



## **CO<sub>2</sub> Emissions From Rivers Draining the Tibetan Plateau Reveal Substantial Increases Over the Past 20 Years**

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## RESEARCH ARTICLE

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### Key Points:

- During 2000–2019, riverine CO<sub>2</sub> emissions on the Tibetan Plateau was  $37 \pm 3$  Tg C yr<sup>-1</sup>, twice that of previous estimates
- Tibetan Plateau riverine CO<sub>2</sub> emissions significantly increased by 14% from 2000 to 2019 due largely to increased discharge
- Future climate-driven increases in discharge are expected to further amplify CO<sub>2</sub> emissions from the Tibetan Plateau

### Supporting Information:

Supporting Information may be found in the online version of this article.

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## CO<sub>2</sub> Emissions From Rivers Draining the Tibetan Plateau Reveal Substantial Increases Over the Past 20 Years

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**Abstract** Inland water CO<sub>2</sub> emissions are a critical component of the global carbon cycle, yet their magnitude and long-term response to climate change remain poorly constrained, primarily because of uncertainties in the air–water CO<sub>2</sub> concentration gradient. This knowledge gap is particularly pronounced on the Tibetan Plateau, where thin air and exceptional climate sensitivity complicate accurate quantification. To improve estimates of the air–water CO<sub>2</sub> concentration gradient in this remote region, we developed an optimized, locally adapted model that explicitly accounts for the region's thin air and low atmospheric CO<sub>2</sub> partial pressure. Our results indicate that annual riverine CO<sub>2</sub> emissions from the Tibetan Plateau averaged  $\sim 37 \pm 3$  Tg C yr<sup>-1</sup> during 2000–2019, nearly double the previous estimate of 18.60 Tg C yr<sup>-1</sup>. Spatially, riverine CO<sub>2</sub> emissions exhibited a distinct pattern, with lower emissions across the central plateau and higher emissions in the southeastern region, largely driven by steeper channel gradients and enhanced gas transfer velocities. Over the study period, riverine CO<sub>2</sub> emissions increased significantly by  $\sim 14\%$ , corresponding to an average rate of 0.25 Tg C yr<sup>-2</sup>. This increase was closely linked to the expansion of discharge-dependent river surface area. These findings suggest that projected climate-driven increases in river discharge are likely to further enhance CO<sub>2</sub> emissions from the Tibetan Plateau. The high riverine CO<sub>2</sub> emissions and their temporal increase identified in this study also indicate an increasing role of inland waters in the regional carbon cycle, undermining efforts toward carbon neutrality.

**Plain Language Summary** Rivers and streams are important sources of carbon dioxide (CO<sub>2</sub>) to the atmosphere and play a key role in the global carbon cycle. However, CO<sub>2</sub> emissions from rivers on the Tibetan Plateau remain poorly constrained because the region is remote, difficult to sample, and characterized by unique high-altitude conditions that are highly sensitive to climate change. Here, we applied models specifically adapted to the environmental conditions of the Tibetan Plateau to estimate riverine CO<sub>2</sub> emissions. We found that rivers and streams released  $\sim 37$  Tg C yr<sup>-1</sup> during 2000–2019, approximately twice the previous estimates. Emissions were substantially higher in the southeastern Plateau than in the central region, primarily because steeper river gradients enhanced CO<sub>2</sub> transfer from water to the atmosphere. Riverine CO<sub>2</sub> emissions also increased by 14% over the study period, largely driven by rising river discharge associated with climate change. These results provide a more robust assessment of CO<sub>2</sub> emissions from Tibetan rivers and demonstrate the importance of incorporating regional environmental characteristics into emission models. Continued climate-driven increases in riverine CO<sub>2</sub> emissions could strengthen their contribution to the regional carbon cycle and complicate future carbon neutrality efforts.

## 1. Introduction

Inland waters play a pivotal role in the global carbon (C) cycle by receiving vast amounts of C from terrestrial ecosystems (Aufdenkampe et al., 2011; Battin et al., 2009; Cole et al., 2007) and recycling it back to the

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atmosphere via carbon dioxide (CO<sub>2</sub>) outgassing. These aquatic CO<sub>2</sub> emissions show significant spatiotemporal variability, which is difficult to capture from (ground-based or remotely sensed) observations (Lauerwald et al., 2015, 2023; Liu et al., 2022; Ran et al., 2021). Riverine CO<sub>2</sub> emissions are constrained by gas transfer velocities, air–water gas concentration gradients, and river surface areas available for gas exchange, which in the past have been estimated by distinct methods trained on different input data applied at different spatiotemporal resolutions (e.g., Liu et al., 2022; Ran et al., 2021). As a result, global riverine CO<sub>2</sub> emission estimates still vary widely, ranging from 0.23 to 2.26 Pg C yr<sup>−1</sup> (Battin et al., 2023; Lauerwald et al., 2023; Liu et al., 2022; Tian et al., 2023). Moreover, much less is known regarding long-term (decadal to centennial) trends in emissions in response to ongoing global change, whether induced by climate change or other direct anthropogenic disturbances such as land-use change (Battin et al., 2008; Regnier et al., 2013, 2022; Tian et al., 2023). Model derived estimates suggest that global aquatic CO<sub>2</sub> emissions may have increased by about 30% since pre-industrial times (Regnier et al., 2022; Tian et al., 2023), but direct observational evidence in support of these findings is sparse, spatially biased and of too short duration.

Global estimations of riverine CO<sub>2</sub> emissions based on empirical data are skewed toward Europe and North America, while data availability for Asian streams and rivers is limited (Lauerwald et al., 2015; Liu et al., 2022). This is particularly true for the Tibetan Plateau, one of the most climate-sensitive regions globally (Qu et al., 2018; Ran et al., 2021). The Plateau is characterized by high elevation and abundant freshwater resources (Qu et al., 2018; Ran et al., 2021) and is therefore often referred to as both the “Water Tower of Asia” and the “Third Pole.” However, due to its remoteness, high altitude, and scarcity of in situ observations, it has been difficult to understand and quantify the dynamics of aquatic CO<sub>2</sub> cycling on the Tibetan Plateau (Q. Yang et al., 2024). This is a major knowledge gap, given that this region is currently undergoing large-scale changes caused by a complex interplay between rapid warming, intensive industrialization and expanding agricultural activities (Ciais et al., 2014; Doblas-Reyes et al., 2021; Qu et al., 2018; J. Yang & Huang, 2021), all of which are anticipated to amplify in the future. Hence, accurately capturing the spatiotemporal variations in aquatic CO<sub>2</sub> emissions and their underlying drivers is crucial for predicting future changes in the greenhouse gas budget of the Tibetan Plateau.

Although previous studies have attempted to quantify riverine CO<sub>2</sub> emissions from the region (Qu et al., 2018; Ran et al., 2021), they did not account for the high altitude causing locally low atmospheric CO<sub>2</sub> partial pressure above the Tibetan Plateau. This omission, likely induced a significant underestimation of the air–water gas concentration gradients and thus aquatic CO<sub>2</sub> emissions. Furthermore, recent trends in emissions and the identification of potential drivers of their spatial-temporal variability have not yet been addressed. To address these knowledge gaps, the aims of this study are to (1) estimate riverine CO<sub>2</sub> emissions from the Tibetan Plateau using a locally tailored model obtained via parameter optimization; and using the optimized model to (2) investigate the environmental factors that drive the spatial patterns and trends of riverine CO<sub>2</sub> emissions.

## 2. Materials and Methods

### 2.1. Study Site

The Tibetan Plateau (26°–39.5°N; 73° to 104.5°E) is a high-altitude plateau covering 4.4 million km<sup>2</sup>, with elevation from 81 to 8,535 m.a.s.l. (average: 3,719 m.a.s.l.) (Figure 1). This region is characterized by a cold and semi-arid climate, with a mean annual temperature of 2.8°C and mean annual precipitation of 912 mm yr<sup>−1</sup> (meteorological period: 1990–2020) (Muñoz Sabater, 2019). Due to the high elevation, the atmosphere above the Tibetan Plateau is thinner than in lower-elevation regions, resulting in more intense solar radiation, extreme temperature variations, and lower atmospheric concentrations of oxygen and CO<sub>2</sub> (Anslan et al., 2020). The Tibetan Plateau has an abundance of freshwater resources, with large rivers such as the Yellow, Yangtze, Mekong, and Indus rivers originating there (Anslan et al., 2020; Ghimire et al., 2020).

