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A new biogeochemical modelling framework (FLaMe v1.0) for lake methane emissions on the regional scale: Development and application to the European domain

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Abstract

This study presents a new physical-biogeochemical modelling framework for simulating lake methane (CH_4) emissions at regional scales. The new model, FLaMe v1.0 (**F**luxes of **La**ke **Me**thane), rests on an innovative, computationally efficient lake clustering approach that enables the simulation of CH_4 emissions across a large number of lakes. Building on the Canadian Small Lake Model (CSLM) that simulates the lake physics, we develop a suite of biogeochemical modules to simulate transient dynamics of organic Carbon (C), Oxygen (O_2), and CH_4 cycling. We first test the performance of FLaMe by analyzing physical and biogeochemical processes in two representative lakes (an oligotrophic, deep lake driven by cold climate *versus* a trophic, shallow lake driven by warm climate). Next, we evaluate the model by comparing simulated and observed timeseries of CH_4 emissions in four well-surveyed lakes. We then apply FLaMe at the European scale to evaluate simulated diffusive and ebullitive lake CH_4 fluxes against *in-situ* measurements in both boreal and central European regions. Finally, we provide a first assessment of the spatio-temporal variability in CH_4 emissions from European lakes smaller than 1000 km^2 (n=108407, total area = $1.33x10^5$ km^2), indicating a total emission of 0.97±0.23 Tg CH_4 yr^{-1}, with the uncertainty constrained by combining FLaMe and machine learning techniques. Moreover, 30% and 70% of these CH_4 emissions are through diffusive and ebullitive pathways, respectively. Annually averaged CH_4 emission rates per unit lake area during 2010–2016 have a South-to-North decreasing gradient, resulting in a mean over the European domain as 7.39 g CH_4 m^{-2} yr^{-1}. Our simulations reveal a strong seasonality in European lake CH_4 emissions, with late summer emissions nearly ten times higher than winter values. This pronounced seasonal variation highlights the importance of accounting for the sub-annual variability in CH_4 emissions to accurately constrain regional CH_4 budgets. In the future, FLaMe could be embedded into Earth System Models to investigate the feedback between climate warming and global lake CH_4 emissions.

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1. Introduction

Methane (CH_4) is the second most important greenhouse gas after carbon dioxide (CO_2), with a Global Warming Potential (GWP) per mass ~28 times higher than that of CO_2 over a 100-year horizon (Saunois *et al.*, 2020). Over the last centuries, the atmospheric CH_4 concentration has increased from 722 ppb in the pre-industrial period (year 1750) to 1923 ppb in year 2023 (Saunois, et al., 2020; Dlugokencky, 2022; Forster *et al.*, 2024) and this increase is expected to continue in the future. The critical role of CH_4 in global warming has called for the establishment of a comprehensive global CH_4 budget, which embraces both natural and anthropogenic sources (Saunois *et al.*, 2016; 2020; 2024). This budget identified inland waters (wetlands, lakes, reservoirs, ponds, rivers, etc.) as an important, yet highly uncertain atmospheric CH_4 source (Jackson *et al.*, 2020, 2024; Saunois, *et al.*, 2020, Canadell *et al.*, 2021). Global lake CH_4 emissions, which has been estimated to account for ~5 to 20% of total CH_4 emissions (576 $TgCH_4\ yr^{-1}$), are the largest contributors to this inland water source (Jackson *et al.*, 2020; Saunois *et al.*, 2020). However, estimates of its magnitude vary depending on the assessment methods, with discrepancies of up to a factor of four (Saunois *et al.*, 2020; DelSontro and John 2018; Rosentreter *et al.*, 2021; Bastviken *et al.*, 2011; Deemer *et al.*, 2016; Johnson *et al.*, 2021; Holgerson and Raymond 2016; Stavert *et al.* 2022). This variability in global estimates also manifests itself at the continental scale. For instance, estimates of European lake methane emissions range from 0.9 to 2.5 Tg $CH_4\ yr^{-1}$ (Petrescu *et al.* 2021, 2023; Lauerwald *et al.*, 2023).

Observation-based upscaling approaches are highly dependent on the availability and quality of *in-situ* measurements, which are unevenly distributed across the globe and biased towards summer months (Canadell *et al.*, 2021; Johnson *et al.*, 2022). Although the number of CH_4 emission

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63 measurements from lakes has increased considerably in recent decades, the two largest current
64 databases together contain only 1081 records from 575 lakes worldwide (Rosentreter *et al.*, 2021;
65 Johnson *et al.*, 2022). This relatively small data compilation is unlikely to capture the full diversity
66 of physical and biogeochemical patterns of >1.4 million lakes worldwide, which vary by morphology,
67 climate, trophic status, and underlying sediment characteristics (Rinta *et al.*, 2017; Bastviken 2004,
68 2022; Deemer and Holgerson 2021; Johnson *et al.*, 2022). Even more critically, the underlying data
69 collection was not designed to capture the inter-annual and decadal variability in CH_4 emissions
70 driven by climate change and nutrient dynamics, hence rendering the decomposition of the total lake
71 CH_4 fluxes into natural and human-induced components challenging (Saunois *et al.*, 2020). Finally,
72 although current instruments and techniques can effectively capture CH_4 fluxes through diffusive
73 (driven by gradients of aqueous CH_4 concentrations) and ebullitive (via gas bubbles in the sediments
74 due to oversaturation) emission pathways, measurements related to lake turnover events (release of
75 previously accumulated CH_4 due to stratification and ice cover) remain highly challenging (Denfeld
76 *et al.*, 2018; Mayr *et al.*, 2020; Zimmermann *et al.*, 2019). These limitations induce large uncertainties
77 in observation-based upscaling methods. In this context, process-based modelling approaches – that
78 rely on detailed representations of lake physical and biogeochemical processes informed and tested
79 with the available observational data – offer complementary strategies to help reduce these
80 uncertainties.

81 Process-based biogeochemical models provide powerful tools to upscale scarce observations,
82 both in space and in time. In combination with the available observational datasets, they can help
83 deliver regional and global estimates of lake CH_4 emissions from daily to decadal timescales, as well
84 as future projections. These mechanistic models can also help identify the drivers such as climate,

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land-use and atmospheric composition changes responsible for the complex temporal dynamics of lake CH_4 emissions. Over the last decades, several process-based models have thus emerged, e.g., LAKE 2.0 (Stepanenko *et al.* 2016), bLake4Me (Tan *et al.*, 2015), and ALBM (Tan *et al.*, 2018; 2024), to estimate lake CH_4 emissions to the atmosphere. These models explicitly account for the physical and biogeochemical processes that govern lake CH_4 dynamics and resulting emissions. For instance, using ALBM, Zhuang *et al.* (2023) recently estimated that global lakes (larger than 0.1 km^2) emit 24.0 ± 8.4 Tg CH_4 yr^{-1}, which is at the lower end of the range reported by Saunois *et al.* (2020) and represents 11% of total global CH_4 emissions from natural sources as estimated from atmospheric inversions. Yet, these process-based models also have limitations that need to be addressed. A central limitation is the omission of lake phytoplankton productivity, which is one of the most reactive organic C sources and thus substrates for CH_4 production. In most of existing models, this key process and the associated microbial degradation of organic C is not simulated explicitly but taken as prescribed model inputs. If phytoplankton productivity and associated contributions of methane substrates can be incorporated in lake CH_4 models, this would allow capturing the impacts of environmental conditions beyond the commonly included direct temperature effects on organic matter decomposition and CH_4 production. Such additional important impacts include feedback of C metabolism on lake oxygen (O_2) cycling along with eutrophication effects on CH_4 emissions (Del Sontro *et al.*, 2018; Rosentreter *et al.*, 2021; Stavert *et al.*, 2022). However, including an explicit description of these processes is challenging, because it requires to account for a suite of key physical and biogeochemical processes controlling the coupled C-O_2-CH_4 cycles while at the same time maintaining model complexity, as well as the needs for observational data and computational costs for regional and global scale applications with tractable bounds. In addition, it also requires the

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quantification of nutrient inputs from the surrounding catchments, which exert a key control on lake productivity.

To tackle with these challenges, we here develop a new process-based model framework of intermediate complexity, FLaMe v1.0 (**F**luxes of **La**ke **Me**thane version 1.0, hereafter referred as FLaMe) that couples the C-O_2-CH_4 cycles in lakes using a one-dimensional representation. Specifically, FLaMe builds upon the existing physical lake model CSLM (Canadian Small Lake Model–MacKay, 2012; MacKay *et al.*, 2017), and extends with a novel biogeochemical module that captures the production, oxidation, storage, transport and emissions of CH_4 in/from lakes. Importantly, FLaMe introduces lake primary production and turnover of autochthonous organic carbon as a major driver of lake O_2 and CH_4 dynamics. The coupled, mechanistic lake model is then embedded in a computationally efficient clustering approach that allows for the application of the new, coupled, mechanistic lake model for (i) large parameter/input ensemble runs on regional/global scales for uncertainty assessments, (ii) long-term model projections for the assessment of future climate change and its feedback on the Earth system, (iii) a potential coupling to ESMs in subsequent stages of its development.

The structure of this paper is described as follows. In section 2, we provide a general description of the lake model with a focus on a detailed description of the novel biogeochemical modules, as well as the parameter choices and numerical solutions. In section 3, we first test the overall behavior of FLaMe using two representative lakes (an oligotrophic, deep lake driven by cold climate *versus* a trophic, shallow lake driven by warm climate), and then evaluate the simulated temporal variations of CH_4 fluxes against observational data at four well-surveyed lakes in real world. Next, we apply

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FLaMe at the European scale and evaluate the results against *in-situ* measurements in boreal and central European lakes compiled by Rinta *et al.* (2017). Finally, we provide a spatio-temporally resolved estimate of CH_4 emissions from European lakes (2010–2016), assess their sensitivity to key model parameters, and constrain their uncertainty range using a machine-learning approach. In section 4, we discuss model limitations and potential directions for further research. Main conclusions and outlooks are drawn in section 5.

2. Methodology

2.1 General model approach

We developed a new process-based physical-biogeochemical model, FLaMe v1.0 (Fluxes of Lake Methane), to simulate lake CH_4 production and emission at large spatial scales. FLaMe resolves the interplay of physical and biogeochemical processes that governs organic matter ($C_{OC,auto}$), oxygen (O_2), and methane (CH_4) dynamics to estimate (diffusive and ebullitive) lake CH_4 emissions, as well as CH_4 storage fluxes due to lake turnover and ice melting. To enable a continental-scale application of FLaMe (e.g., in Europe, n=108407 and total area = 1.33×10^5 km^2 for lakes with $0.1 \leqslant A_0 \leqslant 1000$ km^2 according to Messager *et al.*, 2016; A_0 is the lake surface area), we here propose a lake clustering strategy to reduce the computational and data/input costs (Fig. 1) while resolving the variability in lake sizes, morphology, and trophic status as well as climate conditions across Europe. Within each grid cell (2.5°×2.5°), lakes are binned into four classes according to surface area (0.1–1 km^2, 1–10 km^2, 10–100 km^2, 100–1000km^2). We then run a FLaMe set-up for one representative lake per size class within each grid cell, using the arithmetic means of lake area, depth and trophic status of all lakes pertaining to the respective size class across the respective grid cell. The total emission flux

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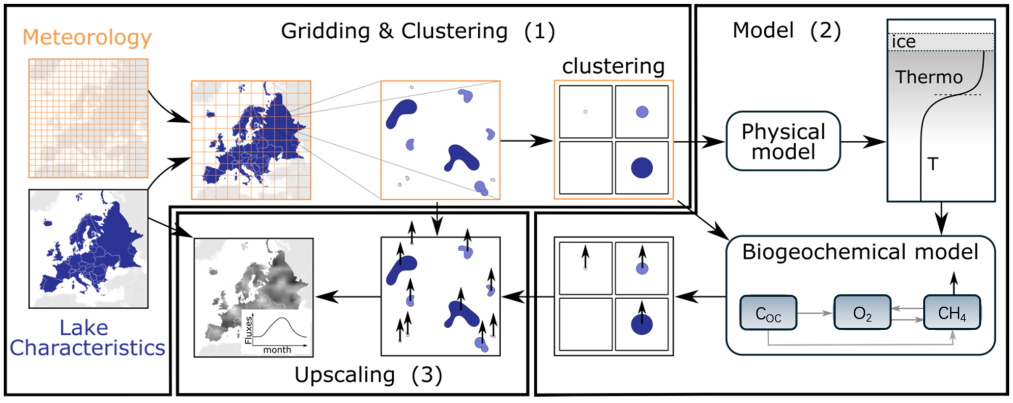
149 from a given size class can be obtained by multiplying the CH_4 emission rates simulated by FLaMe
150 with the total lake area of this size class (Fig. 1).



151

152 **Fig. 1. Illustration of the lake clustering and upscaling strategy for the continental application of FLaMe (Europe**
153 **as an example). (1) Gridding and clustering: The European domain was divided into grid cells at a coarse spatial**
154 **resolution of 2.5°×2.5°. Within each grid cell, lakes are clustered into four classes according to their surface areas.**
155 **(2) FLaMe parallelization: FLaMe simulates the lake metabolic dynamics, vertically resolved concentration and**
156 **rate profiles of the coupled O_2-CH_4 cycle as well as diffusive and ebullitive CH_4 fluxes through the water-air**
157 **interface. The model was parallelized under transient conditions for each grid cell and each lake class. (3)**
158 **Upscaling: The areal rates (i.e., fluxes per unit lake surface area) simulated by FLaMe were then multiplied by the**
159 **total surface area of each lake class within each grid cell (available from HydroLAKES) and aggregated at the**
160 **monthly timescale. The arrows pertaining to clustered and original lakes represent the CH_4 emissions.**

161 ## 2.2 Model description

162 FLaMe builds on an online coupling approach between the Canadian Small Lake Model (CSLM;
163 MacKay, 2012; MacKay *et al.*, 2017) – a widely used lake physics model (Garnaud *et al.*, 2022;
164 Verseghy and MacKay, 2017; William *et al.*, 2014) – and a set of newly developed biogeochemical

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modules that resolve lake OC, O_2 and CH_4 dynamics. CSLM simulates the physical variables of temperature profile (T), thermocline depth (h_{mix}), photic depth (h_{phot}), and ice cover, which will be used to force the biogeochemical modules (Fig. 2). In turn, the biogeochemical module will later modify the photic depth simulated by CSLM to account for the effect of phytoplankton growth and self-shading on light penetration, thus resolving the feedback between lake biogeochemical processes and lake physical dynamics, hence forming a complete feedback loop. A detailed description of the well-established CSLM model can be found in MacKay (2012) and MacKay *et al.* (2017) and is also briefly presented in Supplementary Text S1. In what follows, we provide a detailed description of the vertically resolved 1D model set-up (section 2.2.1) used here, as well as of the novel biogeochemical modules (section 2.2.2 and Table 1 for selected model parameter values).

