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, 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).

The responses of C fluxes in small lakes (lake area < 1 km^2) 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). 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). 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 (pCO_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 pCO_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 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., 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). 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 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 overturns 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 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 refrigerated in opaque bottles at around -25 °C in a refrigerator until the samples were analyzed. 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.

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, 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 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}, 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 (Fig. 1) using the Manning formula. Transparency was estimated using Secchi disc data measured at a certain interval in that time frame 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}). However, no typhoon rainfall was recorded at YYL in 2017 and 2018; the total precipitation was around 1,398 mm yr^{-1}, below half of average years. The annual average wind speed from 2015 to 2016 was higher by 0.09 m s^{-1} than in 2017 and 2018 (Table 1). Despite no significant difference in average water depth between 2017 and 2018, the average discharge in 2015 and 2016 was higher than 774 m^3 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 ($d\mathrm{DIC}/dt$ and $d\mathrm{DOC}/dt$, 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:

$$V_U \frac{d\mathrm{DIC}_U}{dt} = Q_U \mathrm{DIC}_R - Q_{out}\mathrm{DIC}_U - V_U \alpha_{PU} Chl_U + V_U \alpha_{MU} DOC_U + A_I w_I (\mathrm{DIC}_L - \mathrm{DIC}_U) + Q_L \mathrm{DIC}_L - \frac{A_s F_{CO2}}{C_U} + Pa_U \tag{1}$$

$$V_U \frac{d\mathrm{DOC}_U}{dt} = Q_U \mathrm{DOC}_R - Q_{out}\mathrm{DOC}_U - V_U \alpha_{MU} DOC_U + A_I w_I (\mathrm{DOC}_L - \mathrm{DOC}_U) + Q_L \mathrm{DOC}_L + Pb_U \tag{2}$$

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

$$V_L \frac{d\mathrm{DIC}_L}{dt} = Q_L \mathrm{DIC}_R - V_L \alpha_{PL} Chl_L + V_L \alpha_{ML} DOC_L + A_I w_I (\mathrm{DIC}_U - \mathrm{DIC}_L) - Q_L \mathrm{DIC}_L + \frac{A_B BF_{DIC}}{C_U} + Pa_L \tag{3}$$

$$V_L \frac{d\mathrm{DOC}_L}{dt} = Q_L \mathrm{DOC}_R - V_L \alpha_{ML} DOC_L + A_I w_I (\mathrm{DOC}_U - \mathrm{DOC}_L) - Q_L \mathrm{DOC}_L + Pb_L \tag{4}$$

$$V_{total} = V_U + V_L \tag{5}$$

$$Q_{in} = Q_U + Q_L \tag{6}$$

where, as shown in Table 2, total lake volume (V_{total}, 53,544 m^3) comprises to the upper layer (V_U, 45,456 m^3) and to the lower layer (V_L, 8,808 m^3) (Equation 5), and where the lake surface area (A_s) is 36,000 m^2 and the bottom of the lake area (A_B) is 3,520 m^2. The interface is 2.5 m vertically, and the interface area (A_I) is 7,264 m^2 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_{CO2} (mg-C m^{-2} d^{-1}), considering the air–water CO_2 exchange by Fick's law as follows:

$$F_{CO2} = k_{CO2} \cdot K_H (pCO2_{water} - pCO2_{air}) \tag{7}$$

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 (temperature, pH, and DIC concentrations). The empirical equation (Cai and 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 ($Q_{in} = Q_{out}$), divided 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} (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 (α_{PU}, d^{-1}) and lower layer (α_{PL}, d^{-1}). The coefficients of DOC remineralization rates in the upper layer (α_{MU}, d^{-1}) and lower layer (α_{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 (α_{PU}, α_{MU}, α_{PL}, α_{MU}, α_{ML}, w_I, 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). 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.

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

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:

$$\begin{aligned}
\text{Upper flux}_{\text{DIC}} &= C_U\alpha_{PU}Chl_U - C_U\alpha_{MU}DOC_U - C_U\frac{A_I w_I(\text{DIC}_L - \text{DIC}_U)}{V_U} - C_U\frac{Pa_U}{V_U} \\
&= C_U\frac{Q_U\text{DIC}_R + Q_L\text{DIC}_L - Q_{out}\text{DIC}_U}{V_U} - \frac{A_S}{V_U}F_{CO2} \\
&= C_U\frac{1}{t_{rU}}\left(\frac{Q_U}{Q_{in}}\text{DIC}_R + \frac{Q_L}{Q_{in}}\text{DIC}_L - \text{DIC}_U\right) - F_C \\
t_{rU} &= \frac{V_U}{Q_{in}}
\end{aligned} \tag{9}$$

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

$$\begin{aligned}\text{Upper flux}_{\text{DOC}} &= C_U \alpha_{MU} DOC_U - C_U \frac{A_I w_I (\text{DOC}_L - \text{DOC}_U)}{V_U} - C_U \frac{Pb_U}{V_U} \\ &= C_U \frac{Q_U \text{DOC}_R + Q_L \text{DOC}_L - Q_{out} \text{DOC}_U}{V_U} \\ &= C_U \frac{1}{t_{rU}} \left(\frac{Q_U}{Q_{in}} \text{DOC}_R + \frac{Q_L}{Q_{in}} \text{DOC}_L - \text{DOC}_U \right)\end{aligned} \tag{10}$$

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

$$\begin{aligned}\text{Lower flux}_{\text{DIC}} &= C_U \alpha_{PL} Chl_L - C_U \alpha_{ML} DOC_L - C_U \frac{A_I w_I (\text{DIC}_U - \text{DIC}_L)}{V_L} - \frac{A_B BF_{DIC}}{V_L} \\ &- C_U \frac{Pa_L}{V_L} = C_U \frac{Q_L (\text{DIC}_R - \text{DIC}_L)}{V_L} = C_U \frac{1}{t_{rL}} \frac{Q_L}{Q_{in}} (\text{DIC}_R - \text{DIC}_L) \\ t_{rL} &= \frac{V_L}{Q_{in}}\end{aligned} \tag{11}$$