### 2.2. Data Overview

In general, estimating CO<sub>2</sub> emissions from the river network requires two key inputs: the effective river surface area and the areal CO<sub>2</sub> emissions (see detailed equations below). Estimating these two quantities requires hydrological, meteorological, and watershed data, including stream length, discharge, air temperature, precipitation, air pressure, and watershed properties, including population density, slope derived from elevation data, and gross and net primary production (GPP and NPP) (Liu et al., 2022). In this study, discharge was derived from the Global

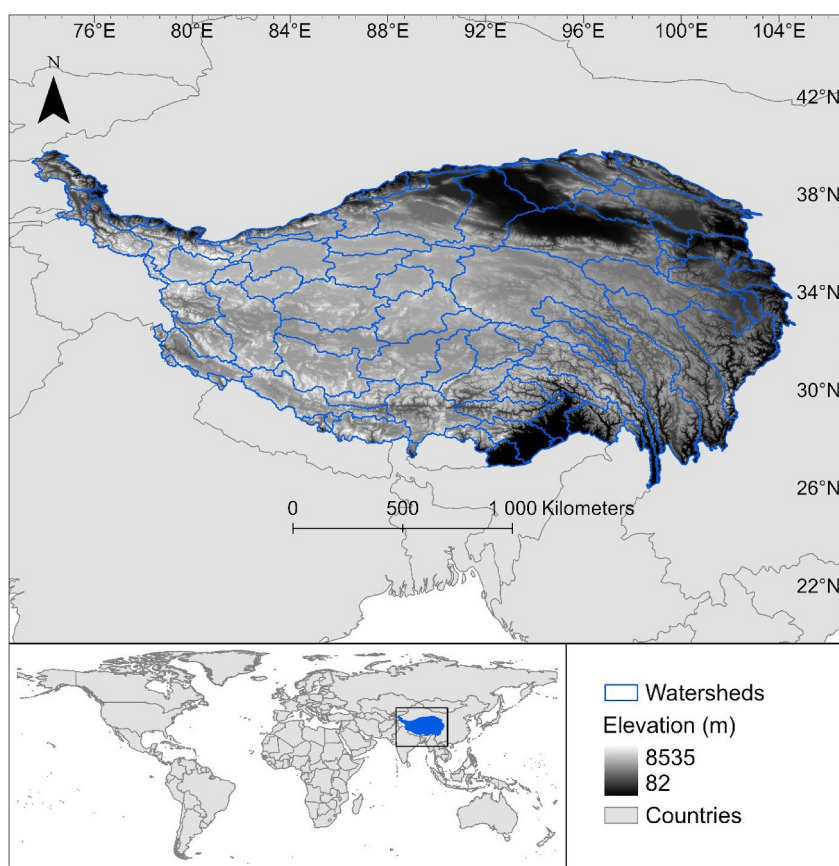


**Figure 1.** Overview map showing elevation (m) and watersheds on the Tibetan Plateau. Details of this elevation map are provided in Table S1 in Supporting Information S1. Also, the location of the Tibetan plateau is shown on the world map.

Reach-level Flood Reanalysis (GRFR) data set with high temporal (3-hourly discharge aggregated to daily values from 1980 to 2019) resolution (Y. Yang et al., 2019). GRFR uses the high-resolution MERIT-Basins vector network to route waterflow globally. Stream length and stream order for each river segment are available in the vector-based hydrography database MERIT-Basins (Lin et al., 2019; Yamazaki et al., 2019). The MERIT-Basins database relies on HydroBASINS Levels 1 and 2 to define its highest-level boundaries. Here, watershed emissions were aggregated to the HydroBASINS Level 5, as this density allowed for a good spatial comparison of the basins. The ERA5-Land data layers provided air temperature, precipitation and air pressure from 1950 to present at a spatial resolution of  $0.1^\circ$  (Muñoz Sabater, 2019). In total, 15 watershed predictors were included for the estimate of riverine areal  $\text{CO}_2$  emission, namely slope, elevation, soil pH, temperature, soil respiration, watershed area, GPP, base saturation, population density, soil gravel content, soil organic carbon content, sand content, precipitation, NPP, and the proportion of wetlands in the watershed (Lauerwald et al., 2015; Liu et al., 2022). More information about data sources and spatiotemporal extent can be found in Table S1 in Supporting Information S1.

The in situ  $p\text{CO}_2$  measurements were taken from 2001 to 2021, with one-time measurements and averages over multiple months and/or years (Liu, 2022; Ran, 2021). For a detailed description of the data used in this study, see Table S1 in Supporting Information S1.

### 2.3. $\text{CO}_2$ Emissions Estimations

The riverine  $\text{CO}_2$  emissions were calculated based on Equations 1 and 2 (Lauerwald et al., 2015; Liu et al., 2022; Ran et al., 2021).

$$\text{CO}_2 \text{ evasion flux} [\text{Tg C yr}^{-1}] = A_{\text{river,eff}} [\text{m}^2] * F_{\text{CO}_2} [\text{mmol C m}^{-2} \text{ day}^{-1}] * 12.01 [\text{g C mol}^{-1}] * 365 [\text{day yr}^{-1}] * 10^{-15} \quad (1)$$

$$F_{\text{CO}_2} [\text{mmol C m}^{-2} \text{ day}^{-1}] = k_{\text{actual}} [\text{m day}^{-1}] * \Delta \text{CO}_2 [\text{mol C L}^{-1}] * 10^6 \quad (2)$$

where  $F_{\text{CO}_2}$  represents the rate of  $\text{CO}_2$  efflux per unit area (i.e., areal riverine  $\text{CO}_2$  emissions),  $A_{\text{river,eff}}$  refers to the effective river area ( $\text{m}^2$ ),  $k_{\text{actual}}$  is the gas transfer velocity ( $\text{m day}^{-1}$ ), and  $\Delta \text{CO}_2$  is the water-air gas concentration gradient ( $\text{mol C L}^{-1}$ ) (Lauerwald et al., 2015; Liu et al., 2022; Ran et al., 2021). Details of the calculation of each parameter are shown in Equations 1 and 2 are explained below.

There are different approaches to model each of these  $\text{CO}_2$  emission parameters, with differences in underlying hydrologic assumptions, geographic coverage, and mathematical algorithms (Lauerwald et al., 2015; Liu et al., 2022; Raymond et al., 2012, 2013), introducing uncertainties to the  $\text{CO}_2$  emission estimations. We determined the most accurate calculation for each of the parameters based on literature values (e.g., Ran et al., 2021), and integrated these parameters into an optimized model. The resulting model was used to assess spatiotemporal  $\text{CO}_2$  emission trends across the Tibetan Plateau. Note that all  $\text{CO}_2$  emission parameters presented below are represented at a monthly temporal resolution.

### 2.3.1. Effective River Surface Area

The river surface area ( $A$ ;  $\text{m}^2$ ) was calculated using Equation 3 at the annual ( $_{\text{ann}}$ ) and monthly ( $_{\text{mon}}$ ) scales by obtaining time-varying river widths ( $W$ ; annually: Equation 4 (Raymond et al., 2012), monthly: Equation 5 (Liu et al., 2022)) and multiplying them with a constant river segment length obtained from GRFR data (Y. Yang et al., 2019). The monthly and annual river widths were calculated using natural logarithmic and linear functions of monthly discharge ( $Q$ ;  $\text{m}^3 \text{ s}^{-1}$ ), respectively.

$$A [\text{m}^2] = W [\text{m}] * \text{RiverLength} [\text{m}] \quad (3)$$

$$\ln(W_{\text{ann}} [\text{m}]) = 2.56 + 0.423 * \ln(Q_{\text{ann}} [\text{m}^3 \text{ s}^{-1}]) \quad (4)$$

$$W_{\text{mon}} [\text{m}] = a [\text{s m}^{-2}] * Q_{\text{mon}}^b [\text{m}^3 \text{ s}^{-1}] \quad (5)$$

