIC and DOC models showed different regimes between the typhoon and non-typhoon years (Table 3). In the typhoon years, the photosynthesis absorption rate 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 values of the transportation coefficients ($w_t$) were higher in autumn than in the other seasons (Table 3). 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.

We simulated the responses of DIC and DOC 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. 5). Overall, the C level declined in the upper layers but increased in the lower layers (Fig. 5). 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. 5a, 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. 5b, d).

Under extreme weather conditions, Level 2 usually shifted to different typhoon responses for each season (Fig. S1–S2) due to extreme river intrusions and strengths of thermal stratification (Lin et al., 2021). DIC changes more than DOC under Level 1 and Level 2 (Fig. 5), because the photosynthesis, transportation, and remineralization rates may crucially affect the seasonal and interannual patterns of DOC as well. The DIC concentration in the upper layer was significantly lower in the typhoon years than in the non-typhoon years during spring and autumn under Level 2 (Fig. 7a–c). Under the best fit and Level 1 conditions, the DIC concentrations decreased significantly from winter to spring (Fig. 7e–d). The lower layer DIC values under the best fit and Level 1 conditions differed significantly between the typhoon and non-typhoon years (Fig. 7e–h). The DIC of the lower layer under Level 2 differed significantly from winter to spring only (Fig. 7e, h). The highest averaged DIC in the lower layer was 10 mg-C L$^{-1}$ under the Level 2 in spring in non-typhoon years (Fig 7e).

The upper layer DOC showed significant typhoon responses for each condition from winter to spring (Fig. 8a, d). The DOC of the upper layer tended to differ more significantly...
under the extreme climates from summer to autumn (Fig. 8b–c). The DOC of the lower layer showed different typhoon responses between spring and the other seasons (Fig. 8e–h), and the DOC values in the lower layer under Level 2 conditions in spring showed no significant difference between the typhoon and non-typhoon years (Fig 8e).
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⁻² d⁻¹ of DIC and 415 mg C m⁻² d⁻¹ of DOC during the typhoon years; during the non-typhoon years, it released 51.7 mg C m⁻² d⁻¹ of DIC and 22.8 mg C m⁻² d⁻¹ of DOC (Table 4). The average $F_C$ was one to two times larger than $F_{DOC}$, 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 exhibited declines in both DIC and DOC concentrations, with the decline in DIC being declined approximately 3.7 times higher than in the non-typhoon years (Table 4). In autumn in the typhoon years, 216 mg C d⁻¹ of upper layer DIC was released, but 46.1 mg C m⁻³ d⁻¹ of upper layer DOC was produced. The upper layer $F_{DOC}$ was negative in autumn in the typhoon years, when 268 mg C m⁻³ d⁻¹ more $F_C$ was released compared to autumn in the non-typhoon years. In addition, the lower layer exhibited the largest release of C into the outflow in the typhoon years; however, the flux in the lower layer was more than twice as high in the summer as in the autumn of those years (Table 4). The average total $F_{DIC}$ exhibited a release of approximately 3.14 times more C in the typhoon years than in the non-typhoon years. The average total $F_{DOC}$ showed an increased of 62.3 mg C m⁻³ d⁻¹ of DOC between the typhoon years and non-typhoon years due to the over 10-fold higher flux in the upper layer (Table 4).
3.34 Interannual responses of \(C_{fluxes}\) in NEP, DIC, and DOC to typhoons under extreme weather scenarios

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. 6–8). Overall, the C level declined in the upper layers but increased in the lower layers (Fig. 6). 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. 6b, d).

The DIC concentration in the upper layer was significantly lower in the typhoon years than in the non-typhoon years during spring and autumn under Level 2 (Fig. 7a–c). Under the best-fit and Level 1 conditions, the DIC concentrations decreased significantly from winter to spring (Fig. 7e–d). The lower layer DIC values under the best-fit and Level 1 conditions differed significantly between the typhoon and non-typhoon years (Fig. 7e–h). The DIC of the lower layer under Level 2 differed significantly from winter to spring only (Fig. 7e, h). The highest-averaged DIC in the lower layer—was 10 mg C L\(^{-1}\) under the Level 2 in spring in non-typhoon years (Fig. 7e).–

The upper layer DOC showed significant typhoon responses for each condition from winter to spring (Fig. 8a–d). The DOC of the upper layer tended to differ more significantly under the extreme climate from summer to autumn (Fig. 8b–c). The DOC of the lower layer showed different typhoon responses between spring and the other seasons (Fig. 8a–h), and the DOC values in the lower layer under Level 2 conditions in spring showed no significant difference between the typhoon and non-typhoon years (Fig. 8c).