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

$$\begin{aligned}\text{Lower flux}_{\text{DOC}} &= C_U \alpha_{ML} DOC_L - C_U \frac{A_I w_I (\text{DOC}_U - \text{DOC}_L)}{V_L} - \frac{A_B BF_{DOC}}{V_L} - C_U \frac{Pb_L}{V_L} \\ &= C_U \frac{Q_L (\text{DOC}_R - \text{DOC}_L)}{V_L} = C_U \frac{1}{t_{rL}} \frac{Q_L}{Q_{in}} (\text{DOC}_R - \text{DOC}_L)\end{aligned} \tag{12}$$

Thus, the total flux of DIC and that of DOC are:

$$\text{Flux}_{\text{DIC}} = \frac{V_U \text{Upper flux}_{\text{DIC}} + V_L \text{Lower flux}_{\text{DIC}}}{V_{total}} \tag{13}$$

$$\text{Flux}_{\text{DOC}} = \frac{V_U \text{Upper flux}_{\text{DOC}} + V_L \text{Lower flux}_{\text{DOC}}}{V_{total}} \tag{14}$$

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 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 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 NSE values of DIC_U and DIC_L were 0.61 and 0.70, respectively. On the other hand, the DOC 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. 5c–d, 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 (α_{PU}, α_{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_I) 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 α_{MU} and α_{ML}. In the non-typhoon years, the remineralization rates were negative (Table 3)..

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.

3.3 Interannual responses of NEP to typhoons under extreme weather scenarios

We used filed observation 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, one to two times larger than $\mathrm{Flux}_{\mathrm{DIC}}$ (Table 4). In summer, the upper layer exhibited declines in both DIC and DOC concentrations, with DIC being approximately 3.7 times higher in DIC in typhoons than in the non-typhoon years (Table 4). In autumn, 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 $\mathrm{Flux}_{\mathrm{DIC}}$ was negative in autumn in the typhoon years when 268 mg C m^{-3} d^{-1} more F_C was released than 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 $\mathrm{Flux}_{\mathrm{DIC}}$ was a release of approximately 3.14 times more C in the typhoon years than in the non-typhoon years. The average total $\mathrm{NEP}_{\mathrm{DOC}}$ showed an increase of 62.3 mg C m^{-3} d^{-1} of DOC between the typhoon and non-typhoon years due to the over 10-fold higher flux in the upper layer (Table 4).

We compared the fluxes with different model conditions (*Best-fit*, *Level 1*, and *Level 2*) as shown in Fig. 6, demonstrating that the responses of $\mathrm{Flux}_{\mathrm{DIC}}$ to typhoons differed dramatically between *Level 1* and *Level 2* (Fig. 6a-c); especially, the Upper $\mathrm{Flux}_{\mathrm{DIC}}$ released more C in the typhoon years and absorbed more C in the non-typhoon years than *Obs* (Fig. 6a). Not only were the absolute values of $\mathrm{Flux}_{\mathrm{DIC}}$ 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. 6). However, DOC fluxes changed less under *Level 1* and *Level 2* (Fig. 6d–f), a finding that is consistent with our continuous DOC data (Fig. 5c–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 ($\Delta\mathrm{Flux}_{\mathrm{DIC}}$ and $\Delta\mathrm{Flux}_{\mathrm{DOC}}$) under *Leve 1* and *Leve 2* scenarios (Fig. 7). Typhoon disturbances contributed -102.2, -62.3 mg-C m^3 d^{-1} $\mathrm{Flux}_{\mathrm{DIC}}$ and $\mathrm{Flux}_{\mathrm{DOC}}$ in measurements (observation) data (*Obs*), respectively (Fig. 7). The averages of $\Delta\mathrm{Flux}_{\mathrm{DIC}}$ under *Best-fit* was declined 7.0 % (*Leve 1*) and 30.0 % (*Leve 2*); $\Delta\mathrm{Flux}_{\mathrm{DOC}}$ 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

The total precipitation was 35.6% 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 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 measurement 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. 5a–b; Fig. S1), which proves to our hypothesis and corresponds to the results of previous studies (Chiu et al., 2020; Lin et al., 2021; Lin et al., 2022). In summer, the DOC and DIC concentrations were spatial differences between layers as a “two-layer system” within the water column because the upper and lower layers were inhibited due to strong thermal stratification (Lin et al., 2021; Lin et al., 2022), thereby resulted in the positive upper DIC and DOC fluxes and lower negative upper DIC and DOC fluxes (Table 4).

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 (α_{PU}, α_{PL}), remineralization (α_{MU}, α_{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). Thus, the conceptual two-layer C models well characterizes 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.

Ejarque et al. (2021) 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. 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 processes within lake, such as sediment burial, photo-degradation 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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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.

Records	Typhoon years	Non-typhoon years
Time period (year)	2015-2016	2017-2018
Total precipitation (mm)	6,332	3,795
Total typhoon rainfall (mm)	2,254	0
Annual average wind speed (m s^{-1})	1.20	1.11
Average water depth (m ± SD)	4.54 ± 1.7	4.51 ± 1.5
Average river discharge (m^3 d^{-1})	3,717	2,943
Transparency (Secchi disc depth, m ± SD)	1.58 ± 0.45	1.38 ± 0.28