2.2.1 Model Scope: Idealized representation of lake morphology

Figure 2 illustrates the vertically resolved, one-dimensional model set-up of FLaMe that is used for both the physical and biogeochemical modules. The original version of CSLM usually adopts a "bucket" shaped morphology which assumes a constant area (A) versus water depth (z), i.e., $A(z) = A_0$, where A_0 is the lake surface area at $z = 0$ m. However, this morphology is unsuitable for the simulation of biogeochemical processes, especially when the variations in water depth within each lake are important. Therefore, we, instead, adopted a "valley" shaped lake morphology, with lake area $A(z)$ given by:

$$A(z)=\frac{A_0}{2s}(s+\operatorname{arctanh}((1-2(z/h_{max}))\tanh(s))) \quad (1)$$

where A is the lake area (m^2), z is the water depth (m), s is a shape parameter that controls the slope of the lakebed (a larger s indicates a steeper slope), and h_{max} is the maximum lake depth. To ensure

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that the volume (and hence heat exchange) is conserved between the "bucket" and "valley" shape set-ups, the maximum depth of the valley-shape lake, h_{max}, must be twice that of the mean depth of the bucket-shape lake, h_{mean} (i.e., $h_{max} = 2h_{mean}$), which was extracted from the global HydroLAKES database compiled by Messager *et al.* (2016). The bottom temperature profiles simulated by CSLM were then extended to the maximal depth of the valley shape lake.

Physical processes in the water column are simulated by CSLM, on a one-dimensional, vertically resolved, evenly distributed grid with a grid spacing of 50 cm. Each water layer is connected to a vertically integrated lake sediment column of 5 m depth (h_s, m) (Fig. 2). Since the CH_4 production rate decreases exponentially with sediment depth, it is typically negligible at 5 m within the sediment column, thus ensuring that the total, depth-integrated benthic CH_4 production becomes insensitive to this arbitrary choice (Langenegger *et al.*, 2019). The surface area of each sediment column in contact with the water column is determined by difference in the widths of two adjacent water layers $A(z)$ (Eq. (1)). In addition, it should be noted that we assume no ***horizontal*** material exchanges (O_2 and CH_4) between the sediments and water columns (i.e., across the interface where left and right edges of a water layer touch the sediment box; Fig. 2), while only the vertical exchanges are simulated in this first version of the model (see details in the following sections).

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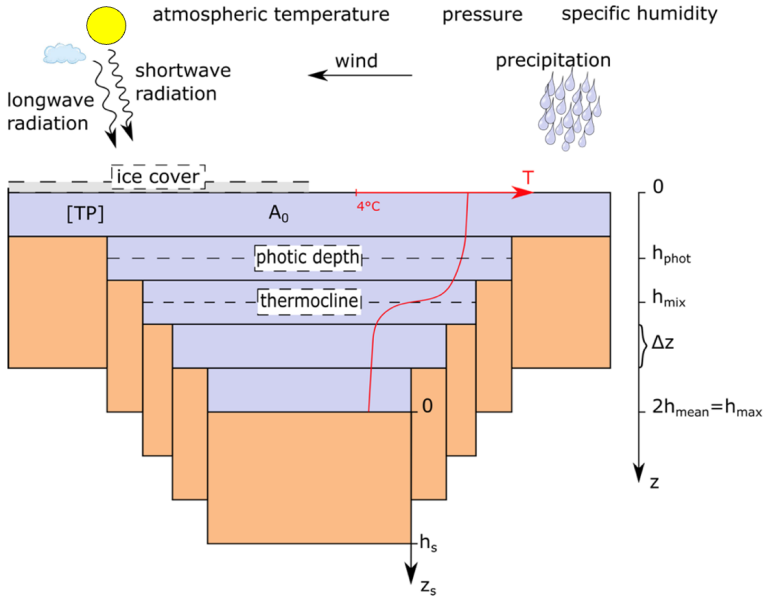

Fig. 2. Schematic representation of the lake morphology in FLaMe. The lake is represented by a "valley" shape (denoted by Eq. (1)). A_0 denotes the lake surface area, A is the area of each water layer, and h_{max} is the maximal water column depth. z represents the depth of a water column down to the surface of a sediment column while z_s stands for the depth inside a sediment column ($z_s = 0$ at the sediment water interface). The physical model is forced by longwave and shortwave radiation, near-surface wind, precipitation, atmospheric temperature, pressure, and specific humidity.

2.2.2 Biogeochemical Modules

2.2.2.1 Organic carbon module

Following the approach of Maavara *et al.* (2017), FLaMe does not resolve the vertical distribution of labile (i.e., microbial degradable) organic carbon (OC) concentrations ($[C_{OC,auto}]$) produced by in-lake primary production, but only simulates the temporal dynamics of the volume-integrated autochthonous OC stock ($\overline{C_{OC,auto}}$, g C) (the overbar here indicates a volume-integrated value). $\overline{C_{OC,auto}}$

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should be understood as a simple indicator or an overall reflection of the resulting lake trophic status, itself controlled by the combined effects of climate and nutrient loads from the catchment. The allochthonous C inputs delivered from surrounding catchments are more refractory and thus perform as less important substrates for CH_4 production (Guillemette *et al.*, 2017; DelSontro *et al.*, 2018). Thus, the autochthonous primary production is considered as the only labile OC source, neglecting the allochthonous OC contribution in this first version of FLaMe.

The temporal evolution of volume-integrated labile OC stock is determined by the interplay between autochthonous primary production, pelagic and benthic mineralization and burial fluxes (Maavara *et al.*, 2017):

$$\frac{\partial \overline{C_{OC,auto}}}{\partial t} = \overline{F_{PP}} - \overline{F_{Min}} - \overline{F_{Bur}} \quad (2)$$

where t is time (day), and $\overline{C_{OC,auto}}$ is the volume-integrated OC stock (g C). $\overline{F_{PP}}$, $\overline{F_{Min}}$ and $\overline{F_{Bur}}$ are the volume-integrated primary production, mineralization, and sedimentary burial fluxes (g C d^{-1}), respectively. Following Maavara et al. (2017), we assume that autochthonous primary production rates are controlled by the light regime, water temperature, and the lake total dissolved phosphorus (TDP) concentration ([TDP], g P m^{-3}) (Reynolds, 2006). The volume-integrated $\overline{F_{PP}}$ can then be expressed using a classical Michaelis-Menten formulation (Mavaara *et al.*, 2017):

$$\overline{F_{PP}} = B\, P_{Chl,max}\, M(T_{mean})\, \frac{[TDP]}{K_{s,P}+[TDP]}\, V_{phot} \quad (3)$$

where B is the phytoplankton biomass (expressed as chlorophyll-a concentration, g *Chl-a* m^{-3}) in the photic zone (Eq. (5)), $P_{Chl,max}$ is the maximum carbon fixation rate per unit of chlorophyll-a (g C (g Chla)$^{-1}$ h^{-1}), M is a dimensionless metabolic correction factor that depends on the mean lake water temperature in photic zone T_{mean} (°C) (see Eq. (4)), $K_{s,P}$ is the half-saturation constant for phosphorus

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limitation (g P m^{-3}), and V_{phot} is the water volume above the photic depth (m^3). Parameters $P_{Chl,max}$ and $K_{s,P}$ are constrained based on published observations (see section 2.3), while the metabolic factor M is given by:

$$M(T_{mean})=\begin{cases} 1, & T_{mean}\geq 28°C \\ Q_{10,prod}^{\frac{T_{mean}-28}{10}}, & T_{mean}<28°C \end{cases} \quad (4)$$

where $Q_{10,prod}$ is the temperature sensitivity for primary production, quantifying the increases of the metabolic factor per 10 degree increase in temperature. Surface water phytoplankton biomass, B, is approximated by a function of the photosynthetically active radiation (PAR), which is determined by shortwave radiation and light extinction in the water column:

$$B=\left(\frac{1}{k_c}\right)\left(0.75\left(\frac{PP}{RP}\right)\ln\left(\frac{0.7PAR_0}{0.5PAR_k}\right)\left(\frac{1}{h_{prod}}\right) - (K_{dw}+K_{dp}+K_{dg})\right) \quad (5)$$

where k_c is the absorbance of PAR per unit of chlorophyll-a (m^2 (g Chl-a)$^{-1}$), and PP/RP is the ratio of maximum gross photosynthesis to respiration per unit chlorophyll-a. PAR_0 is the PAR at the lake surface (µmol m^{-2} s^{-1}), determined by the incoming shortwave radiation, as well as the daytime that is specified by lake latitude and phenology (represented by the day of the year). PAR_k is the PAR at the onset of photosaturation (µmol m^{-2} s^{-1}). The productive depth h_{prod} is determined as the maximum of the thermocline and the photic depth simulated by the physical model. K_{dw}, K_{dp}, and K_{dg} represent nonalgal PAR attenuations, due to pure water, inorganic suspended particulate matter, and labile carbon, respectively. Following Lewis (2011), K_{dg} is calculated as a function of $[C_{OC,auto}]$ as:

$$\ln(K_{dg}) = -4.44 + 1.80\ln([C_{OC,auto}]) - 0.149(\ln([C_{OC,auto}]))^2. \quad (6)$$

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Eq. (5) was derived based on the assumption of a balance between production and respiration (Reynolds, 2006; Lewis, 2011). Here we slightly relax this assumption and assume near-equilibrium conditions for given climate conditions at the monthly timescale, allowing us to simulate seasonal variations of primary production and associated biogeochemical processes. Note that this assumption is only used for biogeochemical variables related to primary production, while physical variables simulated by CSLM are resolved at a sub-daily time step.

Following Hanson *et al.* (2011; 2014) and Maavara *et al.* (2019), the volume-integrated mineralization rate is simulated as a function of temperature and labile OC availability:

$$\overline{F_{Min}}=k_{20}\,\theta^{T_{mean}-20}\overline{C_{OC,auto}} \tag{7}$$

where k_{20} is a first-order rate constant for the mineralization of $\overline{C_{labile}}$ at 20 °C (d^{-1}). T_{mean} is the mean water temperature (°C) in photic zone, and θ is the temperature dependence of mineralization of organic matter (Hanson *et al.*, 2014).

Following Maavara *et al.* (2019), the burial flux $\overline{F_{Bur}}$ is represented by a first order process driven by the labile OC stock $\overline{C_{OC,auto}}$:

$$\overline{F_{Bur}}=k_{bur}\overline{C_{OC,auto}} \tag{8}$$

where k_{bur} is the burial rate constant and here set as half of the mineralization rate constant following the ratios between these two processes reported in the global lake dataset (n=230) assembled by Mendonça *et al.* (2017). This ratio is likely an upper bound because it accounts for contributions of both autochthonous and allochthonous carbon sources in the dataset, while allochthonous inputs

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should have distinct (but unquantified) ratios from autochthonous ones (Mendonça *et al.*, 2017; Guillemette *et al.*, 2017).

2.2.2.2 Methane module

The methane module simulates the dynamics of CH_4 concentration in both sediment and water columns as controlled by benthic CH_4 production, aerobic CH_4 oxidation, and diffusive and ebullitive transport from sediment to water and atmosphere (Fig. 3). Since the observational evidence suggests that benthic CH_4 production is the dominant CH_4 source in lakes (Tan *et al.*, 2015; Bastviken, 2022), we neglect the CH_4 production within the lake's water column in the model.

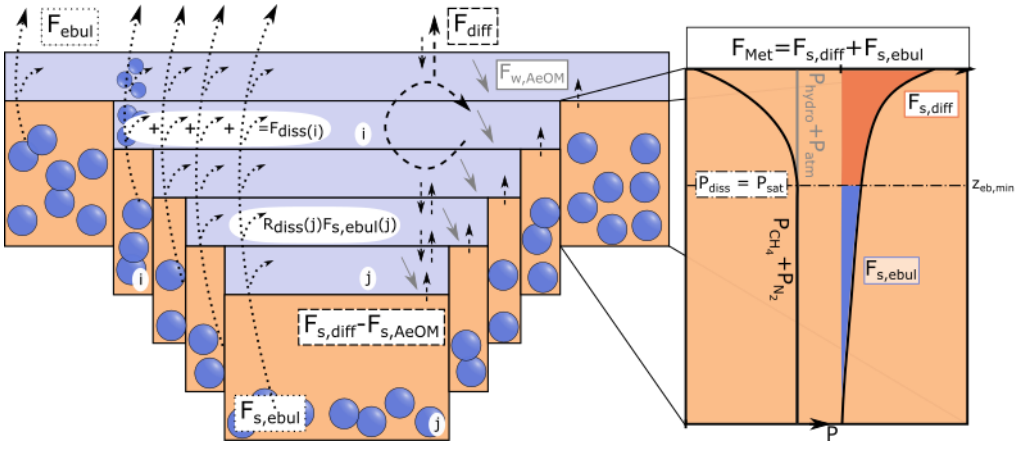

Fig. 3. Illustration of the methane (CH_4) module of FLaMe with a zoom into benthic CH_4 dynamics (zoom modified from Langenegger *et al.*, 2019). Benthic CH_4 production (zoom) assumes an exponential decrease in CH_4 production rate (F_{Met}) with depth. The CH_4 and N_2 partial pressures ($P_{CH4} + P_{N2}$) is mainly driven by the CH_4 production and follows the black curve profile, which starts to exceed the sum of the hydrostatic and atmospheric pressure ($P_{hydro} + P_{atm} - P_{H2O}$, grey line) at $z_{eb,min}$. Thus, this depth ($z_{eb,min}$) divides F_{Met} into a diffusive ($F_{s,diff}$, red filled region) and an ebullitive ($F_{s,ebul}$, cyan filled region) flux. $F_{s,AeOM}$ and $F_{w,AeOM}$ are the CH_4 oxidation in the

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sediment and water column, respectively. F_{diss} is the dissolution of the gas bubbles during transport through the water column. F_{diff} and F_{ebul} are diffusive and ebullitive CH_4 fluxes through the water-air interface, respectively.