We used filed observations data (Fig 5) to estimate the C fluxes in Table 4. 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 fluxes (Table 4). The average \(F_C\) was 219 and 133 mg C m\(^{-3}\) d\(^{-1}\) released from YYL into the atmosphere in the typhoon and non-typhoon years, respectively, which were higher one to two times larger than \(Flux_{DIC}\) (Table 4). In summer, the upper layer exhibited declines in both DIC and DOC concentrations, with the decline in DIC being decreased approximately 3.7 times higher in DIC in the typhoon years than in the non-typhoon years (Table 4). In autumn in the typhoon years, 216 mg C d\(^{-1}\) of upper layer DIC was released in the typhoon years, but 46.1 mg C m\(^{-3}\) d\(^{-1}\) of upper layer DOC was produced. The upper layer \(Flux_{DIC}\) was negative in autumn in the typhoon years, when 268 mg C m\(^{-3}\) d\(^{-1}\) more \(F_C\) was released compared to autumn in the non-typhoon years. In addition, the lower layer exhibited the largest release...
of C into the outflow in the typhoon years; however, the flux in the lower layer was more than
twice as high in the summer as in the autumn of those years (Table 4). The average total FluxDIC
exhibited was a release of approximately 3.14 times more C in the typhoon years than in the
non-typhoon years. The average total NEPDOC showed an increase of 62.3 mg C m$^{-3}$ d$^{-1}$ of
DOC between the typhoon years and non-typhoon years due to the over 10-fold higher flux in
the upper layer (Table 4).

Under extreme weather conditions, Level 2 usually shifted to different typhoon
responses for each season (Fig. 7–8) due to extreme river intrusions. DIC changes more
significantly than DOC under Level 1 and Level 2 (Fig. 6–8), because the photosynthesis,
transportation, and remineralization rates may crucially affect the seasonal and interannual
patterns of DOC as well. Moreover, we compared the fluxes with different model conditions
(Best-fit, Level 1, and Level 2) as shown in Fig. 9, demonstrating that the responses of FluxDIC
to typhoons differed dramatically between Level 1 and Level 2 (Fig. 9a–c); especially, the
Upper FluxDIC released more C in the typhoon years and absorbed more C in the non-typhoon
years than Obs (Fig. 9a). Not only were the absolute values of FluxDIC over 3 times higher in
the typhoon years than in the non-typhoon years (Table 4), but SD was higher in the typhoon
years as well (Fig. 9b). However, DOC fluxes changed less under Level 1 and Level 2 (Fig.
9d–f), a finding that is consistent with our continuous DOC data (Fig. 6c–d).

We not only attempted to know the contributions of thermal stratification and river
intrusion to DIC and DOC fluxes by using Leve 1 and Leve 2 scenarios in this conceptual
model, but also the contributions of typhoon disturbances to DIC and DOC fluxes (Fig. 7). We
found the DIC and DOC fluxes were both released (ΔFluxDIC and ΔFluxDOC) under Leve 1 and
Leve 2 scenarios (Fig. 7). Typhoon disturbances contributed -102.2, -62.3 mg-C m$^{-3}$ d$^{-1}$ FluxDIC
and FluxDOC in measurements (observation) data (Obs), respectively (Fig. 7). The averages of
ΔFluxDIC under Best-fit was declined 7.0 % (Leve 1) and 30.0 % (Leve 2); ΔFluxDOC was
declined 56.9 % (Leve 1) and 118 % (Leve 2), respectively (Fig. 7).
4. Discussion

4.1 Biochemical and physical differences of DIC and DOC fluxes between typhoon and non-typhoon-years in YYL

Annual The total precipitation was 35.640% higher in the typhoon years than in the 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). The absence of typhoon-induced upwelling affected water quality data differences between the upper and lower layers (Chiu et al., 2020; Lin et al., 2022; Tsai et al., 2008; Tsai et al., 2011). DIC, DOC, and Chl. a concentrations differed significantly between upper and lower layers in the non-typhoon years (Fig. 2). Further, the abundance of microorganisms leads to intensive respirations in the lower layers during the non-typhoon period in YYL; for example, an anoxic condition at the hypolimnion may decrease the efficiency of 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). Thus, 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. 5).