Table 2. Parameters of the two-layer conceptual model in Yuan-Yang Lake

Parameters		Value	Unit
Measurements			
Q_{out}	Outflow discharge	Daily data	m^3 d^{-1}
Q_{in}	Inflow discharge	Daily data	m^3 d^{-1}
Q_U	Upper layer discharge	Daily data	m^3 d^{-1}
Q_L	Lower layer discharge	Daily data	m^3 d^{-1}
DIC_R	River inflow DIC	Monthly data	mg-C L^{-1}
DIC_U	Upper layer DIC	Monthly data	mg-C L^{-1}
DIC_L	Lower layer DIC	Monthly data	mg-C L^{-1}
Chl_U	Upper layer Chl *a*	Monthly data	mg L^{-1}
Chl_L	Lower layer Chl *a*	Monthly data	mg L^{-1}
DOC_U	Upper layer DOC	Monthly data	mg-C L^{-1}
DOC_L	Lower layer DOC	Monthly data	mg-C L^{-1}
F_{CO2}	Carbon emission (equation 7)	Monthly data	mg-C m^2 d^{-1}
Constants			
V_{total}	Total lake volume	53,544	m^3
V_U	Upper layer volume	45,456	m^3
V_L	Lower layer volume	8,808	m^3
A_s	Lake surface area	36,000	m^2
A_I	Interface area	7,264	m^2
A_B	Bottom of lake area	3,520	m^2
C_U	Coefficient of the standard unit	1,000	L m^{-3}
Unknow Constants			
α_{PU}, α_{PL}	Coefficients of photosynthesis	Constant	d^{-1}
α_{MU}, α_{ML}	Coefficients of mineralization	Constant	d^{-1}
w_I	Coefficient of vertical transportation	Constant	d^{-1}
BF_{DIC}, BF_{DOC}	Sediment DIC and DOC emission	Constant	mg-C L^{-1}
Pa_U, Pb_L	Equations constant at lower layer	Constant	mg m^{-3} d^{-1}
Extreme scenarios		Q_U	Q_L
Best-fit	Q_U and Q_L are followed buoyancy frequency for each season (Lin et al. 2021)	75% (spring), 80% (summer), 45% (autumn), 50% (winter) of Q_{in}	25% (spring), 20% (summer), 55% (autumn), 50% (winter) of Q_{in}.

Level 1	Best-fit scenario but change upper- and lower-layers discharges (Q_U, Q_L)	77% (spring), 82% (summer), 47% (autumn), 50% (winter) of Q_{in}	23% (spring), 18% (summer), 53% (autumn), 50% (winter) of Q_{in}.
Level 2	Best-fit scenario but change upper and lower layers discharges (Q_U, Q_L)	80% (spring), 85% (summer), 50% (autumn), 50% (winter) of Q_{in}	20% (spring), 15% (summer), 50% (autumn), 50% (winter) of Q_{in}.

Table 3. Best-fit parameters of a two-layer conceptual model of DIC and DOC in Yuan-Yang Lake from 2015 to 2018.

	2015–2016 Typhoon years				**2017–2018 Non-typhoon years**			
	Spring	*Summer*	*Autumn*	*Winter*	*Spring*	*Summer*	*Autumn*	*Winter*
<u>*Upper layer*</u>								
F_{CO2} (mg-C m^2 d^{-1})	291	245	422	127	231	143	104	175
α_{PU} (d^{-1})	-1.20	-33.1	-183.5	-29.1	8.0	6.0	30.0	7.77
α_{MU} (d^{-1})	-0.0227	0.0203	0.08	-0.031	-0.01	-0.039	-0.033	-0.195
w_I (d^{-1})	0.230	0.172	1.38	0.30	0.10	0.0478	0.120	0.180
Pa_U (d^{-1})	12560	-1317	-23750	9597	9880	14000	17600	10100
Pb_U (d^{-1})	-21930	9461	-42130	-17070	-3630	-1251	-20820	-9289
$d\mathrm{DIC}_U$ (R^2, NSE)	0.305, 0.614							
$d\mathrm{DOC}_U$ (R^2, NSE)	0.909, 0.953							
<u>*Lower layer*</u>								
α_{PL} (d^{-1})	-0.627	-22.1	15.0	-0.878	1.49	-6.87	6.0	-16.6
α_{ML} (d^{-1})	-0.025	0.123	0.0755	0.00973	-0.010	-0.0376	-0.04	-0.048
Pa_L (d^{-1})	100	-5662	-10500	-1013	151.6	2032	1216	909
Pb_L (d^{-1})	-6012	-7395	-53940	-9639	-1338	-6296	-19470	-8748
BF_{DIC}, BF_{DOC} (mg-C L^{-1})	0.04, 0.00							
$d\mathrm{DIC}_L$ (R^2, NSE)	0.452, 0.707							
$d\mathrm{DOC}_L$ (R^2, NSE)	0.460, 0.728							

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).

		Flux (mg C m^{-3} d^{-1})			Total (mg C m^{-3} d^{-1})	
		$\boldsymbol{F_C}$	**Upper**	**Lower**	**$\mathbf{Flux_{DIC}}$**	**$\mathbf{Flux_{DOC}}$**
Typhoon years	**Average**	**-219**	-	-	**-150**	**-9.69**
Spring	DIC	-231	-243	-45.2	-210	62.1
	DOC	-	70.8	17.2		
Summer	DIC	-194	29.1	-313	-26.4	18.8
	DOC	-	118	-495		
Autumn	DIC	-351	-216	-659	-288	-151
	DOC	-	46.1	-1167		
Winter	DIC	-100	-96.4	36.5	-74.8	31.2
	DOC	-	40.5	-16.9		
Non-typhoon years	**Average**	**-133**	-	-	**-47.8**	**52.6**
Spring	DIC	-129	-180	-94.9	-166	-7.06
	DOC	-	21.4	-67.1		
Summer	DIC	-183	5.80	-58.1	-4.57	73.8
	DOC	-	115	-140		
Autumn	DIC	-82.6	95.0	35.9	85.5	95.9
	DOC	-	168	-272		
Winter	DIC	-138	-128	6.04	-106	33.7
	DOC	-	34.0	32.1		