Within the lake sediment, CH_4 dynamics are determined by the balance between CH_4 production via methanogenesis and CH_4 migration to the water column through diffusive and ebullitive transport:

$$\frac{\partial \widetilde{CH_{4,s}}(z)}{\partial t} = \widetilde{F_{Met}}(z) - \widetilde{F_{s,diff}}(z) - \widetilde{F_{s,ebul}}(z) \quad (9)$$

$$\widetilde{F_{Met}}(z) = f_{mm} \frac{M_{CH4}}{M_C} \overline{F_{Min}} \frac{V_s(z)}{V_{s,tot}} \quad (10)$$

where the tilde overbar here represents the volume-integrated stocks or fluxes in the sediment column, which is different from the straight overbar for volume-integrated values in the water column. Note that we have sediment columns at different water depths, such that the stocks and fluxes are represented as a function of water depth z, which is characterized by the valley-shape model set-up and different from the conventional bucket shape set-up. $\widetilde{CH_{4,s}}(z)$ is thus the volume-integrated CH_4 stock for the sediment column with the sediment-water interface positioned at depth z (g CH_4). $\widetilde{F_{Met}}(z)$ is the volume-integrated flux of CH_4 production through the entire sediment column with a sediment-water interface at depth z (g CH_4 d^{-1}), and $\widetilde{F_{s,diff}}(z)$ and $\widetilde{F_{s,ebul}}(z)$ are volume-integrated diffusive and ebullition fluxes (g CH_4 d^{-1}) through the sediment-water interface at depth z, respectively. f_{mm} denotes the fraction of organic matter mineralization that proceeds via methanogenesis according to data compiled by Hanson *et al.* (2014) and Bastviken (2022). M_{CH4}/M_C is a conversion factor corresponding to the molar ratio of CH_4 to $C_{OC,auto}$. As $f_{mm} \frac{M_{CH4}}{M_C} \overline{F_{Min}}$ is the total CH_4 production flux integrated over the whole water column, we assume that the fractions of

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CH_4 production occurring in different sediment columns are set according to their volume proportions, i.e., $\frac{V_s(z)}{V_{s,tot}}$.

The partitioning of CH_4 production into ebullitive and diffusive fluxes depends on the porewater CH_4 concentration or its partial pressure, which relies mainly on the vertical distribution of CH_4 production rate in the sediment as well as the oxygen concentration (but is of second-order importance). Based on the observation-based assumption that the organic carbon concentration and thus mineralization rates exponentially decrease with sediment depth, we here assume an exponentially decreasing relationship between methanogenesis rate versus depth (Fig. 3), following Langenegger *et al.* (2019):

$$f_{met}(z,z_s)=F_{Met,0}(z)\exp(-\alpha z_s) \quad (11)$$

where $f_{met}(z,z_s)$ is the methanogenesis rate (g CH_4 m^{-3} d^{-1}) at sediment depth z_s for the sediment column with the sediment-water interface positioned at depth z. $F_{Met,0}(z)$ is the maximum CH_4 production at the sediment-water interface (g CH_4 m^{-3} d^{-1}) at depth z, and α is a shape parameter (m^{-1}) that controls the decrease of methanogenesis rate with depth. As the shape of this curve typically depends on the flux of labile carbon settling on the lake bottom, and thus, lake trophic status, the parameter α is here scaled by the F_{PP} empirically:

$$\alpha = \alpha_{min}+\beta \cdot F_{PP}\ \frac{V_w}{V_{phot}} \quad (12)$$

where α_{min} is the minimum or base value, and β is the dependence of α on F_{PP}. The values of α_{min} and β are determined based on the measurements in lakes of different trophic status reported by Langenegger *et al.* (2019).

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To determine the maximum CH_4 production $F_{Met,0}(z)$, the integral of CH_4 production rate over sediment column should equal to the volume-integrated CH_4 production flux $\widetilde{F_{Met}(z)}$ as specified by Eq. (10):

$$A_s(z)\int_0^{h_s} f_{met}(z,z_s)\mathrm{d}z_s = \widetilde{F_{Met}(z)} \tag{13}$$

where $A_s(z)$ is the surface area of sediment column in contact with the water layer at lake depth z and is determined by difference in the areas of two adjacent water layers $A(z)$ (Eq. (1)). The maximum CH_4 production at depth z, $F_{Met,0}(z)$, can be obtained by combining Equations (10), (11) and (13):

$$F_{Met,0}(z) = \frac{\widetilde{F_{Met}(z)}}{A_s(z)} \frac{\alpha}{1-\exp(-\alpha h_s)} \tag{14}$$

Since CH_4 production increases the *in-situ* CH_4 concentration as the sediment depth increases, the CH_4 concentration may exceed its solubility concentration and methane gas bubbles may start forming (Fig. 3). To constrain the partitioning of CH_4 production between diffusion and ebullition, the threshold sediment depth, $z_{eb,min}$, at which CH_4 concentration starts to exceed the solubility limit, is determined based on the equilibrium pressure condition following Langenegger *et al.* (2019) (see details in Supplementary Text S2). In the upper portion of the sediment column ($z_s < z_{eb,min}$), the produced CH_4 will diffuse into water; however, a fraction of the diffusing CH_4 will be oxidized in the transit through the upper sediment column, and only the remaining CH_4 will reach the sediment-water interface. The volume-integrated CH_4 oxidation in the sediment at depth z, $\widetilde{F_{s,AeOM}(z)}$, is here assumed to be controlled by the O_2 concentration in the overlying bottom water, and is represented by a Michaelis-Menten function:

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$$\widetilde{F_{s,AeOM}}(z) = \widetilde{F_{Met}}(z)\frac{[O_2]_z}{K_{s,O2}+[O_2]_z} \quad (15)$$

where $K_{s,O2}$ is the half-saturation constant of O_2 for the sedimentary CH_4 oxidation. As a result, the diffusive flux passing through the sediment-water interface is determined as follows:

$$\widetilde{F_{s,diff}}(z) = A_s(z)\int_0^{z_{eb,min}} F_{Met,0}(z)\exp(-\alpha z_s)\,dz_s - \widetilde{F_{s,AeOM}}(z) \quad (16)$$

In the lower portion of the sediment column ($z_s > z_{eb,min}$; where oversaturation occurs), the produced CH_4 feeds the ebullitive flux, with the volume-integrated value $\widetilde{F_{s,ebul}}(z)$ (g CH_4 d^{-1}) as given by:

$$\widetilde{F_{s,ebul}}(z) = A_s(z)\int_{z_{eb,min}}^{h_s} F_{Met,0}(z)\exp(-\alpha z_s)\,dz_s \quad (17)$$

Note that Equations. (16) and (17) implicitly imply that, at the monthly resolution of our model, the CH_4 dynamics in the sediment is at steady state and all the CH_4 produced during this time interval is either oxidized or released through the water column via diffusive and ebullitive pathways.

Pelagic, dissolved CH_4 diffuses in the water column and its concentration is determined by the diffusive CH_4 flux passing through the sediment-water interface (acting as a source for each water layer), by aerobic CH_4 oxidation in the water column, and by the re-dissolution of the ebullitive CH_4 fluxes during transit through the water column. The mass conservation equation of dissolved CH_4 is then given by:

$$\frac{\partial[CH_4]_w}{\partial t} = \frac{\partial}{\partial z}\left(K_{diff}\frac{\partial[CH_4]_w}{\partial z}\right) + \widetilde{F_{s,diff}}(z)\frac{1}{A(z)dz} - F_{w,AeOM}(z) + F_{diss}(z) \quad (18)$$

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where $[CH_4]_w$ is the pelagic CH_4 concentration (g CH_4 m^{-3}) and K_{diff} is the eddy diffusion coefficient of CH_4 in water (m^2 d^{-1}). $\widetilde{F_{s,diff}(z)}\frac{1}{A(z)dz}$ is the change of CH_4 concentration induced by diffusive inputs from the sediment columns, the term $A(z)dz$ being the volume of the water layer connected to the corresponding sediment column. $F_{w,AeOM}(z)$ is the aerobic CH_4 oxidation rate in the water column, and is described through double Michaelis-Menten reaction kinetics (Stepanenko *et al.*, 2016; Liikanen *et al.*, 2002; Thottathil and Prairie, 2019):

$$F_{w,AeOM}(z)=k_{max}\,Q_{10,ox}^{\frac{T-T_r}{10}}\frac{[CH_4]_{w,z}}{K_{s,CH_4}+[CH_4]_{w,z}}\frac{[O_2]_z}{K_{s,O_2}+[O_2]_z} \quad (19)$$

where k_{max} is the maximum CH_4 oxidation rate (Liikanen et al. 2002), T is the water temperature, T_r is the reference temperature, and $Q_{10,ox}$ expresses the temperature dependency of the CH_4 oxidation rate. $K_{s,CH4}$ and $K_{s,O2}$ are the half-saturation constants for CH_4 and O_2, respectively.

To constrain the redissolution of gas bubbles ($F_{diss}(z)$), we follow the approach proposed by McGinnis *et al.* (2006) where a function ($f_{bdiss}(z)$) is used to represent the fraction of the benthic ebullitive CH_4 flux that redissolves in the water column during gas ascent. This fraction is a function of water depth and gas bubble diameter, and the latter was set to 5 mm following Delwiche and Hemond (2017). With this function $f_{bdiss}(z)$, the redissolved CH_4 fluxes from sediment column at depth z are assumed to be evenly redistributed in the water layers above the sediment, i.e.,

$$f_{rediss}(z)=\frac{f_{bdiss}(z)\widetilde{F_{s,ebul}(z)}}{\int_0^z A(z)dz} \quad (20)$$

where $\int_0^z A(z)dz$ is the water volume above the sediment layer at the depth of interest z. Then, at this particular depth z, the redissolution flux (F_{diss}, g CH_4 m^{-3} d^{-1}) is thus determined as follows:

$$F_{diss}(z)=\int_z^{h_{max}} f_{rediss}(z)dz \quad (21)$$

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where $\int_z^{h_{max}} f_{rediss}(z)dz$ represents the integral of all re-dissolved ebullitive fluxes from sediment columns below z.

By deducing this dissolution flux from the ebullitive flux released from lake sediments, the resultant ebullitive flux reaching the atmosphere (F_{ebul}; g CH_4 m^{-2} d^{-1}) is calculated as:

$$F_{ebul} = \frac{1}{A_0} \int_0^{h_{max}} \left(1 - f_{bdiss}(z)\right) \widetilde{F_{s,ebul}}(z)dz \quad (22)$$

where A_0 is the lake surface area, and $\left(1 - f_{bdiss}(z)\right) \widetilde{F_{s,ebul}}(z)$ is the component of ebullitive flux at depth z that reaches the atmosphere.

In addition to diffusive and ebullitive pathways, FLaMe also calculates a storage flux (F_{stor}) that encapsulates the changes in the total CH_4 mass stored in hypolimnion due to the weakening of lake stratification or turnover events when the lake surface temperature crosses the threshold of 4°C (MacKay, 2012; MacKay *et al.*, 2017). This results in a full mixing of the lake that releases the previously accumulated CH_4 in the anoxic portion of the lake and concomitantly fully aerates the water column. Lake turnovers thus lead to a complete homogenization of O_2 and CH_4 concentration across the vertically resolved water column. Upon full mixing, remobilization of larger CH_4 stocks that accumulated in the hypolimnion abruptly increase the CH_4 concentration near the lake surface, and hence strongly enhance the diffusive flux through the air-water interface. That is, the storage flux in FLaMe is not simulated separately but it is implicitly incorporated into the diffusive flux F_{diff} which increases dramatically following the formation of a very sharp CH_4 concentration gradient at the lake surface.

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2.2.2.3 Oxygen module

The oxygen module is needed to simulate the lake methane processes (section 2.2.2.2). It represents the O_2 cycle within the water column, driven by O_2 production by photosynthesis, O_2 consumption by pelagic and benthic OC mineralization, and aerobic pelagic and benthic CH_4 oxidation. These processes are coupled to the vertical diffusive transport of O_2 through water column (Fig. 4). The one-dimensional conservation equation for O_2 concentration in the water column is thus given by:

$$\frac{\partial[O_2]}{\partial t}=\frac{\partial}{\partial z}\left(K_{diff}\frac{\partial[O_2]}{\partial z}\right)+OF_{PP}(z)-OF_{w,Aer}(z)-\frac{1}{A(z)dz}\widetilde{OF_{s,Aer}}(z)-OF_{w,AeOM}(z)-OF_{s,AeOM}(z) \quad (23)$$

where $[O_2]$ is the O_2 concentration in the water (g O_2 m^{-3}), and K_{diff} is the eddy diffusion coefficient of O_2 (m^2 d^{-1}), assumed identical to that of CH_4. $OF_{PP}(z)$ is the oxygen production through primary production (g O_2 m^{-3} d^{-1}) at depth z. $OF_{w,Aer}(z)$ is the O_2 consumption by heterotrophic respiration (g O_2 m^{-3} d^{-1}) in the water column at depth z, while $\widetilde{OF_{s,Aer}}(z)$ is the volume-integrated O_2 consumption by heterotrophic respiration in the sediment (g O_2 m^{-3} d^{-1}), and $A(z)dz$ is the volume of the water layer connected to the corresponding sediment column. $OF_{w,AeOM}(z)$ and $OF_{s,AeOM}(z)$ are the aerobic CH_4 oxidation in the water column and sediment (g O_2 m^{-3} d^{-1}), respectively, at depth z.

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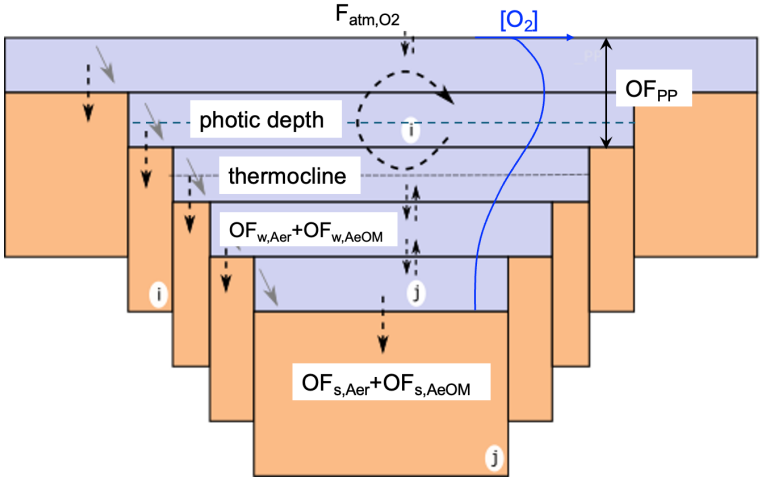

423

424 **Fig. 4. Illustration of the oxygen (O_2) module in the FLaMe model. The O_2 production due to primary production**
425 **occurs only in the photic zone (OF_{PP}), while the O_2 consumption by heterotrophic respiration occurs in both the**
426 **entire pelagic zone and benthic zone ($OF_{w,Aer}$ and $OF_{s,Aer}$). The O_2 consumption due to CH_4 oxidation occurs also**
427 **in both pelagic and benthic zones ($OF_{w,AeOM}$ and $OF_{s,AeOM}$). In this figure, the dotted arrows crossing the sediment-**
428 **water interface represent the O_2 demands in sediments ($OF_{s,Aer}$ and $OF_{s,AeOM}$), the dashed arrows represent the**
429 **eddy diffusion of O_2 between water layers and through the water-air interface, and the tilted grey arrows represent**
430 **the aerobic oxidation of CH_4 in the water column. As a result, the blue curve depicts a typical vertical profile of**
431 **O_2 concentration under lake water stratification.**

432 Photosynthesis occurs only in the photic zone, and the amount of O_2 produced by primary
433 production $\overline{OF_{PP}}$ (volume-integrated value; g O_2 d^{-1}) can be determined according to the
434 stoichiometric ratio M_{O2}/M_C, where and M_{O2} and M_C are the molar masses of oxygen and carbon,
435 respectively. To resolve the vertical O_2 profile, the O_2 produced during primary production is
436 assumed to be evenly redistributed within the water layers above the photic depth (Fig. 4):

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$$OF_{PP}(z) = \begin{cases} \overline{F_{PP}} \frac{1}{V_{phot}} \frac{M_{O_2}}{M_C}, & z < z_{phot} \\ 0, & z \geqslant z_{phot} \end{cases} \quad (24)$$

where V_{phot} is the photic volume.