We found positive correlations significantly between DOC and DIC concentrations in typhoon years (Fig. 4); because substantial amounts of C loading into YYL during the strong typhoon period in 2015 (Fig. 8a–b) due to the last drought year in 2014 (Chiu et al., 2020). Not only the prolonged (or hysteresis) effects might lead to dramatically C emissions dramatically in 2015 due to the drought year in 2014 (Chiu et al., 2020), but also the rapid water retention and C loading during strong typhoon periods (Lin et al., 2022) induced vigorous algal biomass production in the lower layer from May to September 2017 (Fig. 8d). Conversely, without the typhoon-induced mixing and refreshing of the water column might be the mineralization dominated the DIC and DOC concentrations in non-typhoon years (Chiu et al., 2020; Hanson et al., 2015; Lin et al., 2022; Vachon et al., 2021), which could result in the positive linear relationship between DIC and DOC concentrations (Fig. 4b, 8a), and negative remineralization rates in non-typhoon years (Table 3). Therefore, these hydraulic, biogeochemical processes, and the hysteresis effects might describe different patterns of measurements data between the typhoon years and non-typhoon years.

Thermal stratification and allochthonous C loading may drive the responses of fluxes to typhoons in YYL. The absolute values of fluxes were higher in the typhoon years 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 explain the higher fluxes in autumn compared to the other seasons (Table 4). Typhoons dramatically changed the seasonal and interannual patterns of DIC fluxes due to river intrusion (Fig. 6a5a–b; Fig. 7S1), which
Because of the absence of typhoon-induced mixing and allochthonous C loading, the total fluxes were lower in the non-typhoon years than those in the typhoon years (Table 4). In the typhoon years, our results showed that typhoon-induced upwelling and loading increased by 102.2 mg-DIC m\(^{-3}\) d\(^{-1}\) and 62.3 mg-DOC m\(^{-3}\) d\(^{-1}\) in YYL (Table 4). Additionally, the CO\(_2\) emission (\(F_C\)) was 43 % higher (~83 mg C m\(^{-3}\) d\(^{-1}\)) in the typhoon years than in the non-typhoon years (Table 4). Therefore, typhoon disturbances control DIC loading and C emissions in YYL, consistent with our previous studies (Chiu et al., 2020; Lin et al., 2021; Lin et al., 2022).

Simultaneously, 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), resulting in the higher light intensity and water temperature in summer consuming 3.7 times more DIC and DOC than in the other seasons (Table 4). These results suggested that the allochthonous C loading and light duration might be the most crucial factor for DIC and DOC fluxes in the typhoon years. Conversely, the transportation rate shaped the seasonal C concentrations due to thermal stratification in the non-typhoon years.

### 4.2 Model limitation under the extreme weather scenarios

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 thermal stratification strength (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 (\(w_I\)) well represented the seasonal variations in 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, mineralization dominated the DIC and DOC concentrations in non-typhoon years, anoxic...
conditions may occur (Chiu et al., 2020; Hanson et al., 2015; Lin 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 characterize the seasonal and interannual responses of DIC and DOC fluxes to typhoons in YYL.

Under the extreme weather events scenarios (Level 1 and Level 2), the DIC and DOC are more released 30%–118% within YYL, considering thermal stratification and river intrusion. (Fig. 7). In non-typhoon years, the DIC fluxes were more absorbed than DOC fluxes (Fig. 6) or events for each season (Fig. S1–S2) under extreme weather events scenarios. However, the results showed that the typhoon disturbances that impacted the response of DOC fluxes were more sensitive than DIC fluxes (Fig. 7). In autumn, our results are shown that parameter of transportation rates in typhoon years was over ten-fold higher than non-typhoon years (Table 3), but also DOC concentration was dominated by mineralization in typhoon years (Fig. 8) because biogeochemical processes within lakes, such as respiration, mineralization, and sediment burial, may impact DOC fluxes (Bartosiewicz et al., 2015; Hanson et al., 2015; Maranger et al., 2018). Simultaneously, the physical processes such as typhoon-induced mixing and fall overturns upwelled the sediment and lower layer DOC into the water surface in YYL (Kimura et al. 2017; Lin et al. 2022). Therefore, we suggested that fall overturns and mineralization after typhoon disturbances might be vital in the vertical distribution of DOC concentrations in typhoon years.

