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Modeling seasonal salinity variations in a large West African lagoon (Nokoué, Benin): major drivers and mechanisms

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25 **Abstract**

26 Nokoué lagoon is a shallow water body connected to the Gulf of Guinea through the
27 long and narrow Cotonou channel. The salinity dynamics in the lagoon is investigated
28 using the 3D SYMPHONIE numerical model. We first validate the model using
29 salinity and water level data. Simulated and observed salinity and water level
30 variations compare well demonstrating that the reference model simulation correctly
31 reproduces the dynamics of the lagoon. By performing several simulations with
32 varying external forcings and freshwater fluxes, the main drivers of salinity variability
33 in the lagoon are identified.

34 We first focus on the salinization phase of the lagoon at the end of the tropical wet
35 season (between November and February), and we investigate the total change in
36 salinity and the salt fluxes involved in these variations. The high frequency salt fluxes
37 associated with the ocean tide import salt from the ocean via the Cotonou channel to
38 the south of the lagoon. Baroclinic fluxes, associated with the influence of salt on
39 density and associated pressure gradient, increase this local salinity input in the
40 south, but also play a major role in the dispersion of salt throughout the lagoon. Two
41 rivers provide a permanent freshwater inflow in the north/ northeast of the lagoon that
42 limits the salinization of the whole lagoon. Finally, wind-generated high frequency
43 recirculation, partially prevents the salinisation of the North-East of the lagoon.

44 During the desalinization period (between May and September), the lagoon salinity
45 variations are highly sensitive to the magnitude of river inflow. At a constant river flux
46 rate, the lagoon attains an equilibrium state in salinity. As this equilibrium is reached,
47 both the overall salinity level and the time needed to achieve it decrease notably with
48 higher fluxes. Beyond $100 \text{ m}^3 \text{ s}^{-1}$, only a small area near the Cotonou channel retains

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50 higher salinity, while surpassing $500 \text{ m}^3 \text{ s}^{-1}$ results in complete desalination of the
51 entire Nokoué lagoon.

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53 **Keywords:** Lagoon; Salinity variability; numerical model; physical processes.

54

55 **1. Introduction**

56 Coastal lagoons are highly valuable components of the natural heritage and
57 ecosystems in the coastal zone providing important ecosystem services such as
58 agricultural, tourism and fisheries development (Gonenc & Wolflin, 2004). They are
59 highly variable and productive environments (Kjerfve, 1994), and occupy ~13% of the
60 world's coastal zone, including 18% of the African coast (Cromwell, 1971; Barnes,
61 1980). Coastal lagoons are the dominant features along the West African coast
62 (Pauly, 1975) and cover an area of ~ 3,000 km² between Ivory Coast and Nigeria (A.
63 P. Lalèyè et al., 2007).

64 Nokoué Lagoon is one of the most important lagoons along the West African
65 coast in terms of biological productivity, ecological and economic values (Djihouessi
66 et al., 2017; A. P. Lalèyè et al., 2007). It is exploited by ~ 12,000 fishermen (Latifou et
67 al., 2020) and contributes to nearly 65% of Benin's continental fishery catches.
68 Nokoué lagoon also hosts the largest lacustrine villages in West Africa (Djihouessi &
69 Aina, 2018; Yehouenou et al., 2013) and more than one million people live in the
70 immediate vicinity of the lagoon (Gnohossou, 2006). Nokoué lagoon and the
71 surrounding wetlands host a large biodiversity and are classified as protected areas
72 by the Ramsar agreement (<https://rsis.ramsar.org/fr/ris/1018>). Overall, the lagoon
73 environment presents very important socio-economic benefits for the development of
74 Benin, and it is important to better understand its functioning to ensure its
75 preservation.

76 Salinity is one of the key factors affecting coastal lagoon ecosystems (Attrill,
77 2002; García-Oliva et al., 2019; Nunes et al., 2021), controlling the diversity and
78 abundance of fish, the distribution of plankton and the biomass of aquatic plants

79 (Dube et al., 2010; Franco et al., 2019; Nunes et al., 2021). Salinity depends on
80 ocean and river inputs into the lagoon and is very sensitive to physical drivers
81 influencing the hydrodynamics of the lagoon, such as tide, wind (Alekseenko et al.,
82 2017; Cerralbo et al., 2016; Mahanty et al., 2016) or the baroclinic pressure gradient,
83 due to density difference between saline and freshwater (e.g. Gross et al., 1999).

84 The Nokoué Lagoon is primarily fed by substantial freshwater inputs from the
85 Sô and Ouémé rivers in its northern region, along with a minor contribution from the
86 smaller Djonou river in the southwest (Fig. 1). It maintains a continuous exchange of
87 saltwater with the Atlantic Ocean through the Cotonou channel (Fig. 1), primarily
88 influenced by ocean tides (Okpeitcha et al. in 2022). Seasonal variations of
89 freshwater discharge into Nokoué lagoon increase its water level by ~1 m between
90 the dry and wet (African Monsoon) seasons (Chaigneau et al., 2022; Morel et al.,
91 2022), and the mean lagoon salinity increases from 0 in flood period to 25 in the
92 lowest water period (Okpeitcha et al., 2022). These variations in salinity are crucial
93 for the composition and structuring of the Nokoué lagoon ecosystems, from plankton
94 to fish (Chaigneau et al., 2023; P. Lalèyè et al., 2003; Odountan et al., 2019).