The oxygen consumption induced by CH_4 oxidation in the sediment and water column can be calculated from corresponding CH_4 fluxes (Eqs. (15) and (19), respectively) and the stoichiometry of the reactions involved:

$$OF_{s,AeOM}(z) = \frac{2M_{O_2}}{M_{CH_4}} F_{s,AeOM}(z) \quad (25)$$

$$OF_{w,AeOM}(z) = \frac{2M_{O_2}}{M_{CH_4}} F_{w,AeOM}(z) \quad (26)$$

As in Eq. (10), a fraction of the mineralized organic carbon (represented by f_{mm}) is channeled into the methanogenesis pathway according to the data compiled by Hanson *et al.* (2014) and Bastviken (2009). Thus, the remaining fraction ($1-f_{mm}$) of the total mineralization $\overline{F_{Min}}$ is channeled into the aerobic metabolic pathway (F_{Aer}). As a result, the *bulk volumetric rate* of oxygen consumption due to the aerobic metabolic activity (OF_{Aer}) can be represented by the fraction $1-f_{mm}$ and the volume-integrated mineralization $\overline{F_{Min}}$:

$$OF_{Aer} = (1-f_{mm}) \overline{F_{Min}} \frac{1}{V_w} \frac{M_{O2}}{M_C} \quad (27)$$

In the sediment, the aerobic mineralization occurs only in the upper oxic layer. The thickness of this aerobic layer is limited by the oxygen penetration depth z_{ox}. Following Ruardij and Van Raaphorst (1995), this depth z_{ox} can be derived by solving the steady-state reaction-diffusion equation for O_2 in the sediment:

$$z_{ox} = \sqrt{\frac{2K_{s,diff}}{OF_{s,AeOM} + OF_{Aer}}} \quad (28)$$

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where $K_{s,diff}$ is the molecular diffusion coefficient within the sediment, which is dependent on the temperature (Ruardij and Van Raaphorst,1995). The amount of O_2 consumed within the oxic layers of the sediment can thus be determined as:

$$\widetilde{OF_{s,Aer}}(z)=OF_{Aer}A_s(z)z_{ox} \quad (29)$$

where $A_s(z)$ is the area of the corresponding sediment column at depth z. To ensure a mass balance, the volumetric rate of O_2 consumption due to aerobic metabolism in water can then be calculated as follows:

$$OF_{w,Aer}(z)=OF_{Aer}-\widetilde{OF_{s,Aer}}(z)\frac{1}{A(z)dz} \quad (30)$$

where $A(z)dz$ is the volume of the water layer connected to the corresponding sediment column, and it is used here to convert the sedimentary O_2 consumption into a volumetric rate in the water column. Furthermore, following Martin *et al.* (1987), Carlson *et al.* (1994) and Arístegui *et al.* (2003), we redistribute the respiration ($OF_{w,Aer}$) within the water column, assuming that 80% of the respiration occurs in the photic zone, with the remaining 20%, sustained by the export production, occurs in the deeper water layers where it can further degrade.

2.2.3 Boundary conditions for the transport module

The partial differential equations (18) and (23) require boundary conditions to constrain the diffusive transport (i.e., the first term on the right-hand side of both equations). At the sediment-water interface, a zero-flux boundary condition is imposed, because the diffusive exchanges of CH_4 and O_2 between the sediment columns and the overlying waters are already included as source/sink terms in Eq. (18) and (23). This choice was guided by the valley-shape configuration of our lake set-up, and thus by the presence of diffusive CH_4 and O_2 exchange fluxes with sediment in each water

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layer of our model, a situation in stark contrast from a bucket shape model where only a single sediment column would be connected to the bottom water layer.

At the lake surface (z = 0 m), the boundary conditions are determined by the CH_4 and O_2 exchange fluxes with the atmosphere, as given by (Wanninkhof et al., 2009; Cole and Caraco, 1998):

$$F_{\mathrm{atm,CH_4}}=k_{ge}([\mathrm{CH_4}]-f_{CH4,atm}P_{\mathrm{atm}}M_{CH4}K_{H,CH4}\exp(\frac{\partial \ln(K_{H,CH4})}{\partial\frac{1}{T}}(\frac{1}{T_1}-\frac{1}{298.15}))) \quad (31)$$

$$F_{\mathrm{atm,O_2}}=k_{ge}([O_2]-f_{O2,atm}P_{\mathrm{atm}}M_{O2}K_{H,O2}\exp(\frac{\partial \ln(K_{H,O2})}{\partial\frac{1}{T}}(\frac{1}{T_1}-\frac{1}{298.15}))) \quad (32)$$

where $F_{\mathrm{atm},CH4}$ and $F_{atm,O2}$ are diffusive fluxes of CH_4 (g CH_4 $m^{-2}d^{-1}$) and O_2 (g O_2 m^{-2} d^{-1}) through the air-water interface of the lake, respectively. $f_{CH4,atm}$ and $f_{O2,atm}$ are molar fractions of CH_4 and O_2 in the atmosphere, respectively, and P_{atm} is the atmospheric pressure. $K_{H,CH4}$ and $K_{H,O2}$ are Henry's constants of CH_4 and O_2 at 298.15 K and k_{ge} is the piston velocity (m s^{-1}), here constrained from the empirical equation reported by Cole and Caraco (1998), as in Tan *et al.* (2015; 2018) and Stepanenko *et al.* (2016):

$$k_{ge}=(C_{k_1}+C_{k_2}v_{a,10}^{n})\sqrt{\frac{600}{S_{CX}}} \quad (33)$$

where C_{k1}, C_{k2} and n are empirical constants (Cole and Caraco, 1998). $v_{a,10}$ is the absolute wind velocity measured at 10 m above the lake surface (m s^{-1}), and $S_{c,CH4}$ *and* $S_{c,O2}$ are the Schmidt number of CH_4 and O_2, respectively (Wanninkhof et al. 2009). Note that more recently new formulations of k_{ge} have been proposed (McIntire et al., 2020) but we here choose to use Eq. (33) following previous modelling studies (Tan et al., 2015; Stepanenko *et al.* 2016; Tan *et al.*, 2018).

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2.3 Parameter values

Table 1 summarizes all physical and biogeochemical parameters, their values, as well as the original references from which they were extracted. Most of these parameters were either directly taken from relevant modelling studies or constrained based on comprehensive literature reviews. In addition, several key parameters of the FLaMe model, highlighted in Table 1, were adjusted by calibrating the model based on observations of lake C fluxes (i.e., F_{PP}, diffusive and ebullitive CH_4 emissions). For instance, the parameters $P_{Chl,max}$ and $K_{s,P}$ control the lake primary production and were tuned to reproduce broad global patterns of primary production rates across the full range of lake trophic status (Wetzel, 2001). The mineralization k_{20} and burial constants k_{bur} were adjusted based on the observed fraction of $C_{OC,auto}$ that settles onto the lake sediment, either to be decomposed in anerobic or oxic conditions or accumulated in the sediment (Hanson *et al.*, 2011, 2014; Maavara *et al.*, 2019; Mendonça *et al.*, 2017). The temperature dependence of mineralization θ was fine-tuned to reproduce the observational ranges of temperature dependence of net-CH_4 emissions (Aben et al., 2017). f_{mm} specifies the fraction of mineralization that channels to the methanogenesis pathway, which is adjusted to produce the observational patterns of CH_4 emissions. α_{min} is the base value of the exponentially decreasing rate of CH_4 production versus sediment depth, and controls the split of CH_4 production between diffusive and ebullitive pathways, which was calibrated to reproduce observed broad trends of F_{tot}, F_{ebul} and F_{diff} from the literature (Rinta *et al.*, 2017). The parameter values listed in Table 1 provide the reference setup for the simulation of lake CH_4 emissions, and the sensitivity and uncertainty analyses regarding the key model parameters (listed in Table 3) is carried out using wide ranges of values covering most possible lake conditions from the real world (see section 3.3.3).

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2.4 Numerical solution

In FLaMe, the physical (i.e., CSLM) and biogeochemical (OC, CH_4 and O_2) modules are coupled online. For the dynamics of volume-integrated OC and CH_4 in sediments, the involved ordinary differential equations are solved using a forward Euler scheme. For the dynamics of dissolved O_2 and CH_4 concentrations in the water column, the partial differential equations (Eqs. (18) and (23)) are solved numerically using an explicit central difference scheme for depth and Euler forward scheme for time. The diffusion coefficient K_{diff} for both O_2 and CH_4 is depth dependent (Table 1) to capture the reduced transport when the temperature gradient from the epilimnion, metalimnion and hypolimnion is well pronounced (Dong et al. 2020; Imboden and Wuest 1995; Imberger 1985; Boehrer and Schultze 2008).

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527 **Table 1. Model parameters of FLaMe v1.0 and the choice of their values**

Main processes	Key model parameters	Physical meanings (units)	Values	Equations	References
Lake morphology	s	Steepness of lakebed (-)	2	(1)	-
Primary production	$P_{chl,max}$	Maximum carbon fixing rate per unit of Chlorophyll-a (mg C (mg Chl-a)$^{-1}$ h^{-1})	0.5*	(3)	Behrenfeld and Falkowski (1997)
	$K_{s,P}$	Half saturation coefficient of total dissolved phosphorus for the primary production (g m^{-3})	0.09*	(3)	Maavara *et al.*, (2017)
	$Q_{10,prod}$	Temperature sensitivity for the primary production	2	(4)	Lewis (2001) and Reynolds (2006)
	k_c	Absorbance of PAR per unit of chlorophyll-a (m^2 (g Chl-a)$^{-1}$)	0.014× 10^3	(5)	Lewis (2001) and Reynolds (2006)
	PP/RP	ratio of maximum gross photosynthesis to respiration per unit chlorophyll-a (-)	15	(5)	Lewis (2001) and Reynolds (2006)
	PAR_k	PAR at the onset of photo saturation (µmol m−2 s−1)	120	(5)	Lewis (2001) and Reynolds (2006)
	K_{dw}	PAR attenuations due to pure water	0.13	(5)	Lewis (2001) and Reynolds (2006)
	K_{dp}	PAR attenuations due to suspended particulate matter	0.06	(5)	Lewis (2001) and Reynolds (2006)
Mineralization and burial of organic carbon	k_{20}	Mineralization rate at a reference temperature of 20 °C (d^{-1})	0.008*	(7)	Maavara *et al.*, (2017)
	θ	Temperature dependence of mineralization	1.02*	(7)	Maavara *et al.*, (2017)

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	k_{bur}	Carbon burial rate in the lake (d^{-1})	0.004*	(8)	Mendonca *et al.*, (2017)
	f_{mm}	Fraction of mineralization that channels to the methanogenesis pathway	1/4*	(10) and (27)	Hanson *et al.* (2014); Bastviken (2009)
CH_4 oxidation	k_{max}	Maximal rate of CH_4 oxidation (g CH_4 m^{-3} d^{-1})	0.69	(19)	Liikanen et al. (2002)
	$Q_{10,ox}$	Temperature dependence of CH_4 oxidation (-)	1.2	(19)	Liikanen et al. (2002)
	$K_{s,CH4}$	Half-saturation constant for CH_4 (g CH_4 m^{-3})	0.6	(19)	Stepanenko et al. (2016)
	$K_{s,O2}$	Half-saturation constant for O_2 (g O_2 m^{-3})	0.67	(19)	Liikanen et al. (2002)
Shape parameter of sedimentary CH_4 production	α_{min}	Base value of the exponentially decreasing rate of CH_4 production versus sediment depth (m^{-1})	10*	(12)	Langenegger *et al.*, (2019)
Gas transport in the water column and exchange with air	K_{diff}	Depth-dependent eddy-diffusion coefficient (m^2 d^{-1})	8.64 (epilimnion), 8.64×10^{-3} at the termocline, and 8.64×10^{-1} (hypolimnion)	(18) and (23)	Stefan and Fang (1994)
	C_{k1}	Empirical constant for piston velocity (m s^{-1})	5.75×10^{-6}	(33)	Cole and Caraco, (1998)
	C_{k2}	Empirical constant for piston velocity (m s^{-1})	5.97×10^{-7}	(33)	Cole and Caraco, (1998)
	n	Empirical constant for piston velocity	1.7	(33)	Cole and Caraco, (1998)
	$S_{c,CH4}$	Schmidt number of CH_4 (-)	677	(33)	Wanninkhof *et al.* (2009)

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	$S_{c,O2}$	Schmidt number of O_2 (-)	589	(33)	Wanninkhof *et al.* (2009)
	$f_{CH4,atm}$	Atmospheric molar fractions of CH_4	0.18×10^{-13}	(31)	Lan *et al.* (2024)
	$f_{O2,atm}$	Atmospheric molar fractions of O_2	0.2095	(32)	Gatley *et al.* (2008)

528 * indicates that the original parameter values are from the literature, and further adjusted by calibration
529 versus observations.

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2.5 Case studies

We implemented three case studies to assess the performance of FLaMe in simulating lake CH_4 emissions, as well as its application to the European scale. First, we present theoretical simulations for two representative cases (methodological details in section 2.5.1) to assess the general behaviors of FLaMe in capturing the physical-biogeochemical patterns of contrasted lakes. Then, we perform the simulations for four well-surveyed real lakes to assess the model's capability in capturing the observed temporal variations of CH_4 fluxes (section 2.5.2). Next, we apply FLaMe to the entire European domain to assess the model's capability in reproducing the spatial patterns and seasonal variations of CH_4 fluxes at continental scale (section 2.5.3). The European scale application can be considered as a "proof of concept" in support of our proposed modeling strategy. Finally, we examine the sensitivity to key model parameters and assess the uncertainty of the continental-scale emissions using the samples produced by sensitivity analysis, combined with a machine learning approach (section 2.5.4).