Moreover, the DIC fluxes were more absorbed than DOC fluxes in non-typhoon years (Fig. 6), or event for each season (Fig S1–S2) under Level 1 and Level 2 scenarios. The interannual variability of DOC fluxes might more rely on the mineralization rate in non-typhoon years (Fig 7 a–b) than thermal stratification and river intrusion. However, biogeochemical processes within lakes, such as respiration, mineralization, and sediment burial, may impact DOC fluxes (Bartosiewicz et al., 2015; Hanson et al., 2015; Maranger et al., 2018). Simultaneously, the physical processes such as typhoon-induced mixing and fall overturns upwelled the sediment and lower layer DOC into the water surface in YYL (Kimura et al. 2017; Lin et al. 2022). Therefore, we suggested that fall overturns and mineralization after typhoon disturbances might be vital in the vertical distribution of DOC concentrations in typhoon years.
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 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

We successfully developed two-layer conceptual C models to obtain continuous DIC and DOC data in YYL and to simulate extreme conditions. Our conceptual two-layer C model revealed that allochthonous and autochthonous processes both accounted for C flux responses to typhoon disturbances. Without typhoons, thermal stratification was the primary driver of seasonal and interannual patterns of DIC and DOC. In the typhoon years, the changes in seasonal river intrusion regimes in YYL resulted in a 3-fold higher total Flux$_{DIC}$ than in the non-typhoon years. However, our model should be improved for application to extreme climate scenarios by considering other autochthonous processes within lake, such as sediment burial, photo-degradation, 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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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><strong>Annual average wind speed (m s⁻¹)</strong></td>
<td><strong>1.20</strong></td>
<td><strong>1.11</strong></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 (m³ d⁻¹)</td>
<td>3,717</td>
<td>2,943</td>
</tr>
<tr>
<td>Transparency</td>
<td>1.58 ± 0.45</td>
<td>1.38 ± 0.28</td>
</tr>
<tr>
<td>(Secchi disc depth, m ± SD)</td>
<td></td>
<td></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>Outflow discharge</td>
<td>Daily data</td>
<td>m³ d⁻¹</td>
</tr>
<tr>
<td>Inflow discharge</td>
<td>Daily data</td>
<td>m³ d⁻¹</td>
</tr>
<tr>
<td>Upper layer discharge</td>
<td>Daily data</td>
<td>m³ d⁻¹</td>
</tr>
<tr>
<td>Lower layer discharge</td>
<td>Daily data</td>
<td>m³ d⁻¹</td>
</tr>
<tr>
<td>River inflow DIC</td>
<td>Monthly data</td>
<td>mg-C L⁻¹</td>
</tr>
<tr>
<td>Upper layer DIC</td>
<td>Monthly data</td>
<td>mg-C L⁻¹</td>
</tr>
<tr>
<td>Lower layer DIC</td>
<td>Monthly data</td>
<td>mg-C L⁻¹</td>
</tr>
<tr>
<td>Upper layer Chl a</td>
<td>Monthly data</td>
<td>mg L⁻¹</td>
</tr>
<tr>
<td>Lower layer Chl a</td>
<td>Monthly data</td>
<td>mg L⁻¹</td>
</tr>
<tr>
<td>Upper layer DOC</td>
<td>Monthly data</td>
<td>mg-C L⁻¹</td>
</tr>
<tr>
<td>Lower layer DOC</td>
<td>Monthly data</td>
<td>mg-C L⁻¹</td>
</tr>
<tr>
<td>Carbon emission (equation 7)</td>
<td>Monthly data</td>
<td>mg-C m² d⁻¹</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Measurements</th>
<th>Value</th>
<th>Unit</th>
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</thead>
<tbody>
<tr>
<td>Total lake volume</td>
<td>53,544</td>
<td>m³</td>
</tr>
<tr>
<td>Upper layer volume</td>
<td>45,456</td>
<td>m³</td>
</tr>
<tr>
<td>Lower layer volume</td>
<td>8,808</td>
<td>m³</td>
</tr>
<tr>
<td>Lake surface area</td>
<td>36,000</td>
<td>m²</td>
</tr>
<tr>
<td>Interface area</td>
<td>7,264</td>
<td>m²</td>
</tr>
<tr>
<td>Bottom of lake area</td>
<td>3,520</td>
<td>m²</td>
</tr>
<tr>
<td>Coefficient of the standard unit</td>
<td>1,000</td>
<td>L m⁻³</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Unknown Constants</th>
<th>Value</th>
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<tbody>
<tr>
<td>Coefficients of photosynthesis</td>
<td>Constant</td>
<td>d⁻¹</td>
</tr>
<tr>
<td>Coefficients of mineralization</td>
<td>Constant</td>
<td>d⁻¹</td>
</tr>
<tr>
<td>Coefficient of vertical transportation</td>
<td>Constant</td>
<td>d⁻¹</td>
</tr>
<tr>
<td>Sediment DIC and DOC emission</td>
<td>Constant</td>
<td>mg-C L⁻¹</td>
</tr>
<tr>
<td>Equations constant at lower layer</td>
<td>Constant</td>
<td>mg m⁻³ d⁻¹</td>
</tr>
</tbody>
</table>