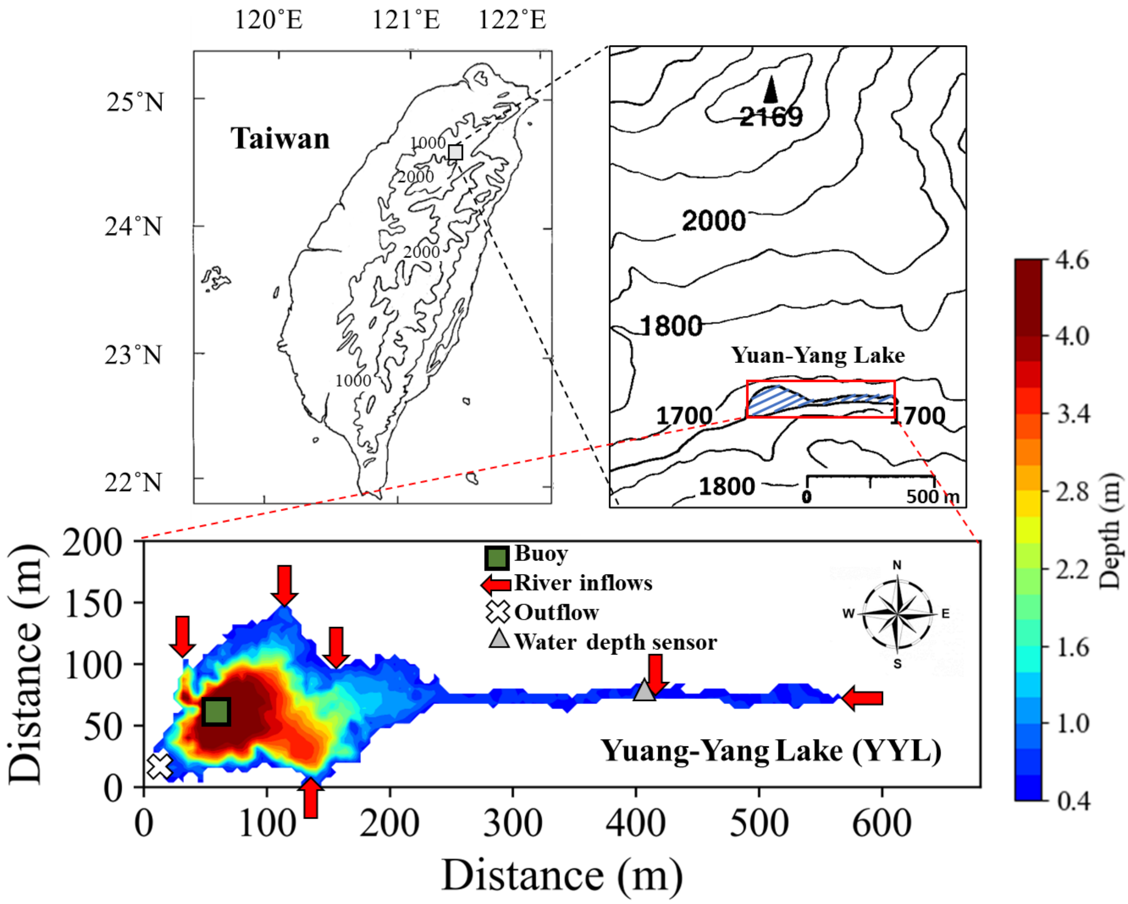

Fig. 1. Sampling locations and bathymetry maps of Yuan–Yang Lake (YYL). The dark green rectangle shows the buoy station, which is located at the deepest site of the lake. The *red points* and *white cross* show the river mouths of the inflows and outflows, respectively. The *gray triangle* shows the location of the water depth sensor.

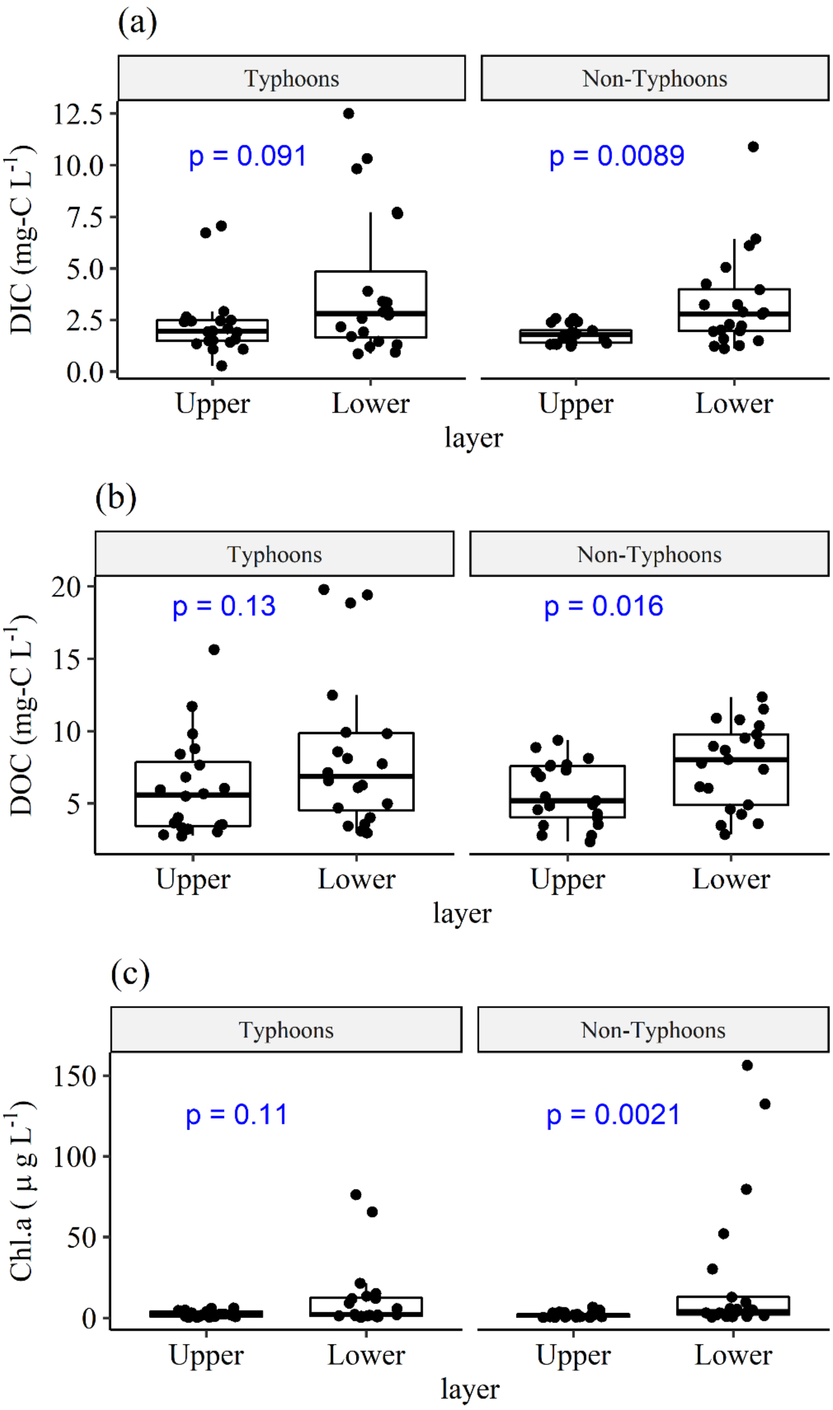

Fig. 2. Comparisons of **(a)** DIC, **(b)** DOC, and **(c)** Chl *a* between upper (DIC_U, DOC_U, Chl_U) and lower (DIC_L, DOC_L, Chl_L) layers, grouped by typhoon and non-typhoon years. *Bullet points* show the water sampling data. We used a t-test to obtain the *p*-values (*blue texts*). Total sampling 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.