2.5.1 Two theoretical representative lakes for testing FLaMe performance

To test if the FLaMe model can capture the contrast in physical-biogeochemical behaviors across shallow vs. deep, eutrophic vs. oligotrophic and warm vs. cold lakes, we set-up the model for two theoretical representative lakes: a "deep oligotrophic" lake (h_{max} = 35 m or h_{mean} = 17.5 m and [TP] = 3 µg P L^{-1}) driven by a "cold" climate (63.75°*N,* 26.25°*E*; Fig. S1) and a "shallow eutrophic" lake (h_{max} = 10 m or h_{mean} = 5 m and [TP] = 80 µg P L^{-1}) driven by a "warm" climate (43.75°*N,* -6.25°*E*; Fig. S2). For these two theoretical representative cases, FLaMe simulates the spatio-temporal evolutions of physical and biogeochemical variables and fluxes, including primary production and mineralization fluxes, and labile autochthonous OC stocks as well as the vertically resolved gradients

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of temperature, CH_4 and O_2 concentrations. Furthermore, we also compared the seasonal patterns of CH_4 productions and emissions for these two contrast lakes. To investigate further how environmental factors affect the FLaMe model behavior, we further decompose the collective responses of shallow and deep lakes into individual effects induced by trophic level, climate (Fig. S1–S3) and lake depth using hypothetical numerical simulations, i.e., (i) changing the maximal lake depth (h_{max}) from 5 to 25 m; (ii) increasing the [TP] levels from 8 to 80 µg P L^{-1}; and (iii) changing the climate from warm (43.75°*N,* -6.25°*E;* Fig. S1) to cold conditions (63.75°*N,* 26.25°*E*; Fig. S2).

2.5.2 Simulations of temporal patterns for four well-surveyed lakes

To evaluate the ability of FLaMe to reproduce the observed temporal patterns of CH_4 fluxes, we selected four lakes from the Inter-Sectoral Impact Model Intercomparison Project (ISIMIP) lake datasets for which monthly resolved temporal CH_4 fluxes were available. These lakes cover different lake depths, areas, climate conditions and trophic statuses, as summarized in Table 2. Since *in-situ* measurements of climatic drivers are not available for these lakes, we extracted them from the 0.5°x0.5° GSWP3-W5E5 global climate forcings released by the ISIMIP3a project as an approximation. The measurements of CH_4 fluxes for these lakes were mostly collected during the first 20 years of the 21st century, and we thus selected the climate forcings for the period 1991–2019, using the period 1991–1999 as spin-up phase. Since the lack of *in-situ* measured climatic drivers, as well as observations of variations in lake water levels and areas, affect the model's ability to capture the full variability in the observed CH_4 emission time series, we here focus only on the magnitudes and broad seasonal patterns in observed CH_4 emissions. Thus, we evaluated the simulated statistics (mean and SD represented by boxplots) of CH_4 fluxes within the year against the observational data.

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574 **Table 2. Characteristic information for the four well-surveyed lakes from ISIMIP**
575 **datasets**

Lake	Coordinates	Lake depth	Lake area (km^2)	Climate	Trophic status	Data
Klöntal	47.026N, 8.981E	21.4m (mean), 45m (max)	2.25	Temperate	Oligotrophic	Annual mean
Erssjön	58.371N, 12.162E	1.3m (mean), 4.75m (max)	0.062	Temperate-Boreal	Mesotrophic	2012–2013
Upper Mystic	42.434N, 71.150W	11.7m (mean), 25m (max)	0.58	Temperate	Eutrophic	2007–2008
Villasjön	68.35N, 19.03E	1.3 m (max)	0.17	Boreal	Oligotrophic	2010–2017

576 **2.5.3 Implementation of FLaMe at continental scale**

577 To implement the model at the scale of Europe (25°W–60°E, 36°–71°N), we extracted the lakes
578 within this domain from the HydroLAKES database (Messager *et al.*, 2016; n=108407, total area =
579 1.33×10^5 km^2 for lakes with $0.1 \leq A_0 \leq 1000$ km^2 within the European domain). Following our
580 clustering strategy, we subdivided, within each grid cell, all lakes into four classes based on their
581 surface area ($0.1 < A_0 < 1$ km^2, $1 < A_0 < 10$ km^2, $10 < A_0 < 100$ km^2, and $100 < A_0 < 1000$ km^2). As FLaMe
582 was derived from the small lake physics model CSLM, we here only considered the lakes with an
583 area smaller than 1000 km^2, and excluded the very large lakes ($A_0 > 1000$ km^2) that account for 40%
584 of the total European lake surface area (but only consist of 21 lakes). Within our model domain, we
585 have 108407 lakes with a surface area larger than 0.1 km^2, which at spatial resolution of 2.5 degree
586 (Fig. S4–S5) result in 365 grid cells and 953 representative lakes (hence reducing computation cost
587 by more than a factor of 100 compared to a case where each individual lake would be simulated).
588 By parallelizing the model simulations on a high-performance cluster, the implementation of FLaMe

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for the entire European domain consumes approximately 365 CPU hours for a single run covering 10 years.

The FLaMe model was forced by meteorological conditions from the GSWP3-W5E5 reanalysis product under ISIMIP3a (Frieler *et al.*, 2024) (Fig. S6), including shortwave solar radiation (W m^{-2}), longwave solar radiation (W m^{-2}), precipitation (mm s^{-1}), near surface air temperature (at 10 m height, °C), specific humidity (kg kg^{-1}), near surface wind velocity (at 10m, m s^{-1}), and atmospheric pressure (Pa). As these forcings were provided at a finer spatial resolution of 0.5°, we only applied the values in the central 0.5° grid cell of our larger 2.5° grid. In addition, the FLaMe model was also driven by the TDP in the representative lakes (Fig. S7–S8), which was estimated by dividing the TDP mass outflow by the water discharge reported in HydroLAKES, hence assuming that the lake water is well mixed. The TP mass outflow from each lake in HydroLAKES was obtained by routing the TP loads (extracted from the GlobalNEWS model; Mayorga *et al.*, (2010)) from the watershed (point and non-point terrestrial sources) into the river network, following the procedure outlined in Lauerwald *et al.* (2019) and topological information provided by the HydroSHEDS drainage network. More details related to the TP routing can be found in Bouwman and Billen (2009), Van Drecht *et al.* (2009), and Mayorga *et al.* (2010).

2.5.4 Sensitivity and uncertainty analysis

To explore how model parameterization affects the European-scale assessments of lake CH_4 emissions, we conducted a sensitivity analysis encompassing the parameters whose variations induce the largest changes in lake CH_4 dynamics and listed in Table 3. The sensitivity was conducted by varying a parameter once at a time: only one parameter is varied with the other parameters kept

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unchanged. All these parameters were assumed to have Gaussian distributions, with their SDs specified as 50% of their original values, except the temperature dependency $Q_{10,ox}$ and θ whose SDs were specified as 50% of their deviation to unity. More specifically, we tested the sensitivity within the ranges of mean±SD at four points, i.e., +SD, +0.5SD, -0.5SD, and -SD.

To constrain uncertainties in European scale CH_4 emissions, we complemented the sensitivity analysis (n=36) with another 28 scenarios under several extreme cases and different combinations of variations in key parameters. With these 64 assessments taken as samples, we then used a machine learning approach to assess the uncertainty associated to our estimation of European lake CH_4 fluxes. Specifically, we trained a Random Forest (RF) model that capture nonlinear relationships between our key model parameters and European lake CH_4 emissions, i.e., the key parameters are taken as predictors and the European lake CH_4 emissions are taken as target variable. Next, we produced 1000 Gaussian-distributed random samples within the parameter space and estimated an ensemble of CH_4 emissions using the trained RF model.

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623 **Table 3 Key model parameters selected for sensitivity and uncertainty analysis**

Main processes	Key model parameters	Physical meanings (units)	Values	Ranges	Eq.	References
Primary production	$P_{chl,max}$	Maximum carbon fixing rate per unit of Chlorophyll-a (mg C (mg Chl-a)$^{-1}$ h^{-1})	0.5	0.5–6	(3)	Behrenfeld and Falkowski (1997)
	$K_{s,P}$	Half saturation coefficient of total dissolved phosphorus for the primary production (g m^{-3})	0.09	0.006–0.189	(3)	Maavara *et al.*, (2017)
Mineralization and burial of organic carbon	k_{20}	Mineralization rate at a reference temperature of 20 °C (d^{-1})	0.008	0.003–0.015	(7)	Maavara *et al.*, (2017)
	θ	Temperature dependence of mineralization	1.02	1.01–1.07	(7)	Maavara *et al.*, (2017)
	k_{bur}	Carbon burial rate in the lake (d^{-1})	0.004	$1/2k_{20}$	(8)	Mendonca *et al.*, (2017)
	f_{mm}	Fraction of mineralization that channels to the methanogenesis pathway	1/4	1/6–1/2	(10) (26)	Hanson *et al.* (2014); Bastviken (2009)
CH_4 oxidation	k_{max}	Maximal rate of CH_4 oxidation (g CH_4 m^{-3} d^{-1})	0.69	0.19–7.68	(18)	Liikanen et al. (2002)
	$Q_{10,ox}$	Temperature dependence of CH_4 oxidation (-)	1.2	1.1–2.0	(18)	Liikanen et al. (2002)
Base value of the shape parameter	α_{min}	Exponentially decreasing rate of CH_4 production versus sediment depth (m^{-1})	10	10–70	(12)	Langenegger *et al.*, (2019)

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3. Results

3.1 Assessing the performance of FLaMe in capturing patterns of CH_4 dynamics across different lake types

The FLaMe model is shown to be able to well capture the typically observed, contrasting physical and biogeochemical behaviors for two representative cases (Fig. 5 and Fig. S9–17; more details in Supplementary Text S3): shallow vs. deep, eutrophic vs. oligotrophic and warm vs. cold lakes. In the deep oligotrophic lake, the mean temperature reveals a lower and narrower seasonal variability (~3−8° C) compared to the shallow eutrophic lake (5−15°C) (Fig. 5a vs. 5b). Large temperature variations in the latter are mainly driven by the smaller water volume and thus faster mean temperature response to fluctuations in atmospheric temperature. In addition, the annual averaged F_{PP} in the shallow eutrophic lake (490 gC m^{-2} yr^{-1}) is approximately 38 times higher than that in the deep oligotrophic lake (13 gC m^{-2} yr^{-1}) (Fig. 5c vs. 5d). This difference can be explained by the differences in water volume (energy exchange), trophic status, and climate forcings. The higher F_{PP} of the shallow eutrophic lake also translates into higher $C_{OC,auto}$ concentration (~110 times) which persist over longer periods (Fig. 5e vs. 5f), despite substantially higher F_{min} rates.

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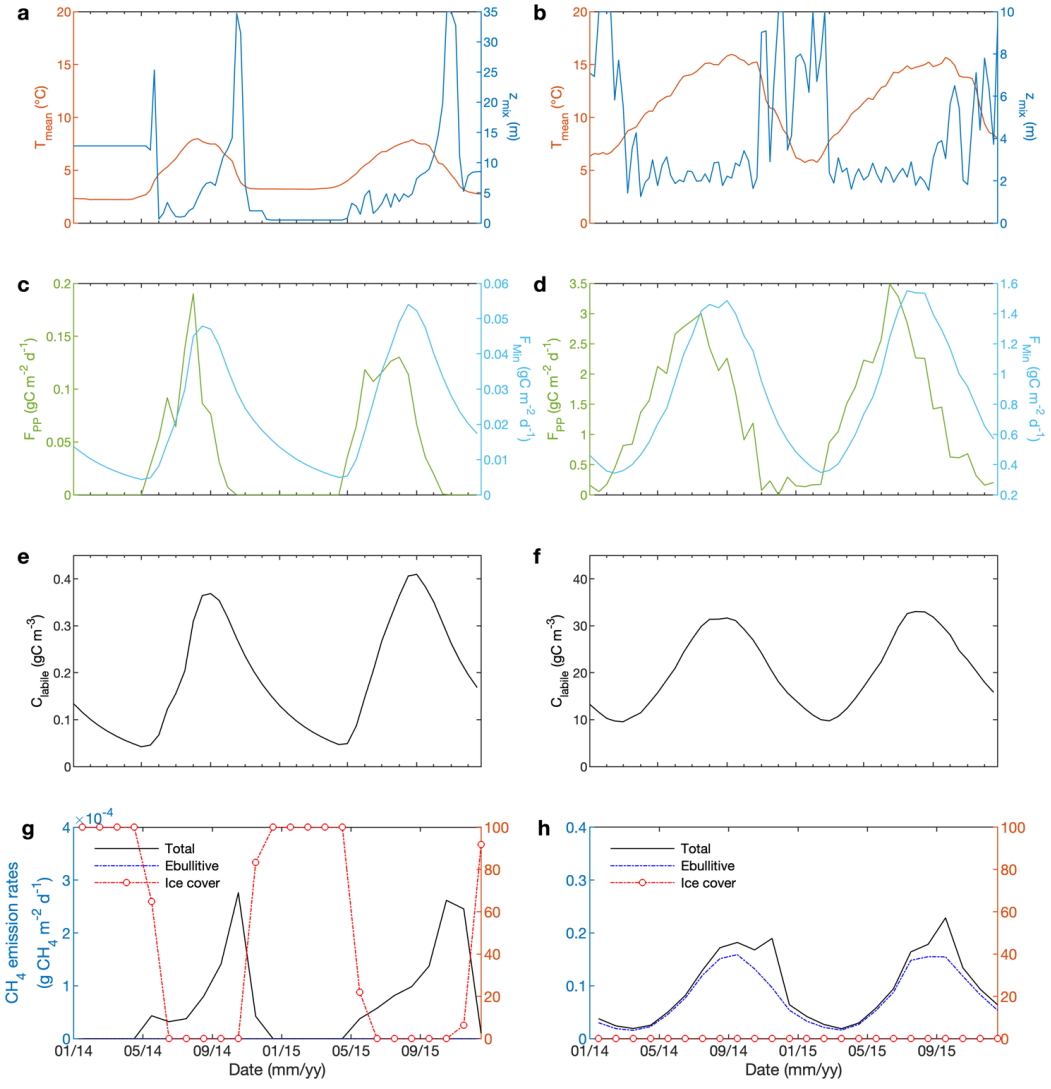

Fig. 5. Depth-integrated temporal evolution of variables and processes in two theoretical representative lakes. The deep oligotrophic lake (left) has a maximal depth of 35 m and [TDP] of 3 µg*P* L^{-1}, and is driven by the climate forcings at the location of 63.75°*N*, 26.25°*E*. The shallow eutrophic lake (right) has a maximal depth of 10 m and [TDP] of 80 µg*P* L^{-1}, and is driven by the climate forcings at the location 43.75°*N*, -6.25°*E*. (a) and (b) show the evolution of lake mean temperature and mixing depth; (c) and (d) show the evolution of primary production (F_{PP}) and mineralization rate (F_{Min}); (e) and (f) show the evolution of concentration of

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autochthonous organic carbon ($C_{OC,auto}$); (g) and (h) show the evolution of CH_4 emission rates and ice cover. Note the difference scales between the left and right panels.