**Extreme scenarios**

- Q<sub>U</sub> and Q<sub>L</sub> are followed buoyancy frequency for each season (Lin et al. 2021): 75% (spring), 80% (summer), 45% (autumn), 50% (winter)
- Q<sub>U</sub> and Q<sub>L</sub> are followed buoyancy frequency for each season (Lin et al. 2021): 25% (spring), 20% (summer), 55% (autumn), 50% (winter)
- Q<sub>U</sub> and Q<sub>L</sub> are followed buoyancy frequency for each season (Lin et al. 2021): 75% (spring), 80% (summer), 45% (autumn), 50% (winter)
- Q<sub>U</sub> and Q<sub>L</sub> are followed buoyancy frequency for each season (Lin et al. 2021): 25% (spring), 20% (summer), 55% (autumn), 50% (winter)
<table>
<thead>
<tr>
<th>Level 1</th>
<th>Best-fit scenario but change upper- and lower-layers discharges ((Q_U, Q_L))</th>
<th>(77%) (spring), (82%) (summer), (47%) (autumn), (50%) (winter) of (Q_{in})</th>
<th>(23%) (spring), (18%) (summer), (53%) (autumn), (50%) (winter) of (Q_{in})</th>
</tr>
</thead>
<tbody>
<tr>
<td>Level 2</td>
<td>Best-fit scenario but change upper and lower layers discharges ((Q_U, Q_L))</td>
<td>(80%) (spring), (85%) (summer), (50%) (autumn), (50%) (winter) of (Q_{in})</td>
<td>(20%) (spring), (15%) (summer), (50%) (autumn), (50%) (winter) of (Q_{in})</td>
</tr>
</tbody>
</table>
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>
<th></th>
<th>2015‒2016</th>
<th>2017‒2018</th>
<th></th>
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<td>Spring</td>
<td>Summer</td>
<td>Autumn</td>
<td>Winter</td>
<td>Spring</td>
<td>Summer</td>
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<td>Winter</td>
<td>Spring</td>
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<tr>
<td><strong>Upper layer</strong></td>
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<tr>
<td>$F_{CO_2}$ (mg-C m$^{-2}$ d$^{-1}$)</td>
<td>291</td>
<td>245</td>
<td>422</td>
<td>127</td>
<td>231</td>
<td>143</td>
<td>104</td>
<td>175</td>
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<tr>
<td>$\alpha_{PU}$ (d$^{-1}$)</td>
<td>-1.20</td>
<td>-33.1</td>
<td>-183.5</td>
<td>-29.1</td>
<td>8.0</td>
<td>6.0</td>
<td>30.0</td>
<td>7.77</td>
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<td>$\alpha_{MU}$ (d$^{-1}$)</td>
<td>-0.0227</td>
<td>0.0203</td>
<td>0.08</td>
<td>-0.031</td>
<td>-0.01</td>
<td>-0.039</td>
<td>-0.033</td>
<td>-0.195</td>
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<tr>
<td>$w_l$ (d$^{-1}$)</td>
<td>0.230</td>
<td>0.172</td>
<td>1.38</td>
<td>0.30</td>
<td>0.10</td>
<td>0.0478</td>
<td>0.120</td>
<td>0.180</td>
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<tr>
<td>$Pa_{U}$ (d$^{-1}$)</td>
<td>12560</td>
<td>-1317</td>
<td>-23750</td>
<td>9597</td>
<td>9880</td>
<td>14000</td>
<td>17600</td>
<td>10100</td>
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<tr>
<td>$Pb_{U}$ (d$^{-1}$)</td>
<td>-21930</td>
<td>9461</td>
<td>-42130</td>
<td>-17070</td>
<td>-3630</td>
<td>-1251</td>
<td>-20820</td>
<td>-9289</td>
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<tr>
<td>$dDIC_{U}$ (R$^2$, NSE)</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>0.305, 0.614</td>
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<td>$dDOC_{U}$ (R$^2$, NSE)</td>
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<td></td>
<td></td>
<td>0.909, 0.953</td>
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<td><strong>Lower layer</strong></td>
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<td>$\alpha_{PL}$ (d$^{-1}$)</td>
<td>-0.627</td>
<td>-22.1</td>
<td>15.0</td>
<td>-0.878</td>
<td>1.49</td>
<td>-6.87</td>
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<td>$\alpha_{ML}$ (d$^{-1}$)</td>
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<td>0.123</td>
<td>0.0755</td>
<td>0.00973</td>
<td>-0.010</td>
<td>-0.0376</td>
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<td>$Pa_{L}$ (d$^{-1}$)</td>
<td>100</td>
<td>-5662</td>
<td>-10500</td>
<td>-1013</td>
<td>151.6</td>
<td>2032</td>
<td>1216</td>
<td>909</td>
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<tr>
<td>$Pb_{L}$ (d$^{-1}$)</td>
<td>-6012</td>
<td>-7395</td>
<td>-53940</td>
<td>-9639</td>
<td>-1338</td>
<td>-6296</td>
<td>-19470</td>
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<td>$BF_{DIC}$, $BF_{DOC}$ (mg-C L$^{-1}$)</td>
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<td></td>
<td></td>
<td>0.04,</td>
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<td>$dDIC_{L}$ (R$^2$, NSE)</td>
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<td></td>
<td></td>
<td>0.452, 0.707</td>
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<tr>
<td>$dDOC_{L}$ (R$^2$, NSE)</td>
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<td></td>
<td></td>
<td></td>
<td>0.460, 0.728</td>
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</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, and negative ones show the values after carbon was released. \(F_C\) was carbon emission across water to air by using empirical equations method (Lin et al.2021).