The exponent  $b$  was calculated from the coefficient of variation based on daily discharge ( $Q_{C.V.}$ ; unitless), annual mean discharge ( $Q_{\text{ann}}$ ), and basin runoff (Equation 6). Then the coefficient  $a$  was resolved using  $b$ ,  $Q_{\text{ann}}$  and mean annual river width ( $W_{\text{mean\_ann}}$ ) as estimated according to Equation 7.

$$b[1] = 0.074 * \ln(Q_{C.V.}[1]) - 0.026 * \ln(\text{runoff} [\text{mm yr}^{-1}]) - 0.011 * \ln(Q_{\text{ann}} [\text{m}^3 \text{ s}^{-1}]) + 0.36 \quad (6)$$

$$W_{\text{mean\_ann}} [\text{m}] = a [\text{s m}^{-2}] * Q_{\text{ann}}^b [\text{m}^3 \text{ s}^{-1}] \quad (7)$$

To obtain the effective river surface area ( $A_{\text{river,eff}}$ ), we excluded both dry and ice-covered streams by setting their area to zero when the mean monthly discharge was  $0 \text{ m}^3 \text{ s}^{-1}$  or when the monthly air temperature was below  $-4.8^\circ\text{C}$  (Lauerwald et al., 2015). In addition,  $A_{\text{river,eff}}$  (with  $A_{\text{mon}}$ ) was calculated using Equation 8 based on a zero discharge days ratio ( $ZDD$ ) from Equation 9 (Liu et al., 2022).

$$A_{\text{mon}} [\text{m}^2] = \begin{cases} A_{\text{mon}}, & (T_{\text{air}} \geq -4.8^\circ\text{C}; Q_{\text{ann}} \geq 60 \text{ m}^3 \text{ s}^{-1}) \\ A_{\text{mon}} * ZDD, & (T_{\text{air}} \geq -4.8^\circ\text{C}; Q_{\text{ann}} < 60 \text{ m}^3 \text{ s}^{-1}) \\ 0, & (T < -4.8^\circ\text{C}) \end{cases} \quad (8)$$

$$ZDD[1] = \left(1 - e^{(11+4*\ln Q_{\text{ann}} [\text{m}^3 \text{ s}^{-1}])}\right)^{-1} \quad (9)$$

### 2.3.2. Gas Transfer Velocity

Gas transfer velocity governs the rate at which riverine CO<sub>2</sub> exchange with the atmosphere can occur (Equation 9), with higher values indicating faster exchange (Ulseth et al., 2019). The actual gas transfer velocity ( $k_{\text{actual}}$ , m day<sup>-1</sup>; Equation 10) was estimated as a function of standardized gas transfer velocity ( $k_{600}$ , m day<sup>-1</sup>) and Schmidt's number ( $Sc$ ; Ulseth et al., 2019).

$$k_{\text{actual}} [\text{m day}^{-1}] = k_{600} [\text{m day}^{-1}] * \left( \frac{Sc[1]}{600} \right)^{-0.5} \quad (10)$$

$Sc$  was calculated based on monthly averaged water temperature (Equation 11). The water temperature was calculated based on the air temperature following Telmer and Veizer (1999). Accounting for low-order streams with steep channel slopes on the Tibetan Plateau, we computed  $k_{600}$  with two different approaches (Equation 12), that is, one that considers energy dissipation ( $eD$ ) for steep channel slopes at low-order streams (Equation 13) and one that does not (Equation 14).

$$Sc[1] = 1742 - 91.24 * T_{\text{water}} + 2.208 * T_{\text{water}}^2 - 0.0219 * T_{\text{water}}^3 \quad (11)$$

$$k_{600} [\text{m day}^{-1}] = \begin{cases} k_{600\_noED}, & (eD \leq 0.02; T_{\text{air}} \geq -4.8^{\circ}\text{C}) \\ k_{600\_ED}, & (eD > 0.02; T_{\text{air}} \geq -4.8^{\circ}\text{C}) \\ 0, & (T_{\text{air}} < -4.8^{\circ}\text{C}) \end{cases} \quad (12)$$

$$k_{600\_ED} [\text{m day}^{-1}] = e^{(1.18 * eD [\text{m}^2 \text{s}^{-3}] + 6.43)} \quad (13)$$

$$k_{600\_noED} [\text{m day}^{-1}] = 2841 * V [\text{m s}^{-1}] * S[1] + 2.02 \quad (14)$$

$eD$  is a function of annual mean flow velocity ( $V_{\text{ann}}$ ; m s<sup>-1</sup>) (Raymond et al., 2012) and channel slope ( $S$ , m m<sup>-1</sup>) (Equation 15).  $V_{\text{ann}}$  was calculated based on annual discharge using Equation 16, an equation proposed for low-flow conditions (Raymond et al., 2012).

$$eD [\text{m}^2 \text{s}^{-3}] = 9.81 [\text{m s}^{-2}] * V [\text{m s}^{-1}] * S[1] \quad (15)$$

$$\ln(V [\text{m s}^{-1}]) = -1.64 + 0.285 * \ln(Q [\text{m}^3 \text{s}^{-1}]) \quad (16)$$

### 2.3.3. Water-Air Gas Concentration Gradient

In streams and rivers, the water-air gas concentration gradient ( $\Delta\text{CO}_2$ ) governs the strength and direction of the resulting CO<sub>2</sub> flux, that is, whether CO<sub>2</sub> is released or taken up. It is based on the difference between the riverine ( $p\text{CO}_{2\_water}$ ) and atmospheric ( $p\text{CO}_{2\_air}$ ) partial pressures of CO<sub>2</sub>, converted to gas concentrations using the adjusted Henry's Law constant  $K\text{CO}_2$  (Equation 17) (Liss & Slater, 1974). In the present study,  $p\text{CO}_{2\_water}$  was modeled using both a generic empirical function constructed based on global observations ( $p\text{CO}_{2\_water, \text{empirical}}$ ; Equation 18) (Lauerwald et al., 2015) and a random forest (RF) model adapted to local conditions ( $p\text{CO}_{2\_water, \text{RF}}$ ; see details below) (Liu et al., 2022), respectively.

$$\Delta\text{CO}_2 [\text{mol L}^{-1}] = K\text{CO}_2 [\text{mol L}^{-1} \text{atm}^{-1}] * (p\text{CO}_{2\_water} [\text{atm}] - p\text{CO}_{2\_air} [\text{atm}]) \quad (17)$$

$p\text{CO}_{2\_water, \text{empirical}}$  was calculated using a multiple linear regression model (Equation 18). For Equation 18, all parameters were calculated at the watershed level (i.e., population density Pop, km<sup>-2</sup>; slope  $S$ , degree; NPP, kg C m<sup>-2</sup> y<sup>-1</sup>) except the air temperature ( $T_{\text{air}}$ , °C), which was calculated at the stream segment level (Table S1 in Supporting Information S1).

$$\log_{10}(p\text{CO}_{2\_water,empirical}[\text{atm}]) = -3.192 + 9.372 * 10^{-3} * \sqrt{\text{Pop}[\text{km}^{-2}]} - 0.279 * \log_{10}(S[^\circ]) + 0.01343 * T_{air}[^\circ\text{C}] + 0.279 * \text{NPP}[\text{kg C m}^{-2}\text{y}^{-1}] \quad (18)$$

Note that while Horgby et al. (2019) presented a linear regression model for riverine  $p\text{CO}_2$  adapted to mountain streams (discharge  $<2.26 \text{ m}^3 \text{ s}^{-1}$ ), the river network on the Tibetan Plateau had more diverse flow conditions (mean monthly discharge:  $51.55 \pm 5.81 \text{ m}^3 \text{ s}^{-1}$  based on GRFR data from Y. Yang et al. (2019)), making their model less applicable to the entire study region. Therefore, we used a generic linear regression model based on global observations instead (Lauerwald et al., 2015). This further allowed us to explore the potential uncertainties in adapting global models without adjustment to local conditions.