In the deep oligotrophic lake, the simulated vertical temperature profiles indicate an almost permanently maintained stratification that is only interrupted by short but intense turnover events during late falls (Fig. S9a). Lake stratification (e.g., lake turnover and O_2 concentrations that depend mostly on solubility and hence, temperature) dominates the spatio-temporal pattern of O_2 such that O_2 concentration is near-saturated during most of the year (Fig. S9c). The oligotrophic status, together with the well oxygenated conditions, results in extremely low CH_4 concentrations. Higher CH_4 concentrations are only simulated near the lake bottom following the productive season, i.e., late summer/fall transition (Fig. S9e). In contrast, in the shallow eutrophic lake, the weaker stratification results in a less pronounced vertical temperature gradient (Fig. S9b). The vertical lake O_2 profile is not only controlled by the lake physics (temperature and O_2 solubility) but also by intense biogeochemical processes (Fig. S9d). During summer, O_2 concentrations in the upper portion of the lake are slightly supersaturated due to photosynthetic activity, followed by a gradual decrease in O_2 concentration as mineralization rates exceed primary production rates. Due to the high primary production in the eutrophic lake, large amounts of OC are exported below the thermocline, where heterotrophic activity progressively depletes O_2, leading to the development of anoxic conditions in the hypolimnion. The combination of high F_{Min} and low O_2 concentrations drive the accumulation of CH_4 in late summer at the bottom of the lake (Fig. S9f), with maximal CH_4 concentration (3.0 g CH_4 m^{-3}) exceeding those simulated in the deep oligotrophic lake by a factor of 600 (Fig. S9e).

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By aggregating CH_4 fluxes over time, we obtained distinct seasonal patterns of CH_4 production and emission for these two representative lakes (Fig. 5g and 5h; Fig. S10). In the cold, deep oligotrophic lake (Fig. 5g and Fig. S10a), winter to early spring ice cover (December–April) blocks CH_4 emissions such that lake CH_4 emissions are limited to the period between May and November. CH_4 production is highest (8.0×10^{-4} g CH_4 m^{-2} d^{-1}) in August and lowest (8.0×10^{-5} g CH_4 m^{-2} d^{-1}) in April. Almost all the produced CH_4 escapes the sediment via diffusion as gas bubbles do not form due to low CH_4 production rates and high-water pressure. However, the benthic CH_4 flux is subsequently largely oxidized in the well oxygenated deep water column. As a result, total lake CH_4 emissions are low (0 to 2.4×10^{-4} g CH_4 m^{-2} d^{-1}) with a slight peak in October. In the shallow eutrophic lake (Fig. 5h and Fig. S10b), the warmer climate prevents ice formation on the lake surface, leading to an emission season about twice as long as under colder climatic conditions. CH_4 production (0.02 to 0.35 g CH_4 m^{-2} d^{-1}) is >1000 times higher than that in cold, deep oligotrophic lake due to the higher nutrient loads, lower O_2 levels, higher irradiance as well as higher temperature (Fig. 5b). Higher CH_4 production rates, together with lower water pressure, drive the formation of gas bubbles, leading to a higher fraction of CH_4 emissions via the ebullitive pathway. The weaker stratification and the shorter transport time scale in the shallow lake limits CH_4 oxidation during diffusive transport, leading to ~900 times higher total CH_4 emissions compared to the deep, oligotrophic lake. Total lake CH_4 emissions are highest (0.21 g CH_4 m^{-2} d^{-1}) in September and lowest (0.02 g CH_4 m^{-2} d^{-1}) in February.

By decomposing the collective responses of shallow and deep lakes into individual effects induced by trophic level, climate and lake depth using additional theoretical numerical simulations, we found that the trophic level exerts the most important control on CH_4 dynamics, followed by climate, and finally, lake depth (Fig. S11–S14). Specifically, the yearly mean CH_4 production is

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increased by a factor of 30 (from 0.003 to 0.089 g CH_4 m^{-2} d^{-1}), and the yearly mean CH_4 emission is increased by a factor of 44 (from 0.0013 to 0.057 g CH_4 m^{-2} d^{-1}) from oligotrophic to trophic status (i.e., [TDP] increased by 10 times) (Fig. S12). From cold to warm climate, the yearly mean CH_4 production and emission increase by a factor of 6 (0.0094 to 0.059 g CH_4 m^{-2} d^{-1}) (Fig. S13), and a factor of 5 (0.0057 to 0.03 g CH_4 m^{-2} d^{-1}), respectively. By increasing lake depth from 15 m to 35 m (Fig. S14), the CH_4 production rates remain almost the same, i.e., 0.02 g CH_4 m^{-2} d^{-1} for the yearly mean and 0.06 g CH_4 m^{-2} d^{-1} for the peak, while the CH_4 emissions are overall lower for the deeper lake.

3.2 Evaluation of simulated temporal lake CH_4 emissions against observations from four well-surveyed lakes

In Klöntal and Erssjön Lakes (Table 2, Fig. 6a and 6b), FLaMe captures the observed seasonal cycles of CH_4 emissions well, albeit with almost a one-month delay. As a result, the simulated CH_4 fluxes are slightly lower in the first half of the year and slightly higher in the second half. This lag between observations and model results is likely due to the use of idealized climate forcings but could also be attributed to the unresolved changes in water levels and in-lake TDP dynamics. In the Klöntal Lake (Fig. 6a), the observed CH_4 fluxes are exceptionally high in April (1.64 g CH_4 m^{-2} d^{-1}) and July (5.03 g CH_4 m^{-2} d^{-1}), interrupting the normal seasonal cycles. These abrupt observed emissions might reflect the contributions from storage fluxes that are not well captured by FLaMe. Apart from these two months with exceptionally high fluxes, the observational data indicates peak emissions of 3.18 g CH_4 m^{-2} d^{-1} in August and no emissions during the ice-covered period. FLaMe simulates similar fluxes, with a peak of 3.4 g CH_4 m^{-2} d^{-1} in September (and 3.17 g CH_4 m^{-2} d^{-1} in August), and a null flux in January–February when the model predicts ice formation. In the Erssjön Lake (Fig. 6b),

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observational data report a peak in CH_4 emission reaching 13.48 g CH_4 m^{-2} d^{-1} in July and no emissions during the ice-covered period, whereas FLaMe simulates a peak emission of 18.76 g CH_4 m^{-2} d^{-1} in August (and 12.82 g CH_4 m^{-2} d^{-1} in July), and no flux in February. Moreover, the simulated CH_4 fluxes are exceptionally high in April (11.10 g CH_4 m^{-2} d^{-1}) due to the release of a storage fluxes that does not seem to be recorded by the observations. These high CH_4 fluxes attributed to storage and lake turnover are usually associated with large variability, i.e., in Klöntal Lake (Fig. 6a), the observed variability (standard deviation, SD) in CH_4 flux in July is almost 8-fold larger than the simulated one, whereas in Erssjön Lake (Fig. 6b), the simulated SD in CH_4 flux in April is almost 6-fold larger than that of the observed one. This suggests that both *in-situ* measurements and FLaMe struggle to accurately capture the storage fluxes. Apart from these storage fluxes, we found that the SDs of CH_4 fluxes simulated by FLaMe are lower than those observed for most months, indicating a more stable behavior in the simulations compared to the observations across the multi-year period considered here.

For the Upper Mystic and Villasjön Lakes (Fig. 6c and 6d), the observed temporal patterns of CH_4 fluxes appear more erratic, either due to the dominant role of short-term water level fluctuations or due to the complex ice cover dynamics. For the Upper Mystic Lake (Fig. 6c), the observed CH_4 fluxes are irregular or fluctuating (0–17.6 g CH_4 m^{-2} d^{-1}) over the year, a pattern which was explained by dynamic variations of lake water levels (Varadharajan, 2009). Since *in-situ* water level measurements are lacking, the simulated temporal variations cannot capture these observed erratic patterns well. Our model produces a smoother seasonal cycle of monthly-mean CH_4 fluxes over the year, i.e., high fluxes (10.02–13.38 g CH_4 m^{-2} d^{-1}) during the productive season (August–October), and low fluxes (0.02–7.56 g CH_4 m^{-2} d^{-1}) during the other months. Moreover, the model predicts a

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733 weak storage flux occurring in November (10.20 g CH_4 m^{-2} d^{-1}). For the Villasjön Lake (Fig. 6d), the
734 observed CH_4 fluxes are limited to the period of June–October, due to the long ice cover period
735 induced by the cold climate. FLaMe captures the observed ice-cover period well and produces similar
736 seasonal cycles of CH_4 fluxes. The simulated means and SDs are very close to observations in June
737 and July, but both, means and SDs, are much lower than observations in August, September, and
738 October.

739 In summary, despite the use of idealized climatic forcing and neglecting variations in lake area
740 and water level, FLaMe broadly captures the observed temporal patterns of monthly mean emissions,
741 albeit sometimes with small delays or diverging extents of high emissions periods. The SDs of
742 simulated CH_4 fluxes are also usually lower than the observed values, which is to be expected
743 considering that our model is not designed to capture high-frequency fluctuations of CH_4 fluxes. The
744 largest biases can be found in the estimations of storage fluxes (timing and magnitude), probably due
745 to 1) the difficulty of capturing these fluxes with existing measurement instruments and techniques,
746 2) the possibility of methane oxidation with greater than expected values during turnover and ice-out
747 (Mayr *et al.*, 2020; Zimmermann *et al.*, 2019; Pajala *et al.*, 2022) and 3) the lack of *in-situ*
748 measurements of climate conditions, dynamical water levels, and dynamic TDP concentrations
749 (Denfeld *et al.*, 2018). Resolving these issues will require to assemble a much larger dataset of
750 observed long time-series of CH_4 fluxes and associated physical and biogeochemical variables. To
751 help further calibrate and evaluate the model, this much larger pool of observations should span a
752 broader range of environmental conditions to be more representative of the lake CH_4 dynamics on
753 the continental to global scales. Overall, given the scarce spatiotemporal observations and the limited
754 possibility to validate current knowledge on process regulation in fields, it is difficult for all existing

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755 models to produce the details of the CH_4 dynamics in specific single lakes. Hence, the temporal
756 patterns of CH_4 fluxes simulated by FLaMe v1.0 are seen as acceptable, as its main focus is to capture
757 the broad spatio-temporal patterns of CH_4 emissions across the thousands of lakes that need to be
758 accounted for in large-scale applications (see section 3.3).

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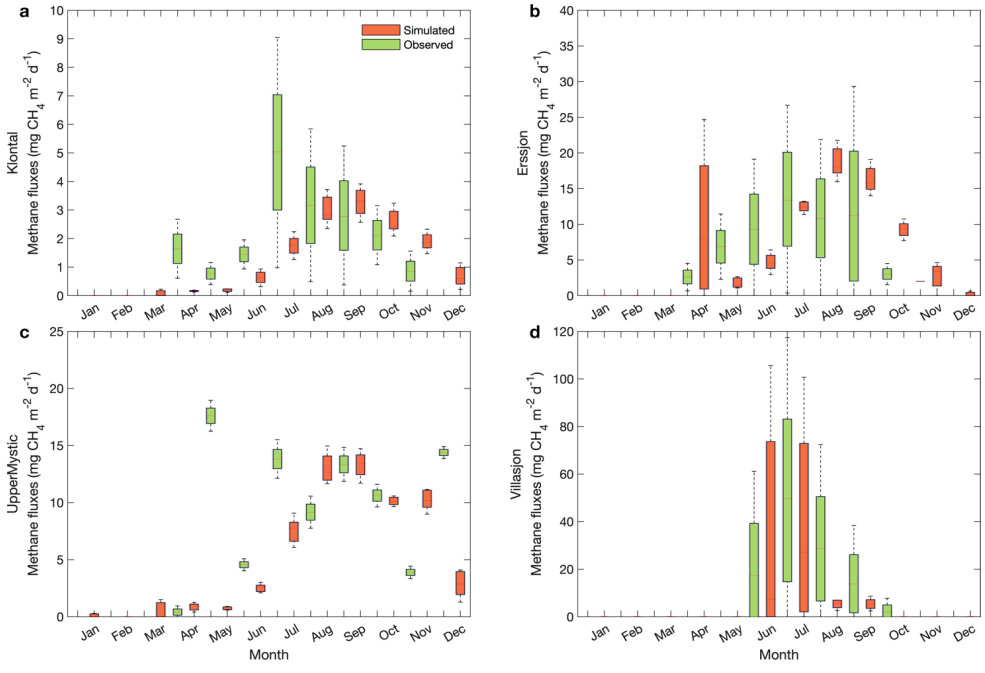

760

761 **Fig. 6. Evaluation of FLaMe against monthly mean CH_4 fluxes recorded in long time-series of observations**
762 **in four real lakes. (a) Klöntal, (b) Erssjön, (c) Upper Mystic, and (d) Villasjön. The detailed lake**
763 **characteristics are listed in Table 2. The climate forcings for these four lakes are extracted from GSWP3-**
764 **W5E5 model from ISIMIP3a. Note the different scales of CH_4 emissions in each lake.**

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3.3 FLaMe application on the European domain

3.3.1 Evaluation of FLaMe in European lakes

In the European scale application of FLaMe, we first evaluated the simulated F_{PP} against the empirical ranges reported by Wetzell (2001) for lakes under ultraoligotrophic (0–5 µgP L^{-1}), oligotrophic (5–10 µgP L^{-1}), mesotrophic (10–30 µgP L^{-1}), and eutrophic (>30 µgP L^{-1}) conditions (Fig. 7 and Fig. S18). Figure 7 shows that, under different trophic status, the means and medians of F_{PP} simulated by FLaMe (for 953 representative lakes) fall well within the reported ranges. Slight deviations could only be observed in ultraoligotrophic lake for which the model tends to slightly overestimate F_{PP} (Fig. 7a). Ultraoligotrophic and oligotrophic lakes reveal very similar mean and median of F_{PP} that fall at the higher ends of the ranges specified by Wetzel (2001) or even exceed it in the case of ultraoligotrophic lakes. In turn, mesotrophic and eutrophic lakes reveal mean and median F_{PP} that fall at the lower ends of the ranges specified by Wetzel (2001). This slight difference of simulated versus observed F_{PP} in lakes with different trophic conditions can be explained by the relatively low value of $K_{s,P}$ (90 µg L^{-1}) compared to the concentration of [TDP] (Fig. S7–S8), as well as the simplified representation of lake primary production in our model. When extending the representative lakes to all real lakes in the European domain (n=108407), the median and mean of simulated F_{PP} are still within the specified ranges but are reduced slightly for all trophic status (Fig. S18), attributed to the positively skewed distribution of [TDP] (Fig. S8), i.e., many lakes have a low [TDP].