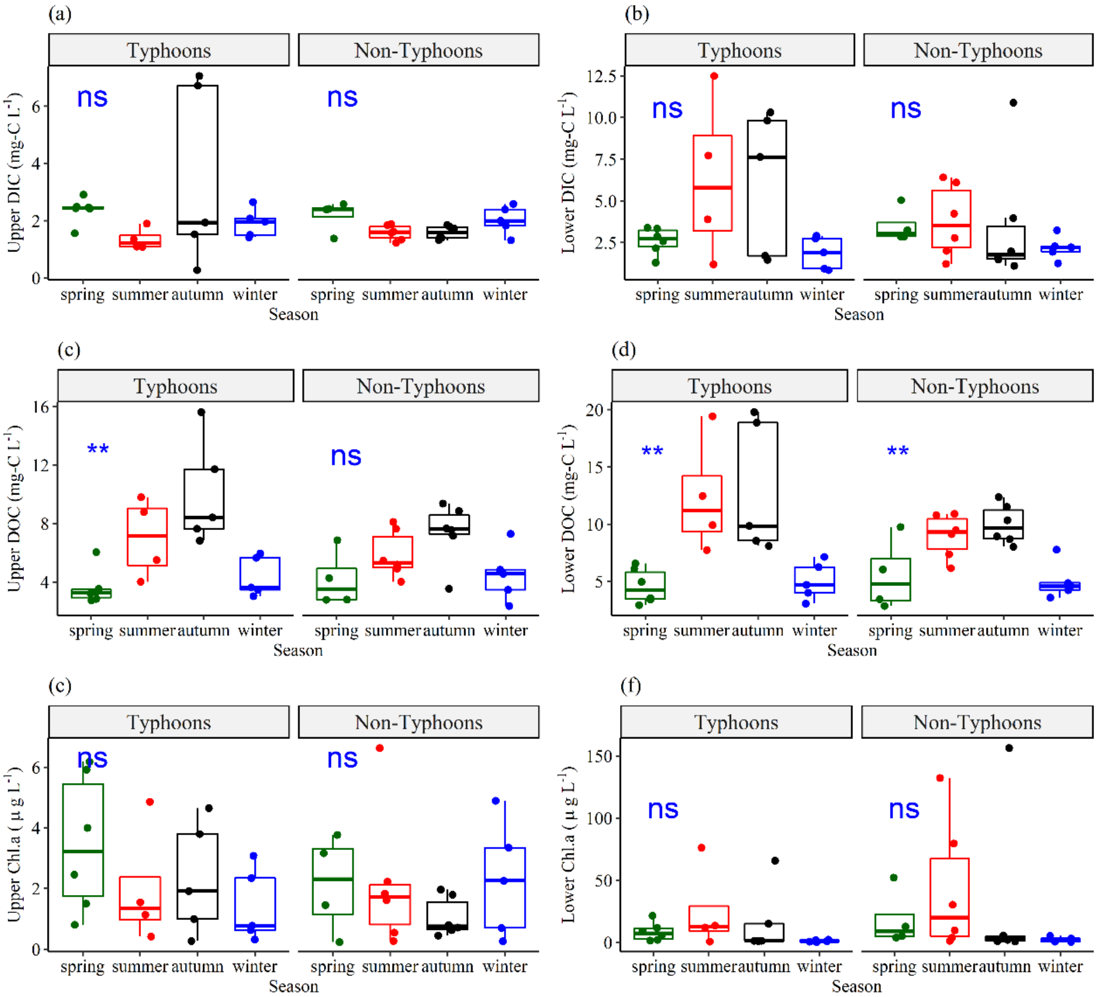

Fig. 3. Seasonal variations of **(a)** upper layer DIC (DIC_U), **(b)** lower layer DIC (DIC_L), **(c)** upper layer DOC (DOC_U), **(d)** lower layer DOC (DOC_L), **(e)** upper layer Chl. a (Chl_U), **(f)** lower layer Chl. *a* (Chl_L) grouped by typhoon and non-typhoon years. The bullet points show 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.