For riverine  $p\text{CO}_2$ , we also constructed our own RF model based on observations from the Tibetan Plateau but using the same drivers as Liu et al. (2022), meaning that there is no equation for  $p\text{CO}_{2\_water,RF}$ . Following their method, all drivers were kept constant at watershed (HYBAS) level, except for temperature and precipitation, which varied over time from 2000 to 2019 (Table S1 in Supporting Information S1) (Liu et al., 2022). The previously mentioned 15 watershed predictors (Table S1 in Supporting Information S1). Although underlying data for population density and NPP varied over time, these indicators were calculated as yearly means and kept constant, as mentioned above. Soil properties of the two soil layers were aggregated into the entire soil column (i.e., 30% topsoil and 70% subsoil) for each watershed.

As  $p\text{CO}_2$  observations ranged from daily measurements to multi-month or multi-year records (Liu et al., 2022; Ran et al., 2021), daily observations were aggregated to monthly values to match the predominant temporal resolution of the data set. For longer-term observations, catchment-scale temperature and precipitation were averaged over the corresponding observation period. In total, direct measurements of riverine  $p\text{CO}_{2\_water}$  ( $n = 167$ ) included rivers ( $n = 95$ ), lakes, and reservoirs with discharge between  $0.004$  and  $1,031 \text{ m}^3 \text{ s}^{-1}$  (mean monthly discharge:  $21 \pm 98 \text{ m}^3 \text{ s}^{-1}$ ; Liu, 2022; Ran, 2021). More details on the modeling data and the riverine  $p\text{CO}_2$  measurements can be found in Table S1, Figures S1, and S2 in Supporting Information S1.

Due to the low number of sampling locations (91 locations), the resulting data set had a 90:10 split for training and validation data, respectively. These training points were put into an RF model from the r-package randomForest (Liaw & Wiener, 2022). The hyperparameters of the RF model were set in accordance with Liu et al. (2022), with the number of tries splitting a node set to 5 and the number of trees set to 500. After 100 iterations, the model with the highest explained variance was selected for the calculation of  $p\text{CO}_{2\_water,RF}$ .

$p\text{CO}_{2\_air}$  was determined by either using constant estimates of atmospheric  $\text{CO}_2$  (Jia et al., 2022; Lauerwald et al., 2015; Liu et al., 2022; Ran et al., 2021), or atmospheric  $\text{CO}_2$  varying with altitude and/or time (Horgby et al., 2019). Specifically, using a constant global mean atmospheric  $p\text{CO}_2$  is referred to “ $p\text{CO}_{2\_air,global}$ ”; using a constant regional mean atmospheric  $p\text{CO}_2$  is termed “ $p\text{CO}_{2\_air,TP}$ ”; atmospheric  $p\text{CO}_2$  expressed as a function of elevation is denoted “ $p\text{CO}_{2\_air,elevation}$ ”; atmospheric  $p\text{CO}_2$  calculated as a function of air pressure is called “ $p\text{CO}_{2\_air,pressure}$ .” Air pressure varied with both elevation and time.

In terms of constant  $p\text{CO}_{2\_air}$ , we compared the global  $p\text{CO}_{2\_air,global}$  estimate of  $390 \mu\text{atm}$  (Lauerwald et al., 2015; Liu et al., 2022; Ran et al., 2021) and average Tibetan Plateau-specific  $p\text{CO}_{2\_air,TP}$  of  $238.78 \mu\text{atm}$  (Jia et al., 2022). Note that the global value ( $p\text{CO}_{2\_air,global}$ ) was used to compare our calculations with previous results for a specific Tibetan Plateau region. This allowed us to fill the missing flux data for this region that were not provided in previous global studies and subsequently to validate our calculations.

In addition, we computed varying  $p\text{CO}_{2\_air}$  based on Horgby et al. (2019) (Equation 19), using the mean atmospheric  $\text{CO}_2$  concentration measured in Switzerland in 2017 ( $400.40 \mu\text{atm}$ ), and the ratio between atmospheric pressure ( $P_{atm}$ ; atm) and the standard pressure of 1 atm ( $P_{std}$ ; atm).  $P_{atm}$  was taken both directly from ERA5-Land (Muñoz Sabater, 2019), that is, spatiotemporally varying  $P_{atm}(p\text{CO}_{2\_air,pressure})$ , and calculated from elevation (Equation 20) following Horgby et al. (2019), that is, spatially but temporal-invariant  $P_{atm}(p\text{CO}_{2\_air,elevation})$ .

$$p\text{CO}_{2\_air}[\text{atm}] = 400.40[\mu\text{atm}] * 10^{-6} * \frac{P_{atm}[\text{atm}]}{P_{std}[\text{atm}]} \quad (19)$$

$$P_{\text{atm}}[\text{Pa}] = P_0 * \frac{T_b \frac{g \cdot m}{R \cdot \lambda}}{T_b + \lambda * E} \quad (20)$$

where  $P_0$  indicates the international standard atmosphere value of sea level pressure (1 atm = 101,325 Pa),  $T_b$  assumed sea level temperature (292.15 K),  $\lambda$  temperature lapse rate ( $-0.0065 \text{ K m}^{-1}$ ),  $g$  gravity acceleration ( $9.80665 \text{ m s}^{-2}$ ),  $m$  molecular weight of dry air ( $0.02897 \text{ kg mol}^{-1}$ ),  $R$  gas constant ( $8.3143 \text{ J mol}^{-1} \text{ K}^{-1}$ ) and elevation  $E$  (in m) extracted from digital elevation models (Table S1 in Supporting Information S1).

#### 2.4. Uncertainties and Trend Attribution of CO<sub>2</sub> Emissions

The riverine CO<sub>2</sub> emissions were obtained by combining two different methods for each parameter (i.e., effective river surface area, gas transfer velocity, and  $p\text{CO}_{2,\text{water}}$ ), as well as four  $p\text{CO}_{2,\text{air}}$  (local or global constant, spatially varying or spatiotemporally varying) values for each river segment, yielding 32 model configurations for comparison. Spatiotemporal patterns and trends were explored at the regional (i.e., the entire Tibetan Plateau), watershed (HydroBASINS watersheds at level 5) (Lehner & Grill, 2013) and stream order (stream orders 1–6) scales. For comparison of average CO<sub>2</sub> emissions and  $F\text{CO}_2$  parameters between 2000 and 2019, as well as spatial patterns, the study by Ran et al. (2021) was taken as a reference baseline, which reported observed aquatic CO<sub>2</sub> emissions, gas transfer velocities, and effective surface areas for the Tibetan Plateau. Estimated riverine  $p\text{CO}_{2,\text{water}}$  was compared to direct observations collected between 2000 and 2021 (Figure S1 in Supporting Information S1) (Liu, 2022; Ran, 2021). Although observational data exist for river surface area, gas transfer velocity, and  $p\text{CO}_{2,\text{air}}$ , they are too sparse to provide statistically robust estimates of uncertainty. Therefore, the uncertainty in CO<sub>2</sub> emissions was quantified using the estimated river surface area, gas transfer velocity, and  $p\text{CO}_{2,\text{air}}$ , together with the uncertainty in modeled  $p\text{CO}_{2,\text{water}}$ , represented by the standard deviation derived from comparisons with observations. We identified the primary parameter contributing the largest uncertainty to CO<sub>2</sub> emission estimates. To do so, we compared annual emissions and their long-term trends derived from conventional global models versus locally adapted models.

To identify the dominant drivers of the trends in CO<sub>2</sub> emission budgets between 2000 and 2019, we replaced each of the following annually varying factors, that is, effective river surface area, gas transfer velocity, concentration gradient, or discharge, with its average value for the full study period, and then calculated the corresponding trend in the CO<sub>2</sub> emission budgets (i.e., Theil-Sen slope). Note that when the discharge is kept constant, both the surface area and flow velocity are constant. The contribution was calculated as the relative difference in percentage between the “actual” trend (i.e., with all varying factors) and the trend when each individual parameter (discharge, riverine  $p\text{CO}_2$ , gas transfer velocity, river surface area) was set as constant. Significance of trends in CO<sub>2</sub> emission was assessed using the modified Mann-Kendall test (Hamed & Rao, 1998). In contrast, the spatial drivers were interpreted by the qualitative spatial relationship between  $F\text{CO}_2$  parameters and the resulting emissions.