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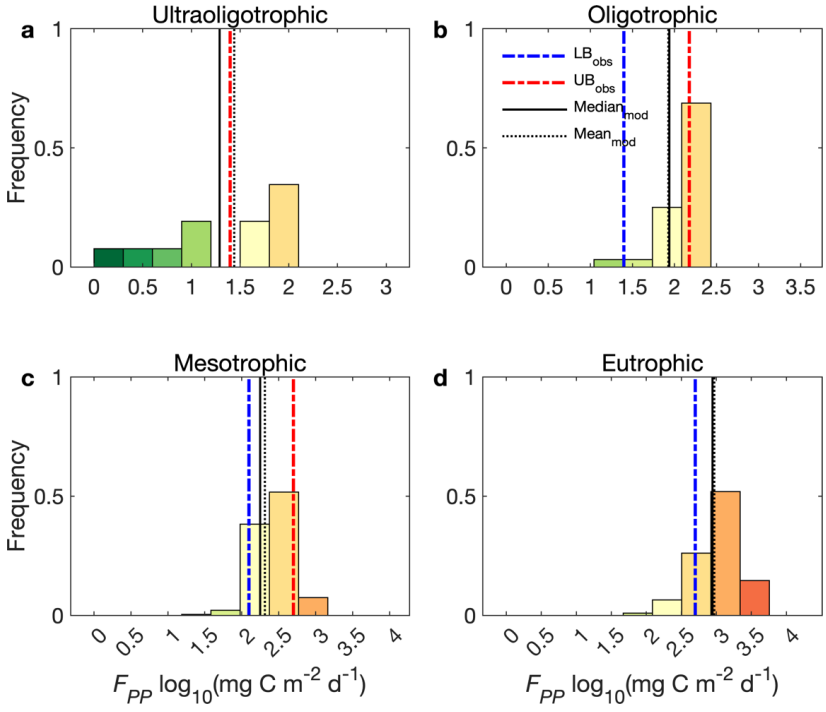

Fig. 7. Comparison of simulated primary production (F_{PP}) with empirical estimates reported by Wetzel (2001). The histograms show the frequency distributions of simulated F_{PP} (log scale) for 953 representative lakes that are grouped into ultraoligotrophic (0–5 µgP L^{-1}), oligotrophic (5–10 µgP L^{-1}), mesotrophic (10–30 µgP L^{-1}), and eutrophic (>30 µgP L^{-1}) lakes. In the figure, blue and red dashed lines are the lower and upper bounds (LB_{obs} and UB_{obs}), respectively, of empirical ranges reported by Wetzel (2001) in this class of lakes; Black solid and dotted lines are the $median_{mod}$ and $mean_{mod}$, respectively, of simulated F_{PP} for this class of lakes.

Next, we evaluated the simulated diffusive and ebullitive CH_4 emission rates against measurements in boreal and central European regions during late summer (August–September, 2010–2011) synthesized by Rinta *et al.* (2017) (Fig. 8 and Fig. S19). As Rinta *et al.* (2017) compiled *in-situ*

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796 measurements of diffusive and ebullitive CH_4 emission rates from 17 boreal lakes (in southern
797 Finland and Sweden) and 30 lakes of central European lakes (in The Netherlands, Germany and
798 Switzerland), we extracted the mean CH_4 emission rates during August–September for representative
799 lakes located in the grid cells corresponding to these two regions. Results indicate that the simulated
800 diffusive CH_4 emissions for boreal European lakes (Fig. 8) agree well with the observations; yet the
801 simulated ebullitive CH_4 emissions are slightly higher than the observations, leading to slightly higher
802 total emissions. For central European lakes, the simulated diffusive CH_4 emissions are slightly lower
803 than the observations, while the simulated ebullitive CH_4 emissions are slightly higher, leading to a
804 good agreement in the total emissions (Fig. 8). Moreover, Rinta *et al.* (2017) reported 6 and 27 times
805 higher diffusive and ebullitive fluxes in central Europe, respectively, while our model simulates a
806 smaller contrast of a 3- and 7-fold difference. This smaller contrast in the simulation can likely be
807 explained by the higher variability in measurements, reflecting diverse climate, light and catchment
808 properties in real lakes, while the variabilities in the simulated fluxes are significantly lower, probably
809 due to more homogeneous representations of environmental conditions in the FLaMe simulations.
810 Specifically, the large differences in measured CH_4 emissions in boreal and central European lakes
811 are attributed to their distinct characteristics, including climate (colder and dryer in the boreal region),
812 light regime (larger absorbance in the boreal region) and catchment properties, in particular land-use
813 (dominance of forests and smaller fraction of managed agricultural land in the boreal region).
814 However, in FLaMe, the catchment properties are not fully captured by our sole, simplified indicator
815 of [TDP], such that the differences between boreal and central European lakes are underestimated.
816 The coarse resolution of our model also likely reduces the represented range of climate conditions in
817 our simulations compared to those experienced by the sampled lakes. In the meantime, observations

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are also associated with uncertainties, because measurements were not continuous in time and might thus not be fully representative of the late summer-early fall period, as well as sampling and measuring CH_4 emissions, in particular via the ebullitive pathway, is all but a trivial task. Nevertheless, the above evaluation of FLaMe against observations overall reveals the ability of our model to reproduce broadly observed patterns in primary production and CH_4 emissions observed across distinct trophic status and landscapes.

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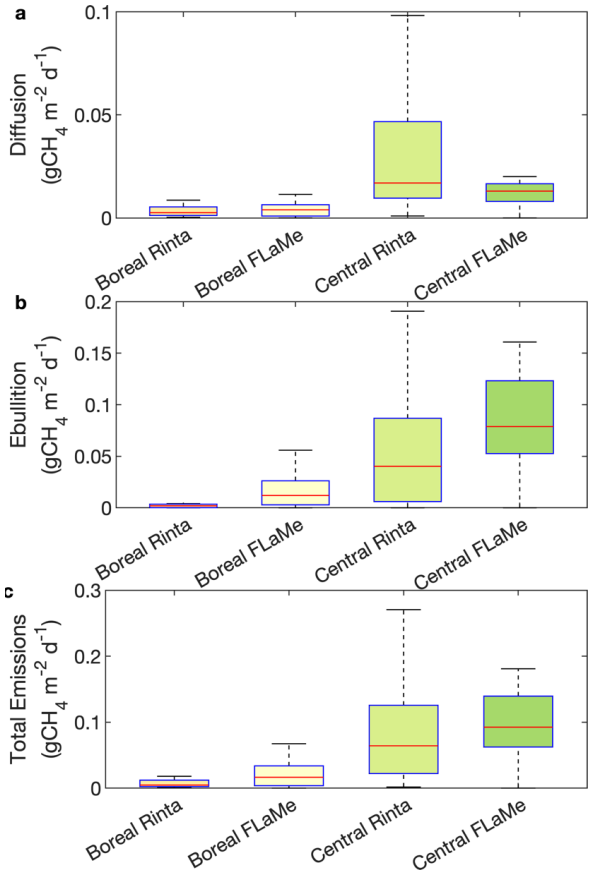

Fig. 8. Comparison of simulated diffusive (top), ebullitive (middle) and total (bottom) CH_4 emission rates with the measurements complied by Rinta *et al.* (2017). The datasets reported by Rinta *et al.* (2017) comprises the diffusive, ebullitive and total emission rates from 17 boreal lakes in Finland and Sweden and 30 lakes of central European lakes in The Netherlands, Germany and Switzerland. The boxes represent the 25% and 75% quartiles, and the whiskers cover the 95% confidence intervals. The same figure with a log scale is presented in Fig. S19.

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3.3.2 European scale assessment of lake CH_4 emissions

The continental-scale assessment indicates that European lakes smaller than 1000 km^2 have an annual mean emission of 0.97 Tg CH_4 yr^{-1} from autochthonous phytoplankton production during the period of 2010–2016, of which 30% and 70% are through diffusive and ebullitive transport pathways, respectively (Fig. 9 and Fig. S20). The mean CH_4 emission rates per unit lake area amounts to 7.39 g CH_4 m^{-2} yr^{-1}, while the mean CH_4 emission rates per unit land surface area amounts to 0.054 g CH_4 m^{-2} yr^{-1}. Both emission rates decrease from South to North, despite the larger number of lakes and lake surface area in Northern Europe (Messager *et al.*, 2016; Fig. S4). This south to north decrease can be explained by a much higher CH_4 emission rate in the South of Europe (reaching 109.6 g CH_4 m^{-2} yr^{-1}) driven by much higher eutrophic status of southern lakes (together with higher temperatures), which outcompetes the effect of the larger lake area in the Scandinavian region and Finland (which contribute to ~30% of the European lake area). This latitudinal pattern of CH_4 emissions per unit lake area is broadly consistent with that reported by Johnson *et al.* (2022) based on observations.

In terms of seasonal variability, our model results are in full agreement with the sparse data set of seasonally resolved observations (Tan *et al.*, 2015) and show that European lakes as a whole act as a continuous CH_4 source including during the winter months. Moreover, the simulated CH_4 production and emission reveal a sharp 10-fold increase from late Spring to late Summer that is largely driven by the increase in ambient temperature and F_{PP} rates. These findings underscore the importance of accounting for seasonal variations in CH_4 emissions when refining regional methane budgets (Tan *et al.*, 2015; Guo *et al.*, 2020; Johnson *et al.*, 2022; Stavert et al., 2022). A simple extrapolation of observed summer emissions to the yearly timescale would thus lead to an overestimation of yearly mean fluxes. In addition, model results also reveal a slight time-lag between

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the most favorable climate conditions (air temperature and light) and the maximum CH_4 production. This time lag in the model can be explained by the cascade of biogeochemical reactions (primary production, mineralization, O_2 depletion and onset of CH_4 production) that ultimately control benthic CH_4 fluxes, and the timescale of heat transfer from the lake surface to the deepest portion of our valley-shape lake bottom. This slight time-lag is further amplified by the time required for the benthic CH_4 to reach the water-air interface, although this effect is secondary due to the dominance of shallow lakes (with mean depth <10 m for 98.7% of lakes) within the European domain. Finally, the broad seasonal pattern in CH_4 emissions is complicated by the episodic releases of storage fluxes during lake turnovers which occur during spring (March and April; emissions>production) and fall (October and November; emission circa 85% of the production). Lake turnovers amplify total emissions for the duration of these short-lived events.

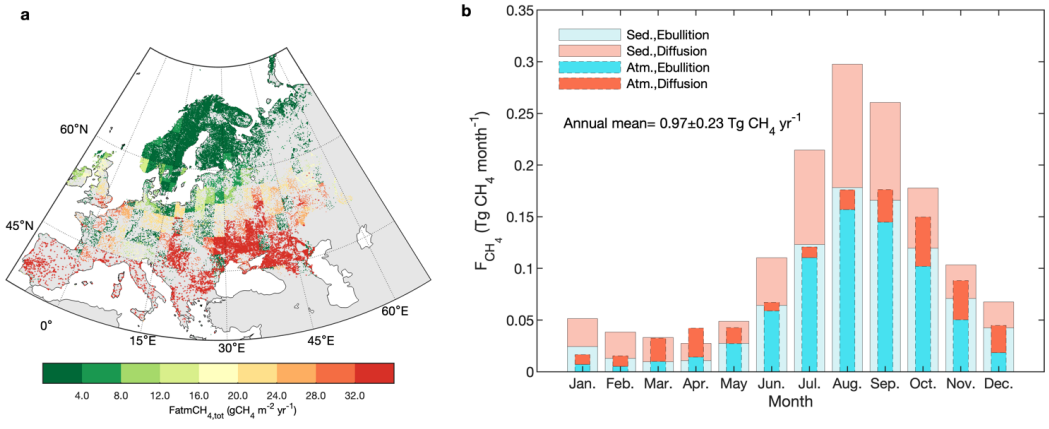

Fig. 9. Methane (CH_4) emissions from European lakes. (a) Spatial distribution of annual mean total CH_4 emissions (sum of diffusion and ebullition) for the period of 2010-2016, expressed in per unit of lake area. (b) Seasonality of total CH_4 production (dashed colors) and emission (full colors) fluxes and their split between ebullitive and diffusive pathways (period 2010-2016).

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3.3.3 Sensitivity and uncertainty analysis

The sensitivity analysis of annual mean CH_4 emissions from European lakes to key model parameters (listed in Table 3) are summarized in Table 4. Table 4 indicates that the fraction of benthic organic matter mineralization channeled to methanogenesis (f_{mm}) is the most sensitive parameter, and the increase (decrease) of f_{mm} by one SD leads to an increase (decrease) of European lake CH_4 emissions by 0.92 Tg CH_4 yr^{-1} or 95% (0.67 Tg CH_4 yr^{-1} or 69%). This is intuitive as a higher fraction of carbon channeled to methanogenesis will increase the continental scale CH_4 emissions, although the response is nonlinear. The second and third most sensitive parameters are the maximum carbon fixation rate per unit of Chlorophyll-a ($P_{chl,max}$) and the half saturation constant of phosphorus ($K_{s,P}$). An increase (decrease) of $P_{chl,max}$ by one SD could increase (decrease) the European lake CH_4 emissions by 0.77 Tg CH_4 yr^{-1} or 79% (0.63 Tg CH_4 yr^{-1} or 65%). This is again logical as a higher $P_{chl,max}$ indicates a stronger capacity of phytoplankton to assimilate carbon, thus resulting in higher amounts of organic carbon available for CH_4 production and emissions. The increase (decrease) of $K_{s,P}$ by one SD decreases (increases) the European lake CH_4 emissions by 0.46 Tg CH_4 yr^{-1} or 48% (0.22 Tg CH_4 yr^{-1} or 22%), a result which can be explained by a stronger TDP limitation of primary production when $K_{s,P}$ increases, resulting in lower CH_4 production and emissions. The next most sensitive parameters are the mineralization and burial rates (k_{20} and k_{bur}), for which an increase (decrease) in k_{20} by one SD result in an increase (decrease) of European lake CH_4 emissions by 0.19 Tg CH_4 yr^{-1} or 20% (0.39 Tg CH_4 yr^{-1} or 40%), while an increase (decrease) of k_{bur} by one SD leads to a decrease (increase) of European lake CH_4 emissions by 0.35 Tg CH_4 yr^{-1} or 36% (0.21 Tg CH_4 yr^{-1} or 22%). This is straightforward to interpret as a higher mineralization rate (k_{20}) will channel more mineralization into methanogenesis (and also via lower O_2 levels in the lake), while a higher

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burial rate (k_{bur}) translates to a lower relative amount of organic matter degradation, and thus lower CH_4 production and emissions.

The other parameters (including the shape parameter of the CH_4 production rate versus sediment depth α_{min}, the temperature dependence of mineralization θ, as well as the maximum CH_4 oxidation rate k_{max} and its temperature dependence $Q_{10,ox}$) are less sensitive, with their relative effects on European lake CH_4 emissions ranging from 1–20%. The shape parameter α_{min} can affect the CH_4 emissions as it determines the split between diffusive and ebullitive pathways, i.e., a higher α_{min} favors a higher fraction of CH_4 emitted to water and atmosphere through the diffusive pathway, a pathway that is more prone to oxidation thus lowering total CH_4 emissions. We also find that a higher temperature dependence of mineralization (θ) results in a lower CH_4 emission. This can be explained by the reference temperature of 20°C in the expression of the θ function, higher than the mean water temperature in most lakes, leading to a faster drop in mineralization for a larger θ when temperature is lower than 20°C. The parameter k_{max} barely impacts the total CH_4 emissions, as this parameter mostly influences the thickness of the water layers where the profiles of oxygen and methane overlap and the oxidation occurs, while the volume-integrated rates remain essentially unaltered (Regnier *et al.*, 2011; Thullner and Regnier, 2019). As for the temperature dependence of oxidation ($Q_{10,ox}$), the sensitivity is even weaker because changing the $Q_{10,ox}$ value has a lower impact on the oxidation rates than changing k_{max}.