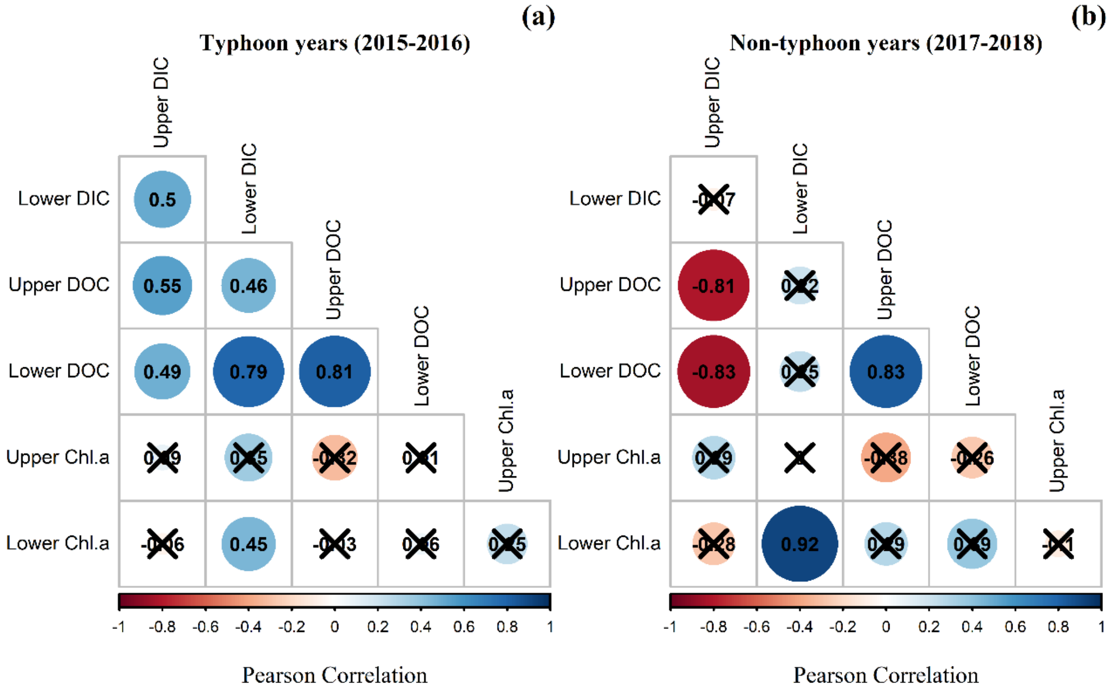

Fig. 4. Pearson correlation coefficients of DIC, DOC, and Chl. *a* concentration at upper layer and lower layer DIC ($DIC_U,\ DIC_L$), DOC ($DOC_U,\ DOC_L$), Chl. a ($Chl_U,\ Chl_L$) during **(a)** typhoon years and **(b)** non-typhoon years. *Black-crosses* show insignificant values (p-values are > 0.05).

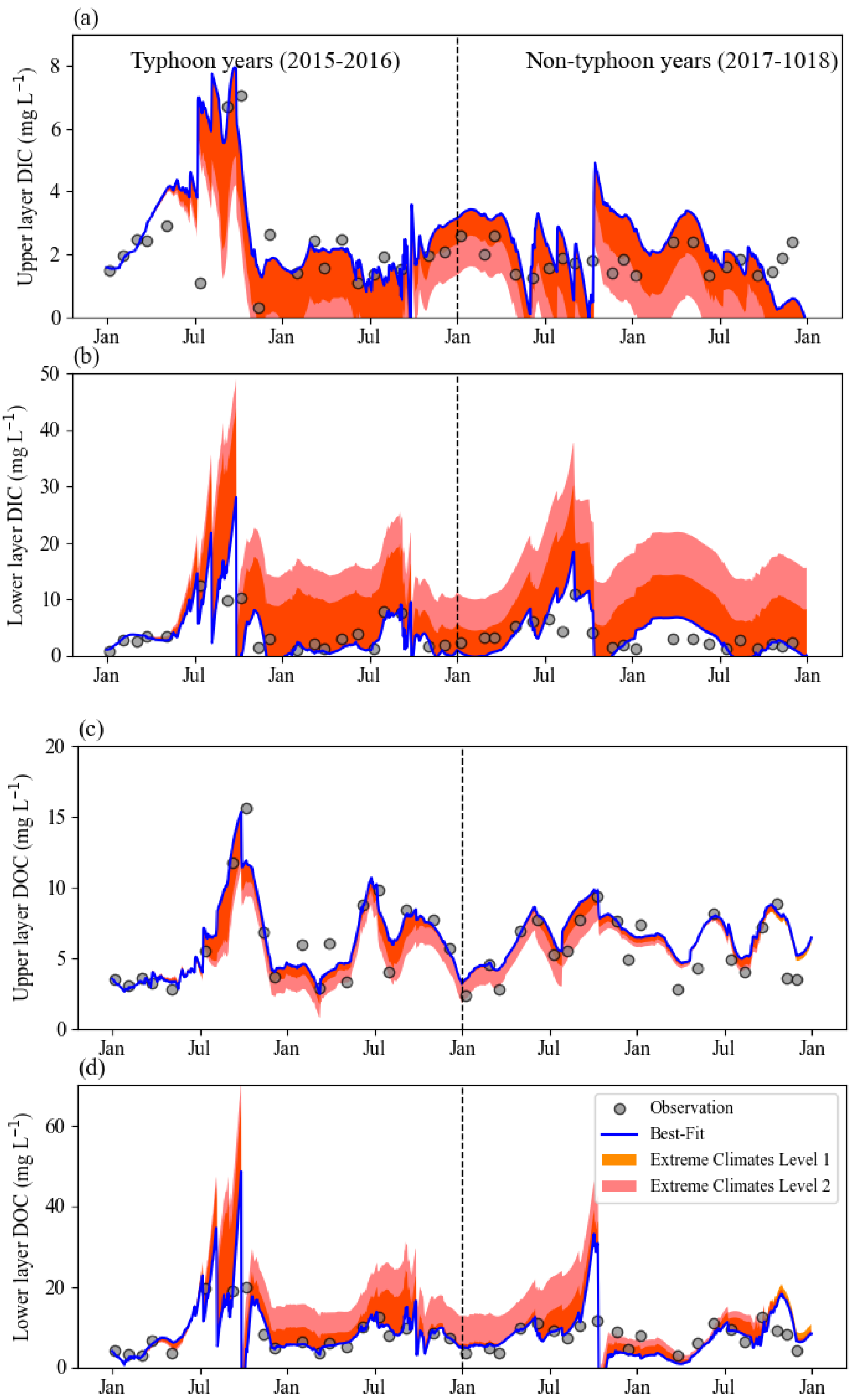
(a)
Typhoon years (2015-2016)
Non-typhoon years (2017-1018)
Upper layer DIC (mg L^{-1})
(b)
Lower layer DIC (mg L^{-1})
(c)
Upper layer DOC (mg L^{-1})
(d)
Lower layer DOC (mg L^{-1})
Jan Jul Jan Jul Jan Jul Jan Jul Jan
Observation
Best-Fit
Extreme Climates Level 1
Extreme Climates Level 2

Fig. 5. 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, 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*.