### 3. Results and Discussion

#### 3.1. Optimal Local Configuration for Riverine CO<sub>2</sub> Emission Modeling

We first compared the 32 alternative parameterizations of the CO<sub>2</sub> emission model against available observations. Among the tested configurations, the most accurate set-up combined the RF-based riverine  $p\text{CO}_2$  ( $p\text{CO}_{2,\text{water,RF}}$ ), monthly river surface area ( $A_{\text{mon}}$ ), gas transfer velocity ( $k_{\text{actual,ED}}$ ), and atmospheric  $p\text{CO}_2$  as a function of observed air pressure ( $p\text{CO}_{2,\text{air,pressure}}$ ) (Table 1; and Table S1 and Figures S3–S5 in Supporting Information S1).

Results for  $p\text{CO}_{2,\text{water}}$  differed substantially between the locally adapted RF model and the global model. The locally adapted RF model predicted a decreasing east-west trend in aquatic  $p\text{CO}_2$ , which was consistent with direct observations (Figure 2; Figure S3 in Supporting Information S1). In contrast, the global model (Lauerwald et al., 2015) revealed a large deviation from observation (Figure S4 in Supporting Information S1) and little spatial variation (Figure 2). The RF model predicted that most streams and rivers were oversaturated with respect to atmospheric CO<sub>2</sub> (average  $p\text{CO}_{2,\text{water,RF}}$ :  $666 \pm 5.50 \mu\text{atm}$ ), which was in agreement with previous studies (average  $p\text{CO}_{2,\text{water}}$ :  $615 \mu\text{atm}$ ) (Qu et al., 2017; Ran et al., 2021). This was in contrast with the results obtained with the global model, which failed to capture the magnitude of observed  $p\text{CO}_{2,\text{water}}$  (average  $p\text{CO}_{2,\text{water,empirical}}$ :  $363 \pm 0.00 \mu\text{atm}$ ; no temporal variation) (Table 1).

**Table 1**

Parameter Values for the  $F_{CO_2}$  Estimation (Mean  $\pm$  One Standard Deviation|Theil-Sen Slope) Using Different Parameter Modeling Approaches

| Parameter                  | Unit                           | Mean $\pm$ St. Dev.   | Theil-Sen slope         | Literature values |
|----------------------------|--------------------------------|-----------------------|-------------------------|-------------------|
| $A_{ann}$                  | km <sup>2</sup>                | 5,516.54 $\pm$ 361.57 | 29.50**                 | 10,515            |
| $A_{mon}$                  | km <sup>2</sup>                | 4,190.28 $\pm$ 279.38 | 24.30*                  | 10,515            |
| $k_{actual\_noED}$         | m day <sup>-1</sup>            | 12.33 $\pm$ 0.41      | 0.05**                  | 12–36             |
| $k_{actual\_ED}$           | m day <sup>-1</sup>            | 21.85 $\pm$ 0.72      | 0.07**                  | 12–36             |
| $pCO_{2\_water,empirical}$ | $\mu$ atm                      | 362.93 $\pm$ 0.00     | (no temporal variation) | 615               |
| $pCO_{2\_water,RF}$        | $\mu$ atm                      | 666.32 $\pm$ 5.52     | −0.62**                 | 615               |
| $pCO_{2\_air,pressure}$    | $\mu$ atm                      | 237.06 $\pm$ 25.26    | 0.02*                   | 238.78            |
| $pCO_{2\_air,elevation}$   | $\mu$ atm                      | 244.38 $\pm$ 0.00     | (no temporal variation) | 238.78            |
| Discharge                  | m <sup>3</sup> s <sup>-1</sup> | 51.55 $\pm$ 5.81      | 0.56*                   | —                 |
| Temperature                | °C                             | −2.91 $\pm$ 0.44      | 0.03                    | —                 |
| Precipitation              | mm month <sup>-1</sup>         | 2.10 $\pm$ 1.52       | 0.00000035              | —                 |

Note. Theil-Sen slope indicates the trend in  $F_{CO_2}$ . \*Refers to  $p < 0.05$  for the Theil-Sen slope, and \*\*refers to  $p < 0.01$  for the Theil-Sen slope. Literature values for the Tibetan Plateau are from Ran et al. (2021). Values without standard deviation and Theil-Sen slope are spatially variable, but constant in time.

The optimized atmospheric CO<sub>2</sub> ( $pCO_{2\_air,pressure}$ ) accounted for both local topography (i.e., high elevation) and monthly atmospheric CO<sub>2</sub> variability (Equation 19). Instead, using uncorrected global atmospheric CO<sub>2</sub> concentrations ( $pCO_{2\_air,global}$ ) resulted in biased air-water CO<sub>2</sub> gradients because the reduction in atmospheric pressure with increasing elevation was not properly captured. Applying a constant Tibetan Plateau mean atmospheric  $pCO_2$  ( $pCO_{2\_air,TP}$ ) partly corrected for the high-altitude environment and improved the estimate but did not account for spatial variation in topography. In contrast, the spatially explicit elevation-based estimate ( $pCO_{2\_air,elevation}$ ) captured this topographic variation and produced spatial patterns very similar to those obtained using pressure ( $pCO_{2\_air,pressure}$ ) (Figures S5–S6 in Supporting Information S1). This similarity indicates that elevation was the primary driver of atmospheric  $pCO_2$ , outweighing the influence of temporal variability.

Annual and monthly modeled river surface areas increased at similar rates for the period 2000–2019 (29.50 km<sup>2</sup> yr<sup>-1</sup> and 24.30 km<sup>2</sup> yr<sup>-1</sup>) and we therefore selected the monthly resolved estimates to account for seasonal variation (Table 1). However, both approaches led to effective river surface areas ( $A_{ann}$ : 5,516.54  $\pm$  361.57 km<sup>2</sup>;  $A_{mon}$ : 4,190.28  $\pm$  279.38 km<sup>2</sup>) that were substantially smaller than those reported by Ran et al. (2021) (i.e., a manually delineated river surface area 10,515 km<sup>2</sup>, Table 1). Ran et al. (2021) considered a larger land area (2.60 million km<sup>2</sup>) than ours (2.5 million km<sup>2</sup>), extending it to the east with more observations available in streams from the Yangtze and Yellow River networks. They also did not account for the effects of ice cover and ephemeral rivers in their analysis. As a result, our estimates gave a reduced river surface area (especially for low-order streams).

Across the Tibetan Plateau, estimates of average standardized gas transfer velocity ( $k_{600}$ ) differed depending on the treatment of energy dissipation. When energy dissipation was not accounted for, we obtained a standardized gas transfer velocity of 18.15  $\pm$  0.55 m day<sup>-1</sup> ( $k_{600\_noED}$ ), which increased to 28.81  $\pm$  0.89 m day<sup>-1</sup> when energy dissipation was accounted for ( $k_{600\_ED}$ ). Our estimated  $k_{600}$  range aligns with the one reported by Ran et al. (2021; 12.0–36.0 m day<sup>-1</sup>) for high-energy streams on the Tibetan Plateau. We found that the local maximum actual gas transfer velocity ( $k_{600\_ED}$ ) on the Tibetan Plateau could reach up to 100 m day<sup>-1</sup>, consistent with the estimate for mountain streams in the Swiss Alps (86.40 m day<sup>-1</sup>) (Horgby et al., 2019). Given the presence of high-energy mountain streams on the Tibetan Plateau,  $k_{actual\_ED}$  was preferred over  $k_{actual\_noED}$  (reaching up to 50 m day<sup>-1</sup>).