With the samples produced by the above sensitivity analysis and complemented by samples from additional tests, we utilized a Random Forest (RF) model to assess the uncertainty of European lake CH_4 emissions (see details in section 2.5.4). The RF model has a R^2 of 0.73 and Root of Mean Square

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Error (RMSE) of 0.24 Tg CH_4 yr^{-1} for the train set (Fig. 10a) and a R^2 of 0.52 and RMSE of 0.30 Tg CH_4 yr^{-1} for the out-of-bag samples (Fig. 10b), suggesting that the it can capture well the relationship between model parameters and European lake CH_4 emissions. Using these ensembles of CH_4 emissions, the uncertainty (or SD) of European lake CH_4 emissions associated with the choice of biogeochemical parameter values was estimated as 0.23 Tg CH_4 yr^{-1}. Therefore, during the period of 2010-2016, the European lakes have an annual mean emission of 0.97±0.23 Tg CH_4 yr^{-1}.

With the RF model, we can also identify the importance of key model parameters involved as predictors (Fig. 10c). We noticed that the first four leading parameters are also the most sensitive parameters as identified in Table 4, while the importance of other parameters are slightly different from the sensitivity analysis. This slight difference can be attributed to the interactions of model parameters that are overlooked in the sensitivity analysis. Overall, from the sensitivity and uncertainty analysis, we find that the European lake CH_4 emissions are strongly controlled by the carbon biogeochemical dynamics, which, however, was not fully accounted for in previous lake models.

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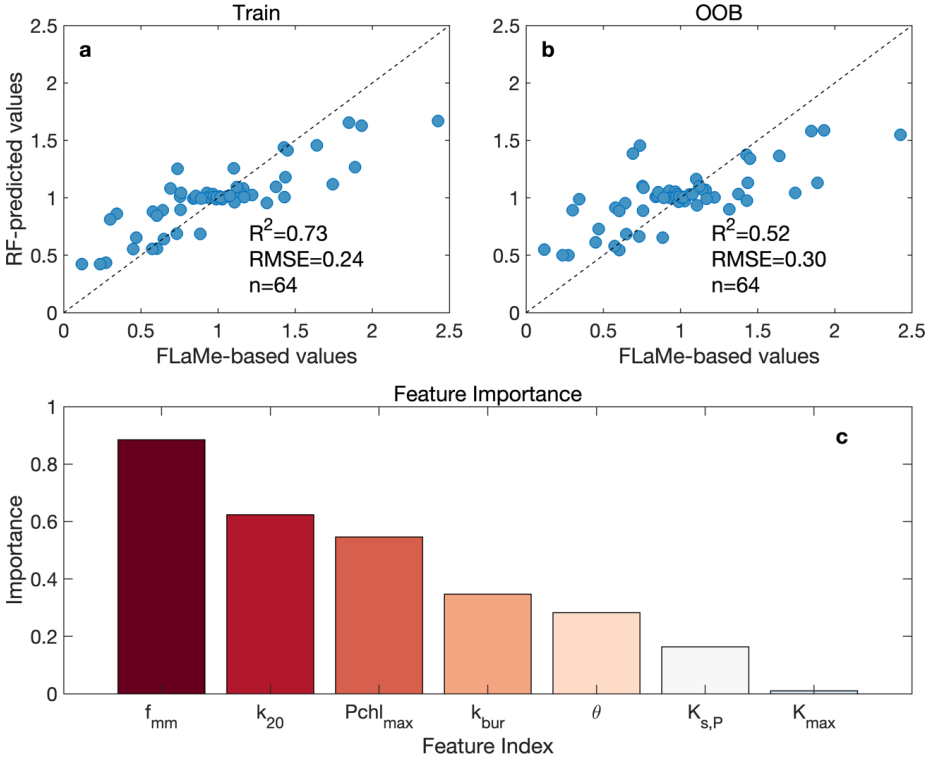
925 **Table 4 Sensitivity of European lake CH_4 emissions (Tg CH_4 yr^{-1}) to key model**
926 **parameters**

Parameter setting		Mean±SD		Mean±0.5SD	
		–SD	+SD	–0.5SD	+0.5SD
Primary production	P_{chl_max}	0.344	1.743	0.642	1.376
	$K_{s,P}$	1.432	0.754	1.170	0.852
Mineralization and burial rates	k_{20}	0.578	1.164	0.758	1.141
	k_{bur}	1.317	0.761	1.107	0.856
	θ	1.028	0.928	0.989	0.968
	f_{mm}	0.302	1.888	0.605	1.437
Methane oxidation	K_{max}	1.057	0.930	1.009	0.953
	$Q_{10,ox}$	0.992	0.983	0.978	0.973
Base value of the shape parameter	α_{min}	1.222	0.840	1.077	0.891



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928 **Fig. 10. Random Forest (RF) model for the uncertainty analysis. (a) and (b) are the train and test (Out-of-**

929 **Bag prediction) of the RF model. (c) shows the importance of key model parameters. Note that the**

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parameters of α_{min} and $Q_{10,ox}$ are excluded from illustration due to their second order of importance (indicated by negative values).

4. Model limitations

We have illustrated that FLaMe is able to capture complex physical-biogeochemical behaviors for lakes with diverse settings and environmental controls. Specifically, the FLaMe model has been evaluated against (i) observational temporal variations of CH_4 fluxes at four contrasting, well-surveyed real lakes, (ii) the empirical ranges of primary production under different trophic status reported by Wetzel (2001), and (iii) observational patterns of CH_4 emissions against trophic and climate gradients spanning the European domain (Rinta *et al.*, 2017). Moreover, the European scale simulation produces a spatial pattern of lake CH_4 emission rates consistent with observation-based upscaling approaches (Johnson *et al.*, 2022). This continental scale application also demonstrates the power of our modelling framework that rests on a lake clustering approach and on a routing of nutrient (TDP) inputs from surrounding catchments to lakes that allow to account for eutrophication effects. Our results thus suggest that the FLaMe modelling framework performs well in providing reliable spatio-temporal patterns of lake CH_4 emissions at the large scale. However, the results also pinpoint to several key aspects to be improved in the model and highlight critical data gaps that must be addressed in the future.

First, the organic carbon module only accounts for autochthonous OC production as the substrate for methanogenesis, but ignores the contribution of allochthonous OC inputs leached from the surrounding catchments, rivers and streamflow. This is based on the distinct reactivity of autochthonous and allochthonous OC inputs, with the latter being more refractory to microbial

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activities (mineralization and decomposition). As a result, FLaMe may provide conservative estimates of CH_4 productions and emissions. However, neglecting the allochthonous C inputs may at the same time minimize the feedback of OC on light penetration, leading to systematically biased estimates (section 2.2.2.1). Moreover, transient lake phosphorus dynamics and the co-limitations by nitrogen, albeit assumed to be less important, are neglected and might increase the uncertainty in the estimates of CH_4 production and emission. In addition, our primary production model does not resolve the short-term (e.g., (sub)daily) dynamics of algae growth induced by climate variability, rendering model-data comparison more difficult. In future model developments, these limitations could be addressed by (i) integrating or routing the lake water, carbon and nutrient fluxes along the global river network, which would allow to simultaneously solve the issue of time-invariant lake water levels in current global lake models (Golub *et al.*, 2022), including ours; (ii) refining the carbon module by incorporating more dynamic models for algal growth as well as P and N uptake and recycling processes within lakes.

Second, several model assumptions and implementations are based on empirical or theoretical knowledge, which may lead to biases in the estimation of CH_4 fluxes. For instance, the present version of FLaMe (i.e., v1.0) neglects the plant-mediated emission pathway (trough rooted plant) in the littoral zone due to the lack of observational data for model calibration. Moreover, in our model, the lake is assumed to follow a "valley" shape. Although this is an advancement from the "bucket" shape used in previous process-based lake models of CH_4 emissions (e.g., LAKE 2.0, ABLM, and bLake4Me), it remains a simplified assumption that captures important but not all features of a realistic lake geometry. Furthermore, several benthic CH_4 processes are highly parameterized. For instance, the split between aerobic and anaerobic decomposition of organic matter is represented by

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a single parameter f_{mm} and is determined based on the data compilation from Bastviken (2022). This simplification leads to the same temperature dependence of CH_4 processes occurring in the sediment as that of pelagic and benthic mineralization. This is a shortcoming although it should be noted that the overall temperature dependence of CH_4 emissions, which results from the combined effects of OC production, mineralization, and subsequent CH_4 processes, was found to fall well within the observed ranges reported by Aben *et al.* (2017) (Fig. S21). The split of diffusive and ebullitive CH_4 fluxes is also currently captured by an empirically determined threshold depth ($z_{eb,min}$) based on limited observations by Langenegger *et al.* (2019). Moreover, the effects of heat transfer and CH_4 bubbles migration in the sediment are not resolved, which may lead to biased simulation of CH_4 fluxes especially for the timing. These are simplified representations related to the highly complex pathways of CH_4 production and emission, which needs to be improved by more mechanistic representations of the biogeochemical processes controlling carbon cycling, CH_4 production and transport via diffusion and bubble ascent.

Third, different modules of the FLaMe model could benefit from more comprehensive calibration and evaluation but those are limited by data availability. Although FLaMe has been evaluated against several timeseries of observed data collected in four well-surveyed lakes with contrasted dynamics, a full evaluation in the context of large-scale application would benefit from a significantly larger and representative set of observational data. Moreover, the *in-situ* climate conditions may vary greatly from the grid-level forcings, and the lake water dynamics may also affect the CH_4 fluxes significantly (e.g., Upper Mystic Lake; Varadharajan, 2009). Thus, a full comprehensive set of *in-situ* measurements of climate, water level, physical and biogeochemical variables would be highly valuable for the purpose of further model development, calibration and evaluation. At the European

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scale, we partly circumvented these limitations by evaluating lake primary production against the broad ranges reported by Wetzel (2001), and the simulated diffusive and ebullitive CH_4 fluxes across the environmental (nutrient and climate) gradients compiled by Rinta *et al.* (2017). In this context, complementary time-series of vertically resolved organic carbon, CH_4 and O_2 concentrations, as well as high frequency measurements of CH_4 fluxes capturing short-lived emissions via the storage and ebullitive pathways and covering heterogeneity of CH_4 fluxes in large lakes (Denfeld *et al.*, 2018; Mayr *et al.*, 2020; Zimmermann *et al.*, 2019) would help further calibrate and evaluate the FLaMe model. These measurements should be performed using a sufficiently large set of representative lakes covering the full range of lake morphologies, landscape properties, and climate.

5. Conclusion and outlook

In this study, we developed and tested a new process-based biogeochemical modeling framework (FLaMe) to simulate lake CH_4 fluxes on the large-scale and, as a "proof of concept", applied the model to European lakes. The physical lake model builds on the Canadian Small Lake Model (CSLM) and is coupled to a set of novel biogeochemical modules describing lake organic matter, oxygen and methane dynamics. We then showcased the abilities and performance of FLaMe by: (1) analyzing the overall behaviors of the coupled C-O_2-CH_4 dynamics in two representative cases (a deep oligotrophic lake driven by cold climate in Northern Europe and a shallow eutrophic lake driven by warm climate in Southern Europe) as well as their decomposition, and (2) evaluating simulated temporal patterns of CH_4 fluxes against observations at four well-surveyed lakes with long-term timeseries. Simulation results were consistent with our common knowledge of lake CH_4 dynamics, suggesting that FLaMe can capture the patterns of CH_4 production and emissions across different lake types as well as their responses to the changes in environment conditions, despite the complexity of underlying

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biogeochemical processes. Furthermore, by applying the model to boreal and central European lakes, we showed that FLaMe captures well the observed magnitudes of both diffusive and ebullitive CH_4 fluxes as well as the difference between boreal and central lakes. Finally, at the European scale, FLaMe estimates total CH_4 emissions from lakes smaller than 1000 km^2 (n=108407, total area = $1.33x10^5$ km^2) as 0.97±0.23 Tg CH_4 yr^{-1}. In addition, the model resolves spatial patterns and seasonal variations of CH_4 emissions, providing a comprehensive view of their contribution to regional methane budgets.

Despite some limitations in its current model configuration, this first version of FLaMe is a significant step forward in biogeochemical simulations of lake CH_4 dynamics. The model explicitly incorporates the dynamics of depth-integrated organic carbon cycling, such that the responses of organic carbon to climate and environmental change can be accounted for in estimating CH_4 emissions. We also have incorporated the primary production as a function of total dissolved phosphorus loads from the surrounding catchments, allowing us to evaluate for the first time the impact of eutrophication on CH_4 emissions in a quantitative way. Moreover, our model is of intermediate complexity, and is thus designed for large scale applications. Although the model was run here at a coarse spatial resolution, its parallelized version offers the possibility to carry simulations at a finer resolution in the future. With these advancements, our model can be used to resolve the spatio-temporal variability of CH_4 emissions at regional and global scales under past and future climates, and has the potential to be coupled to Earth System Models to investigate the feedback between climate warming and global lake CH_4 emissions.

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Data availability

The methane emission data for the four well-surveyed real lakes (Klöntal, Erssjön, Upper Mystic, and Villasjön) were obtained from Tan *et al.* (2024). The *in-situ* measurements of diffusive and ebullitive CH_4 emission rates in boreal and central European regions during late summer (August–September 2010–2011) were obtained from Rinta *et al.* (2017). The lake characteristic information within Europe were obtained from the HydroLAKES database (Messager et al. 2016): https://www.hydrosheds.org/products/hydrolakes. The meteorological variables from GSWP3-W5E5 reanalysis product were obtained from Inter-Sectoral Impact Model Intercomparison Project (ISIMIP3a): https://www.isimip.org/gettingstarted/input-data-bias-adjustment/.

Code availability

The source codes for FLaMe (**F**luxes of **La**ke **Me**thane) model version 1.0 are available at: https://github.com/myFeng818/FLaMe-model-v1.0.git. The preprocessing and postprocessing codes for the model can be obtained upon request.

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Author contributions

M.M., M.F. and P.R. designed the study. M.M. and M.F. co-developed and tested the FLaMe model. D.B., S.A., R.L., A.J., G.G.L., M.D.M., and Z.T. provided plenty of valuable suggestions related to the model development. D.B. and S.A. also provided constructive suggestions on model evaluation against measurements and manuscript writing. M.D.M. helped us in setting up the CSLM at the beginning of developing FLaMe, Z.T. provided us the methane emission data from ISIMIP lake datasets, and W.T. helped us in collecting climate forcings from ISIMIP3a. M.M. and M.F. wrote the first version of the manuscript, and all coauthors helped in improving the manuscript.

Competing interests

The authors declare no competing interests.

Additional information

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