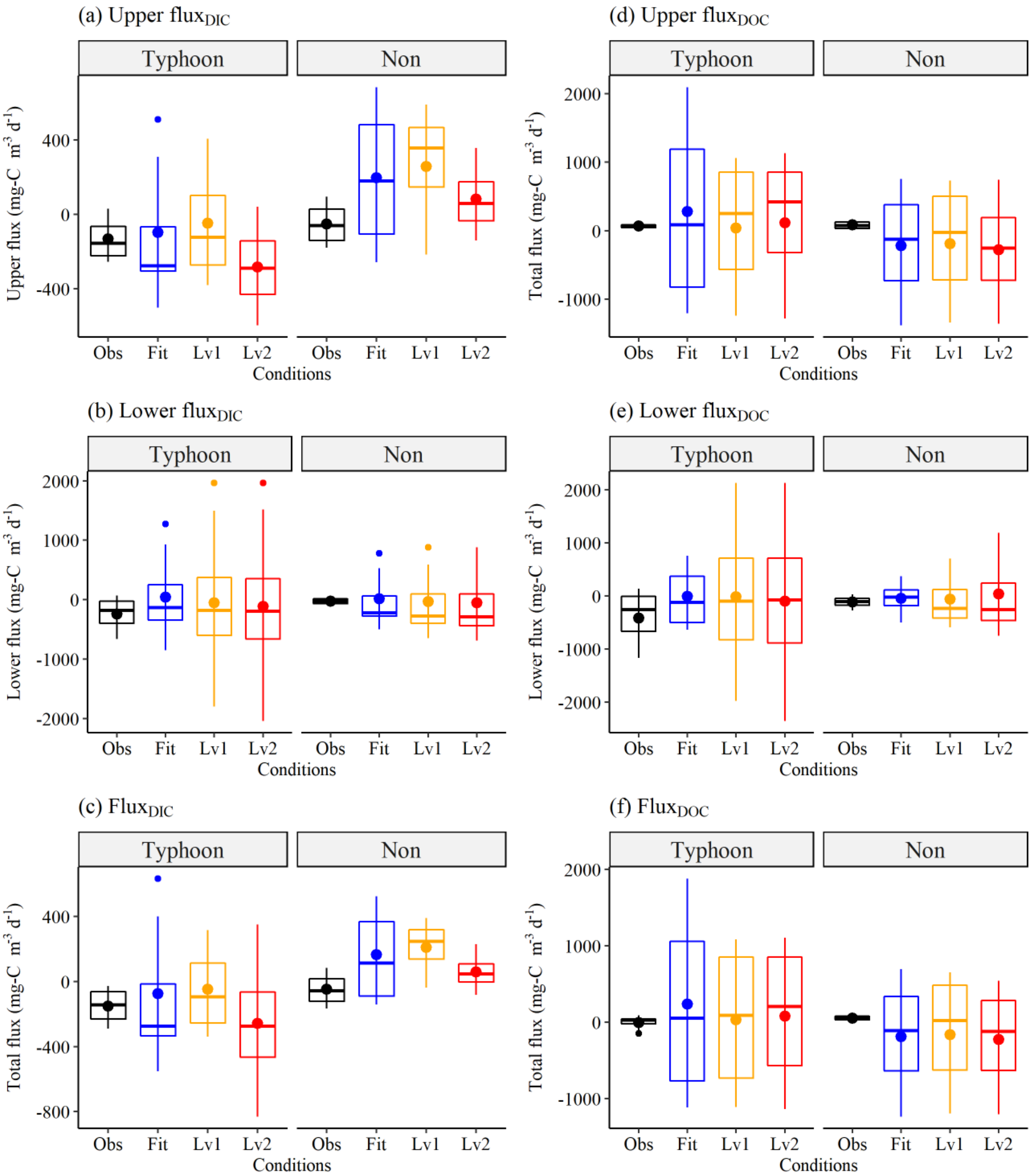

Fig. 6. Interannual (a) Upper flux$_{DIC}$, (b) Lower flux$_{DIC}$, (c) Flux$_{DIC}$, (d) Upper flux$_{DOC}$, (e) Lower flux$_{DOC}$, and (f) Flux$_{DOC}$ (mg C m^{-3} d^{-1}) grouped by typhoon and non-typhoon years. The *Obs* condition (*black boxes*) show the observation data as in Fig. 5. The *Fit* 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.