<table>
<thead>
<tr>
<th>Season</th>
<th>Average</th>
<th>Flux (F_C) (mg C m(^{-3}) d(^{-1}))</th>
<th>Total (\text{Flux}_{\text{DIC}}) (mg C m(^{-3}) d(^{-1}))</th>
<th>(\text{Flux}_{\text{DOC}}) (mg C m(^{-3}) d(^{-1}))</th>
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<tr>
<td>Typhoon years</td>
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<td>Spring</td>
<td></td>
<td>-219&lt;br&gt;DIC: -231&lt;br&gt;DOC: -70.8&lt;br&gt;FluxDIC: -210&lt;br&gt;FluxDOC: 62.1</td>
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<tr>
<td>Summer</td>
<td></td>
<td>-194&lt;br&gt;DIC: -194&lt;br&gt;DOC: 118&lt;br&gt;FluxDIC: -26.4&lt;br&gt;FluxDOC: 18.8</td>
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<tr>
<td>Autumn</td>
<td></td>
<td>-351&lt;br&gt;DIC: -351&lt;br&gt;DOC: -46.1&lt;br&gt;FluxDIC: -288&lt;br&gt;FluxDOC: -151</td>
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<td>Winter</td>
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<td>-100&lt;br&gt;DIC: -100&lt;br&gt;DOC: 40.5&lt;br&gt;FluxDIC: -74.8&lt;br&gt;FluxDOC: 31.2</td>
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<td>Non-typhoon years</td>
<td></td>
<td>-133&lt;br&gt;DIC: -129&lt;br&gt;DOC: -21.4&lt;br&gt;FluxDIC: -166&lt;br&gt;FluxDOC: -7.06</td>
<td></td>
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<td>Spring</td>
<td></td>
<td>-183&lt;br&gt;DIC: -183&lt;br&gt;DOC: 5.80&lt;br&gt;FluxDIC: -4.57&lt;br&gt;FluxDOC: 73.8</td>
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<tr>
<td>Summer</td>
<td></td>
<td>-82.6&lt;br&gt;DIC: -82.6&lt;br&gt;DOC: -155&lt;br&gt;FluxDIC: 85.5&lt;br&gt;FluxDOC: 95.9</td>
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<tr>
<td>Autumn</td>
<td></td>
<td>-138&lt;br&gt;DIC: -138&lt;br&gt;DOC: 34.0&lt;br&gt;FluxDIC: -106&lt;br&gt;FluxDOC: 33.7</td>
<td></td>
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<tr>
<td>Winter</td>
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</table>
Fig. 1. Sampling locations and bathymetry maps of Yuan‒Yang Lake (YYL). The dark green rectangle shows the buoy station, which is located 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. 2. 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. Bullet points show the water sampling data. We used a t-test to obtain the p-values (blue texts). Total sampling numbers are 41 (n = 41) for each measurement from January 2015 to December 2018; n = 20 in typhoon years, and n = 21 in non-typhoon years.
Fig. 3. 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 seasonality, we used one-way ANOVA to obtain the $p$-values. “ns”: $p$-values ≥ 0.05; * show $p$-values from 0.05 to 0.01; **: $p$-values from 0.01 to 0.001.