### 3.2. Strongest Riverine CO<sub>2</sub> Emissions in the Southeastern Tibetan Plateau and Low-Order Streams

Using the optimized model configuration, we found distinct spatial patterns of CO<sub>2</sub> emission budgets (Tg C yr<sup>-1</sup>) and  $F_{CO_2}$  rates (mmol CO<sub>2</sub> m<sup>-2</sup> day<sup>-1</sup>) across the Tibetan Plateau. These spatial patterns reflected those found for the effective river surface area, gas transfer velocity, atmospheric  $pCO_2$  and riverine  $pCO_2$  (Figure 3). Except for atmospheric  $pCO_2$ , river surface area, gas transfer velocity, and the resulting CO<sub>2</sub> emissions consistently exhibited higher values in the southeast part of the domain compared to those found in the northwest (Figure 3). In

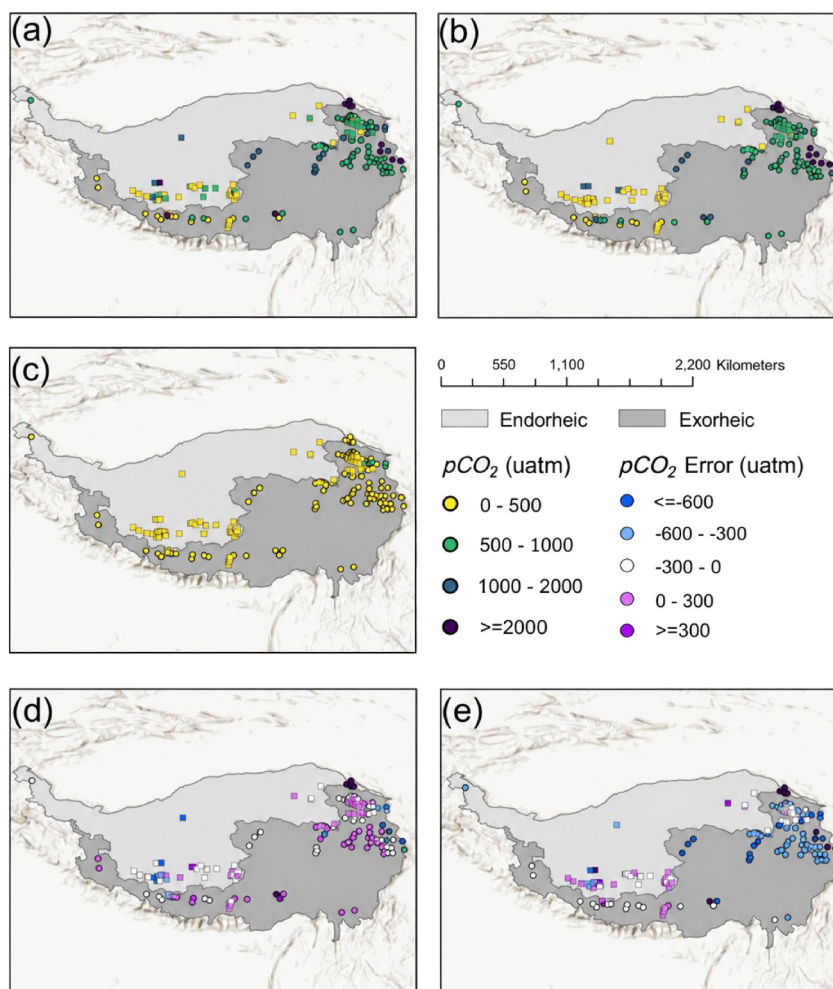


**Figure 2.** Maps of the directly measured and modeled riverine  $p\text{CO}_2$  (uatm) (a) observed  $p\text{CO}_2$ , (b)  $p\text{CO}_2$  modeled using the locally adapted random forest (RF) model, (c)  $p\text{CO}_2$  modeled using the global model, (d)  $p\text{CO}_2$  errors estimated as the difference between the RF model and observations, (e) same for the global model. Squares: Observations from lakes; Circles: Observations from streams and rivers. See Methods for further details.

contrast, atmospheric  $p\text{CO}_2$  was higher in the northeastern portion of the Plateau due to its low elevation. The resulting spatial gradient in emission rates was strong with  $\text{FCO}_2$  lower than  $200 \text{ mmol C m}^{-2} \text{ day}^{-1}$  in north-western basins and exceeding  $2,000 \text{ mmol C m}^{-2} \text{ day}^{-1}$  in south-east basins. The highest values were found in the Jinsha Jiang River on the headwaters of the Yangtze River due to its high riverine  $p\text{CO}_2$  and gas transfer velocity, and low atmospheric  $p\text{CO}_2$  (Figure 3). Interestingly, such spatial patterns emerged irrespective of the method applied to constrain each individual  $\text{CO}_2$  emission modeling parameter (Figures S6 and S8–S10 in Supporting Information S1). However, despite capturing the spatial gradient, the application of a global-mean atmospheric  $p\text{CO}_2$  predicted widespread  $\text{CO}_2$  uptake even in the exoreic regions, which is unrealistic as most freshwater rivers are  $\text{CO}_2$  sources (Figure S8 in Supporting Information S1; Dean & Battin, 2024).

The east-west gradient of riverine  $\text{CO}_2$  emissions and  $\text{FCO}_2$  found in our study was consistent with Ran et al. (2021) and partly reflects the geography of the Tibetan Plateau. The north-western parts have high elevations but gentle slopes (Jarvis et al., 2008), with gas transfer velocities mostly below  $10 \text{ m day}^{-1}$  and  $\text{FCO}_2$  less than  $1,000 \text{ mmol C m}^{-2} \text{ day}^{-1}$  (Figure 3). In contrast, the steeper southeastern parts exhibit much faster gas exchange rates of  $50\text{--}100 \text{ m day}^{-1}$ . An overall plateau-wide  $\text{FCO}_2$  ranges from under 200 to over  $2,000 \text{ mmol C m}^{-2} \text{ day}^{-1}$ , consistent with that reported for Chinese streams and rivers (Zhou et al., 2024).

$\text{FCO}_2$  in rivers varied with stream order, with first order streams emitting more  $\text{CO}_2$  than higher order streams (Figure S11 in Supporting Information S1). This pattern was also captured by other model configurations (Figures

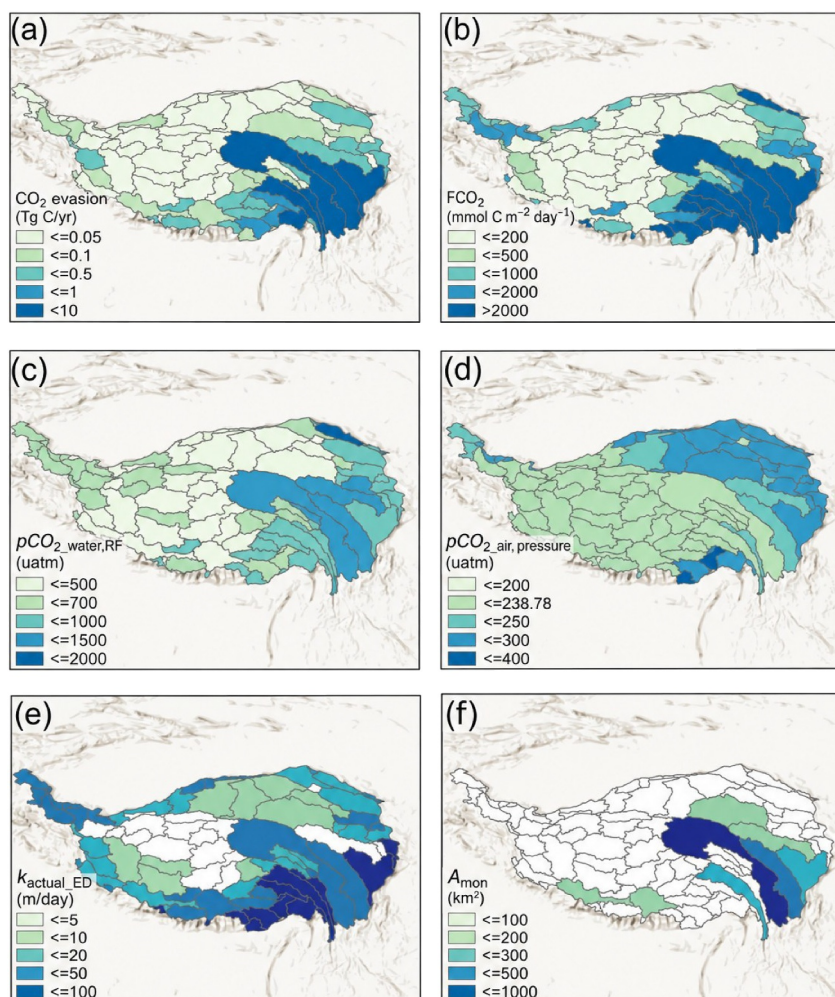


**Figure 3.** Spatial patterns of CO<sub>2</sub> emission rates and their underlying drivers across the Tibetan Plateau averaged over the study period 2000–2019. (a) Annual CO<sub>2</sub> emission budgets, (b) areal emission rate,  $FCO_2$ , (c)  $pCO_{2\_water,RF}$ , (d)  $pCO_{2\_air,pressure}$ , (e)  $k_{actual\_ED}$ , and (f)  $A_{mon}$ .