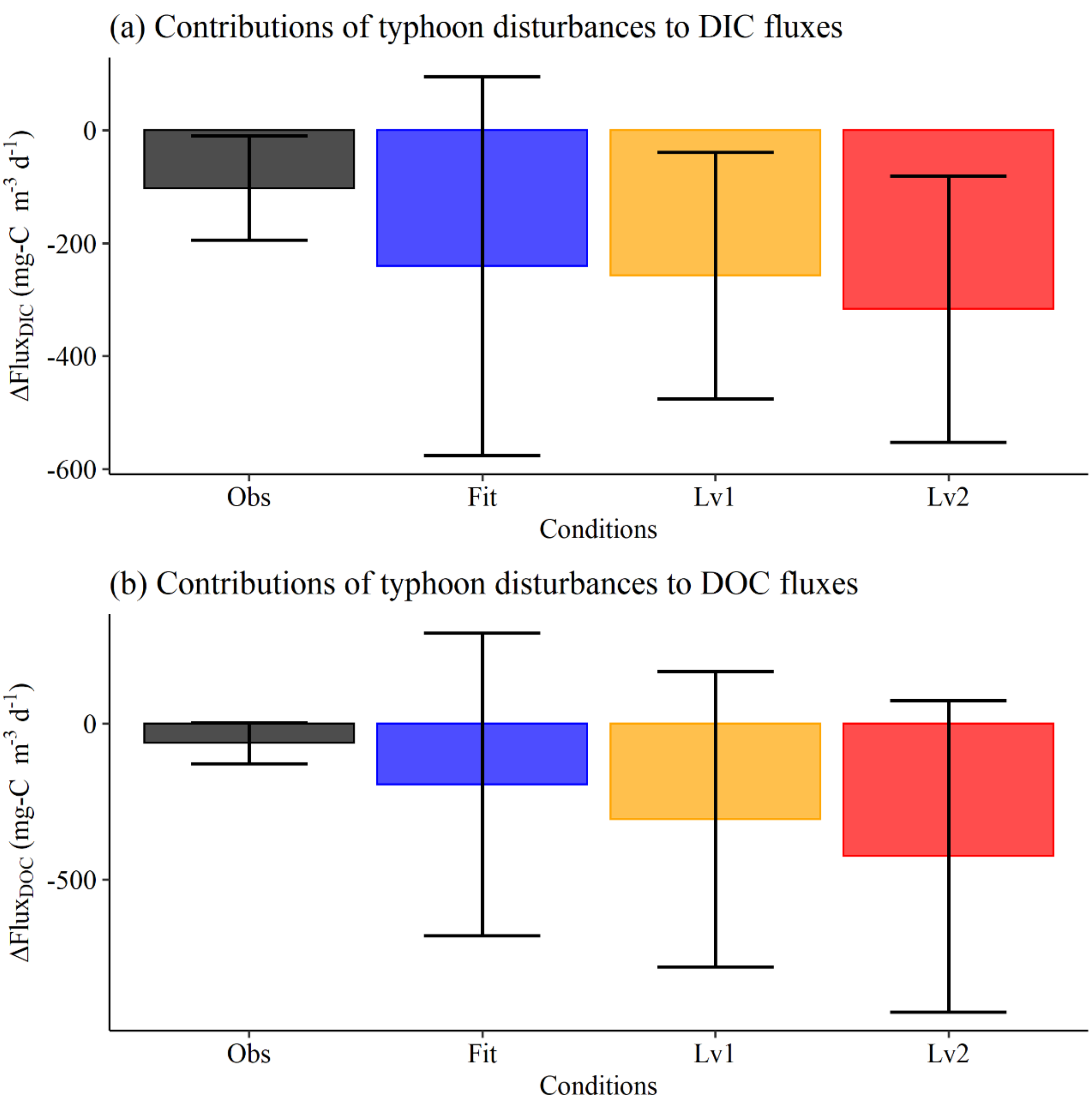

Fig. 7. Contributions of typhoon disturbances to (a) DIC and (b) DOC fluxes (mg-C m^{-3} d^{-1}). Contributions of typhoons to C fluxes ($\Delta Flux_{DIC}$, $\Delta Flux_{DOC}$) represented the intervals of $Flux_{DIC}$ (or $Flux_{DOC}$) in typhoon yeas and non-typhoon years (e.g., $\Delta Flux_{DIC} = Flux_{DIC}$ in typhoon yeas $- Flux_{DIC}$ in non-typhoon years).

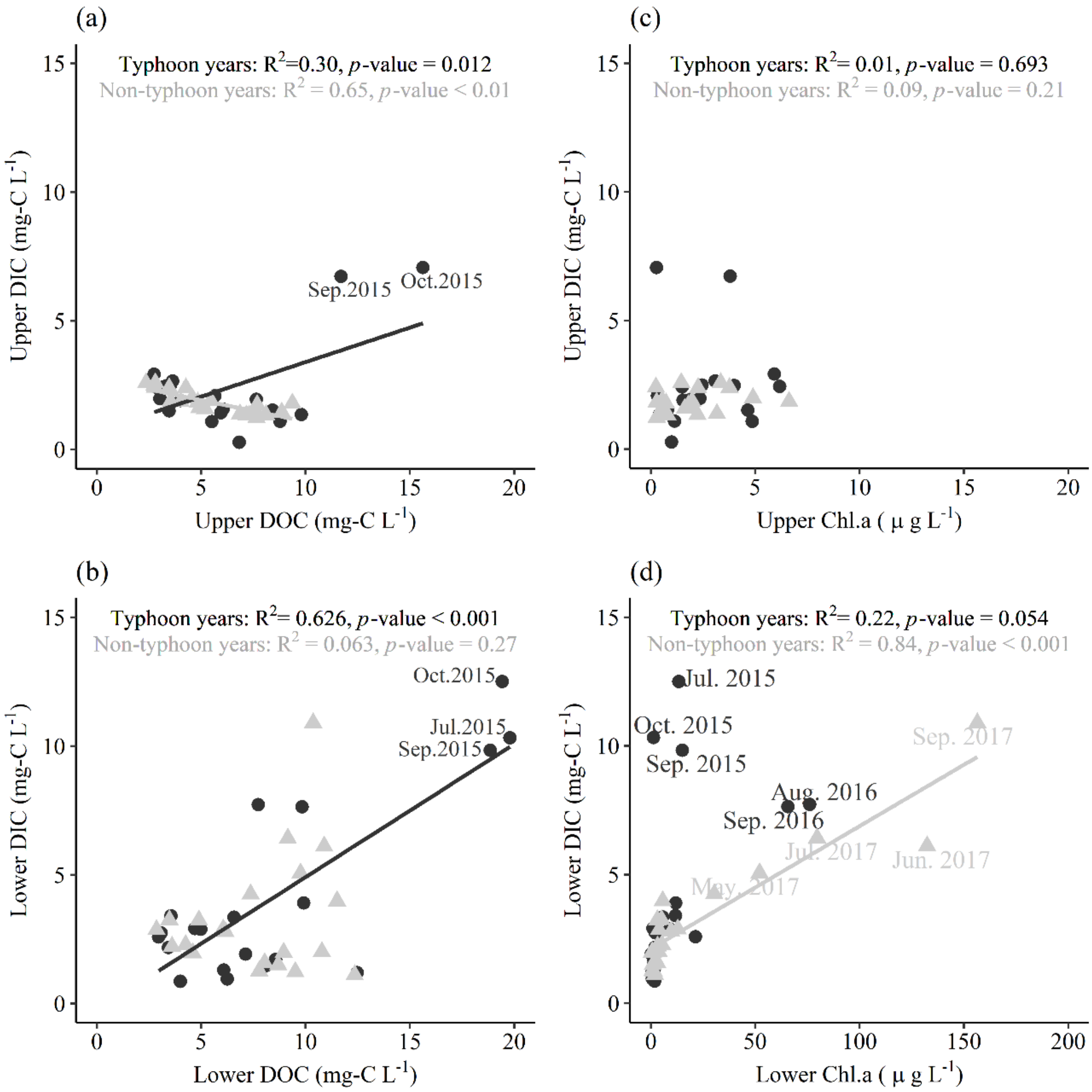

Fig. 8. Interactions of measurements data in typhoon years and non-typhoon years. The *black circles*: typhoon years, *gray triangles*: non-typhoon years. The solid lines represent the linear regression line in typhoon years (*black lines*) and non-typhoon years (*gray lines*).