Fig. 4. 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 ($Chl_a$) during (a) typhoon years and (b) non-typhoon years. Black-crosses show insignificant values ($p$-values are $> 0.05$).
Fig. 5. Continuous daily DIC and DOC data at (a, b, e, f) upper layer (DIC\textsubscript{U}, DOC\textsubscript{U}) and (c, d, g, h) lower layer (DIC\textsubscript{L}, DOC\textsubscript{L}) by using conceptual equations models. The black lines show the best fit for DIC, red lines show the best fit for DOC (Table 3), and empty dots show water sampling (observation) data for each month.
Fig. 56. 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 as shown in Table 3, and gray dots show water sampling (observation) data for each month ($n = 41$) from January 2015 to December 2018. Orange regions show Level 1; pink regions show Level 2.
Fig. 7. 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. 3. 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. Empty dots show the continuous DIC and DOC data. “ns”: p-values $\geq 0.05$; *: p-values from 0.05 to 0.01; **: p-values from 0.01 to 0.001; ****: p-values less than 0.0001 based on a t-test.
Fig. 8. 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. 7. The Fit (blue boxes) condition shows the best-fit data by using the conceptual two-layer carbon model; Lvl (yellow boxes) and Lvl2 (red boxes) show the extreme climates. Empty dots show the continuous DIC and DOC data. “ns”: p-values ≥ 0.05; *: p-values from 0.05 to 0.01; **: p-values from 0.01 to 0.001; ****: p-values less than 0.0001 based on a t-test.
Fig. 6. Interannual (a) Upper $\text{flux}_{\text{DIC}}$, (b) Lower $\text{flux}_{\text{DIC}}$, (c) $\text{Flux}_{\text{DIC}}$, (d) Upper $\text{flux}_{\text{DOC}}$, (e) Lower $\text{flux}_{\text{DOC}}$, and (f) $\text{Flux}_{\text{DOC}}$ (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. 5. The Fit condition (blue- boxes) shows the best-fit data by using the conceptual two-layer carbon model as in Fig.5. Level 1 (yellow boxes) and Level 2 (red boxes) show the extreme scenarios as in Fig. 6. For the definitions of fluxes please see Sect. 2.3.3. Positive values are shown in the carbon sink, and negative ones show the values after carbon was released within YYL.
Fig. 7. Contributions of typhoon disturbances to (a) DIC and (b) DOC fluxes (mg-C m⁻³ d⁻¹).

Contributions of typhoons to C fluxes (ΔFluxDIC, ΔFluxDOC) represented the intervals of FluxDIC in typhoon yeas and non-typhoon years (e.g., ΔFluxDIC = FluxDIC in typhoon yeas - FluxDIC in non-typhoon years).
Fig. 8. Interactions of measurements data in typhoon years and non-typhoon years. The black circles: typhoon years, gray triangles: non-typhoon years. The solid lines represent the linear regression line in typhoon years (black lines) and non-typhoon years (gray lines).