S12 and S13 in Supporting Information S1) except those using global atmospheric CO<sub>2</sub> and the global model configuration for constraining riverine  $pCO_2$  (Figure S14 in Supporting Information S1). Similar patterns were also observed in other regions such as the Yangtze River and Great Pearl River in China (Ran et al., 2021), where first order streams had higher riverine CO<sub>2</sub> emissions and CO<sub>2</sub> emissions decreased as streams and rivers became larger (i.e., higher stream order). This decreasing trend is consistent with the rapid degassing of CO<sub>2</sub> in small streams driven by high gas transfer velocities due to steep slopes (Hotchkiss et al., 2015; Wallin et al., 2013). Across the Tibetan Plateau, around 50% of the bulk riverine CO<sub>2</sub> emissions originated from first and second order streams. The high CO<sub>2</sub> emissions in the first-order streams mainly result from high gas transfer velocities and large cumulative surface areas (Figures S15–S18 in Supporting Information S1). First order streams typically receive C not only from the surrounding terrestrial ecosystems, especially saturated soils adjacent to the stream channel, but also from inorganic C-rich groundwater (Aufdenkampe et al., 2011; Deirmendjian & Abril, 2018). Soil respiration, including root respiration, boosts groundwater  $pCO_2$  supersaturation, thereby sustaining relatively high  $pCO_2$  in headwaters and/or streams with low connectivity (Liu et al., 2016). This CO<sub>2</sub> supersaturation, combined with high gas transfer velocity, generally favor high CO<sub>2</sub> emission rates in low order streams. However, significant uncertainties remain between studies. For instance, dissolved CO<sub>2</sub> concentrations have been reported to increase (Liu et al., 2016) or decrease (Song et al., 2023) with increasing stream order along the Yangtze River. Large differences in CO<sub>2</sub> concentrations between tributaries within the same stream order have been observed,

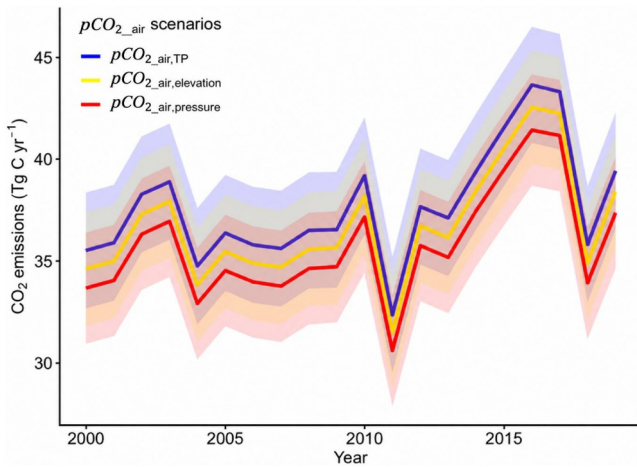


**Figure 4.** Comparison of annual  $FCO_2$  emissions for different sets of model parameters ( $A_{eff}$ ,  $k_{actual}$  and  $\Delta CO_2$ ) combined with constant atmospheric  $pCO_2$  of 238.78  $\mu atm$  ( $pCO_{2\_air,TP}$ ), spatially varying  $pCO_2$  ( $pCO_{2\_air,elevation}$ ), and spatiotemporally varying  $pCO_2$  ( $pCO_{2\_air,pressure}$ ).

suggesting a more complex relationship between stream order and riverine  $C$  dynamics on and around the Tibetan Plateau (Liu et al., 2016).

### 3.3. High and Increasing Riverine $CO_2$ Emissions on the Tibetan Plateau

period 2000–2019, our optimized model configuration (i.e.,  $pCO_{2\_air,pressure}$ ,  $k_{actual\_ED}$ ,  $pCO_{2\_water,RF}$ , and  $A_{mon}$ ) led to annual  $CO_2$  emission budgets for the entire Tibetan Plateau region of  $36.73 \pm 2.79$   $Tg\ C\ yr^{-1}$ , with a statistically significant increasing rate of  $0.25\ Tg\ C\ yr^{-2}$  ( $p < 0.05$ ) (Figure 4 and Table 2). Application of different local atmospheric  $pCO_2$  values (i.e.,  $pCO_{2\_air,TP}$ ,  $pCO_{2\_air,pressure}$ ,  $pCO_{2\_air,elevation}$ ) and locally optimized river surface area and gas transfer velocity revealed slight differences in the mean annual  $CO_2$  emissions over 2000–2019 ( $35.75 \pm 2.73$  to  $37.68 \pm 2.84$   $Tg\ C\ yr^{-1}$ ) (Figure 4 and Table 2). These slight differences stemmed from the widespread local atmospheric  $pCO_2$  values around the region-mean value (i.e., constant  $pCO_{2\_air,TP}$  of 238.78  $\mu atm$ ) (Figure 3) though  $pCO_{2\_air,TP}$  did not account for spatial (and temporal) variability like  $pCO_{2\_air,elevation}$  (and  $pCO_{2\_air,pressure}$ ) did. In contrast, application of the global-mean atmospheric  $pCO_2$ ,  $A_{mon}$  and  $k_{actual\_ED}$  reduced the  $CO_2$  emissions of the Tibetan Plateau by about 10  $Tg\ C\ yr^{-1}$ , with annual  $CO_2$  emission budgets of  $27.77 \pm 2.19$   $Tg\ C\ yr^{-1}$  (Table S3 in Supporting Information S1). Given the same local atmospheric  $pCO_2$  ( $pCO_{2\_air,pressure}$ ), a global model of riverine  $pCO_2$  ( $pCO_{2\_water,empirical}$ ) would yield 76% lower regional  $CO_2$  emissions of  $8.79 \pm 0.62$   $Tg\ C\ yr^{-1}$ . Combinations of global-mean atmospheric  $pCO_2$  with other unrealistic parameters (i.e.,  $k_{actual\_noED}$ ,  $pCO_{2\_water,empirical}$ ) would even lead to drastically different results in terms of magnitude and direction of the flux, with a  $CO_2$  uptake comprised between  $-0.13 \pm 0.073$  and  $-0.075 \pm 0.083$   $Tg\ C\ yr^{-1}$  (Table S3 in Supporting Information S1). This uptake was even amplified by 10%–15% when  $k_{actual\_ED}$  instead of  $k_{actual\_noED}$  was applied in the global model configuration ( $-0.26 \pm 0.081$   $Tg\ C\ yr^{-1}$ ) (Table S3 in Supporting Information S1). The highest riverine  $CO_2$  emission budgets were determined (i.e.,  $43.20 \pm 3.06$   $Tg\ C\ yr^{-1}$ ) when using a larger effective surface area ( $A_{ann}$  instead of  $A_{mon}$ ), regional-mean  $pCO_{2\_air,TP}$  and other locally optimized parameters. Despite significant differences, all 32 model assessments showed a significant increase ( $p < 0.05$ ) in riverine  $CO_2$  emissions over the last two decades of observations, with trends ranging from  $0.06\ Tg\ C\ yr^{-2}$  to  $0.30\ Tg\ C\ yr^{-2}$  while the optimized model showed a trend of  $0.25\ Tg\ C\ yr^{-2}$  (Table 2; Table S3 in Supporting Information S1).

The  $CO_2$  emission based on the optimized model ( $36.73 \pm 2.79$   $Tg\ C\ yr^{-1}$ ) was about twice as high as the  $CO_2$  emissions of  $18.60 \pm 7.90$   $Tg\ C\ yr^{-1}$  previously reported by Ran et al. (2021). This result was obtained despite the smaller river surface area ( $4,190$ – $5,516$   $km^2$ ) estimated here compared to the one ( $10,515$   $km^2$ ) applied by Ran et al. (2021). The main reason for this discrepancy stemmed from the much lower atmospheric  $pCO_2$  used in our study (e.g., 238.78  $\mu atm$  for  $pCO_{2\_air,TP}$ ) compared to the global atmospheric  $pCO_2$  value ( $\mu atm$ ) used by Ran et al. (2021). This finding revealed the large biases introduced by the use of global atmospheric  $pCO_2$  values in high altitude regions such as the Tibetan Plateau.

The increasing temporal trend detected by the optimized model configurations represented a 14% increase in  $CO_2$  evasion over the last two decades. This increase was of similar magnitude to that reported by Ran et al. (2021) for

**Table 2**

Overview of Mean  $CO_2$  Emission Values (Mean  $\pm$  Standard Deviation, in  $Tg\ C\ yr^{-1}$ ) and Temporal Trends (Theil-Sen Slope, in  $Tg\ C\ yr^{-2}$ ) Across Different Model Configurations

| Combinations             |                     |                       |                        |                                                      |                 |
|--------------------------|---------------------|-----------------------|------------------------|------------------------------------------------------|-----------------|
| $pCO_{2\_air}$           | $pCO_{2\_water}$    | Gas transfer velocity | Effective surface area | $CO_2$ emissions ( $Tg\ C\ yr^{-1}$ ; mean $\pm$ SD) | Theil-Sen slope |
| $pCO_{2\_air,TP}$        | $pCO_{2\_water,RF}$ | $k_{actual\_ED}$      | $A_{mon}$              | $37.68 \pm 2.84$                                     | 0.26**          |
| $pCO_{2\_air,pressure}$  | $pCO_{2\_water,RF}$ | $k_{actual\_ED}$      | $A_{mon}$              | $36.73 \pm 2.79$                                     | 0.25*           |
| $pCO_{2\_air,elevation}$ | $pCO_{2\_water,RF}$ | $k_{actual\_ED}$      | $A_{mon}$              | $35.75 \pm 2.73$                                     | 0.25*           |

Note. The standard deviations reflect annual variations. The trends are estimated over the period 2000–2019. \*Refers to  $p < 0.05$  for the Theil-Sen slope, and \*\*refers to  $p < 0.01$  for the Theil-Sen slope.

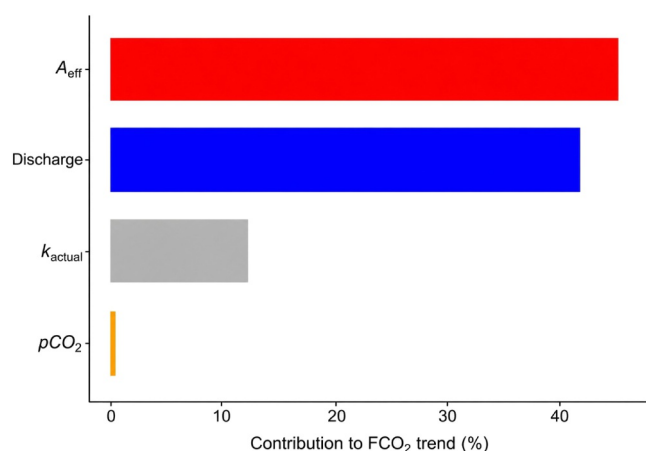


**Figure 5.** Contribution of the main model parameters to the temporal trend in  $CO_2$  emission flux (2000–2019). The contribution was calculated as the relative difference in percentage between the “real” trend and the trend when each individual parameter (discharge,  $pCO_{2\_water}$ , gas transfer velocity, river surface area) was set as constant (see Methods). Trends are expressed as Theil-Sen slopes.

the Tibetan Plateau (18%) for the period 1980–2010. All the modeled riverine  $CO_2$  emission time series showed a similar interannual variability, with the lowest emissions in 2011 and a peak in 2017 (Figure S7 in Supporting Information S1), which was not surprising given that the inter-annual variability is mostly driven by discharge, a parameter applied consistently across models.

### 3.4. Discharge as the Main Driver of Increasing Riverine $CO_2$ Emissions and Its Implications

The significant increase in riverine  $CO_2$  emissions by 13.83% during the period 2000–2019 was primarily driven by increased discharge (42%) (Figure 5 and Figures S19–S22 in Supporting Information S1), which impacted both river surface area and gas transfer velocity. Notably, when the discharge was kept constant, both the surface area and flow velocity were constant. The surface area of rivers and streams was strongly correlated with discharge through width variations (Lauerwald et al., 2015) while gas transfer velocities depended on stream velocities and hence discharge (McDowell & Johnson, 2018). Our attribution analysis further revealed that temporal variations in gas transfer velocity accounted for a smaller contribution (13%) to the  $CO_2$  emission trend, while temporal variation in riverine  $pCO_2$  contributed only marginally to the trend (0.37%). This suggests that, while spatial patterns of fluvial  $CO_2$  emissions were largely determined by the gas transfer

velocity, its impact on temporal changes in  $FCO_2$  over the last two decades was less important. These findings were consistent for all the other model configurations that relied on the local atmospheric  $pCO_2$  (See Figures S20–S22 in Supporting Information S1).

Considering that increased discharge was the most dominant driver of enhanced riverine  $CO_2$  emissions on the Tibetan Plateau, a sustained warming climate in the future will likely further enhance these emissions, provided substrate availability does not become a major constraint (Ciais et al., 2014; Doblas-Reyes et al., 2021). Accelerated melting of glaciers, combined with ongoing and projected increases in precipitation during the wet season, will lead to higher discharge, and wider streams and larger river surface areas (Woolway et al., 2020; L. Zhang et al., 2013). The spatial extension of the river network will lead to enhanced inundation of the riparian zone, which is not only an important source of labile organic C to the river network but also of  $CO_2$  produced by microbial and root respiration in the flooded zone (Liu et al., 2016; Song et al., 2023). Projected climate change-induced precipitation increases (Doblas-Reyes et al., 2021) will stimulate lateral carbon transfers and thus lead to higher organic loadings, potentially higher decomposition rates and higher  $pCO_2$ , which in turn will further enhance riverine  $CO_2$  emissions (Kang et al., 2010; Liu & Raymond, 2018; Qu et al., 2017; H. Zhang et al., 2022; L. Zhang et al., 2013). Therefore, under a warmer climate, increased discharge and greater hydrologic connectivity on the Tibetan Plateau are likely to lead to higher riverine  $CO_2$  emissions in the future.

## 4. Conclusions

Based on a locally adapted optimal modeling configuration, our study provides a new assessment of riverine  $CO_2$  emissions from the Tibetan Plateau. Estimated emissions were almost twice that of the previous estimate (Ran et al., 2021), highlighting the importance of accounting for the low atmospheric  $CO_2$  partial pressure at high elevations. Riverine  $CO_2$  emissions significantly increased by 13.83% from 2000 to 2019, with a clear spatial contrast between low emission watersheds in the western Tibetan Plateau and high-emission watersheds in the east. The temporal increase was primarily driven by concurrent increases in discharge linked to climate warming in the region.

Our results improve the quantitative understanding of the factors controlling the magnitude and spatiotemporal patterns of riverine  $CO_2$  emissions from the Tibetan Plateau and their expected evolution in the future. Our study highlights the need for more long-term observations of riverine  $pCO_2$  and river surface area to better constrain  $CO_2$  emissions estimates and advance our understanding of the carbon dynamics in this environment sensitive to climate change.

## Conflict of Interest

The authors declare no conflicts of interest relevant to this study.

## Availability Statement

All data used in this study are publicly available. Reach-scale hydrology data, including river discharge (NetCDF format) and the corresponding river network (vector format), were obtained from the Global Reach-Level Hydrology data set (Lin et al., 2019). The synthesized  $p\text{CO}_2$  observations for the Tibetan Plateau are available in Liu (2022) and Ran (2021), with the compiled data set archived by Iwan (2025). ERA5-Land climate data including air temperature, precipitation, and surface pressure were obtained from Muñoz-Sabater et al. (2021). Topographic data were derived from the SRTM 90 m Digital Elevation Database (Jarvis et al., 2008). Ecosystem productivity data (MODIS MOD17 products GPP and NPP) were obtained from Running and Zhao (2015). Global gridded soil respiration data were obtained from Hashimoto et al. (2015, 2025) and soil properties were obtained from the Harmonized World Soil Database (Nachtergaele et al., 2010). Wetland vector data were obtained from the Global Lakes and Wetlands Database (Lehner & Döll, 2004). Gridded population density data were obtained from the National Bureau of Statistics of China (Xu, 2017).

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