95 Generally, seasonal variations of lagoon salinity are first order driven by the
96 tide and river discharges. This is also documented for Nokoué lagoon (Djihouessi &
97 Aina, 2018; Laleye, 1995; Le Barbé et al., 1993; Okpeitcha et al., 2022; Zandagba et
98 al., 2016). In a recent study, based on in-situ observations and a simplified box
99 model, Okpeitcha et al. (2022) showed the high sensitivity of the lagoon's salinity to
100 flow variations and estimated that the lagoon could be fully desalinated for a river
101 flow rate greater than $\sim 50\text{-}60 \text{ m}^3 \text{ s}^{-1}$. In addition, they estimated that baroclinic effects
102 related to the salinity and pressure gradients may cause an additional average

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103 transport of approximately $40 \text{ m}^3 \text{ s}^{-1}$ of ocean water into the lagoon, in conjunction
104 with the advective exchange driven by tides.

105 In this study, we analyse the salinity variability in the Nokoué lagoon using a
106 realistic 3D numerical model. We first describe the numerical configuration and the
107 in-situ data available for the model validation (Section 2). The methodology, based
108 on a temporal coarse grain analysis of salt fluxes, is also presented (Section 2.4).
109 The reference simulation is presented and validated in Section 3.1. We then evaluate
110 the influence of key driving processes: tide, salt gradient, wind, river discharge
111 (Section 3.2). Finally, we investigate the role of freshwater fluxes and determine the
112 threshold over which ocean waters and salt cannot penetrate the lagoon.

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114 **2. Study area, model and datasets**

115 **2.1. Study area**

116 The Nokoué lagoon is a choked coastal lagoon spanning approximately 20 km
117 from West to East and 11 km from South to North (Gadel & Texier, 1986; Le Barbé et
118 al., 1993). This shallow lagoon has a mean depth of 1.3 m and 2.2 m during low and high-
119 water seasons, respectively (Chaigneau et al., 2022). It is the largest brackish water
120 body of Benin, covering an area $\sim 150\text{-}170 \text{ km}^2$ during dry and wet seasons and with
121 a volume of $1.5 \times 10^8 \text{ m}^3$ during low water periods (Le Barbé et al., 1993). The lagoon
122 is connected to the Atlantic Ocean via the Cotonou channel, which is approximately
123 4.5 km long and 400 m wide (Fig. 1c). The amplitude of semi-diurnal tide is strongly
124 attenuated within the lagoon compared to the ocean, with 55-90 cm in the ocean to
125 about 3-5 cm within the lagoon (Chaigneau et al., 2022).

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126 Nokoué lagoon receives freshwater from multiple sources, the main ones are
127 the Sô and Ouémé rivers in the northern part of the lagoon (Fig. 1c), which have
128 catchment areas of 50,000 km² and 1,000 km², respectively (Djihouessi & Aina, 2018;
129 Le Barbé et al., 1993). In addition, small amounts of fresh water enter the lagoon
130 from e.g. the Djonou river located southwest of the area and Totchè channel (Le
131 Barbé, 1993; Texier et al., 1980; Djihouessi & Aina, 2018). The seasonal cycle of the
132 regional climate is dominated by the West African monsoon (Fink et al., 2017) with
133 two dry and two rainy seasons (Ahokpossi, 2018; N'Tcha M'Po et al., 2017; Guedje
134 et al., 2019). The long dry season lasts from December to February, while the short
135 one is from August to September (Guedje et al., 2019). The main rainy season
136 occurs from March to July, with a peak in June, while the short rainy season runs
137 from September to November (Ahokpossi, 2018; N'Tcha M'Po et al., 2017). The
138 annual average rainfall is ~1300 mm (Biao, 2017; Lawin et al., 2019; Mama, 2010).
139 However, it is important to note that these double rainy seasons have no significant
140 effect on Nokoué lagoon water level (Chaigneau et al., 2022). Instead, the water level
141 is significantly influenced by the discharge of water from the Ouémé and Sô
142 catchments from the northern part of Benin. In this region to the north of the lagoon,
143 a single rainy season lasts from April to October with the highest rainfall occurring in
144 August (Ahokpossi, 2018; Biao, 2017; Lawin et al., 2019; N'Tcha M'Po et al., 2017).
145 During this period, the lagoon water level increases by ~1 m (Chaigneau et al., 2022)
146 and its salinity reduces to zero (Djihouessi & Aina, 2018; Okpeitcha et al., 2022).
147 Thus, the hydrological regime of the lagoon is primarily influenced by the seasonal
148 variations of river discharges. Unfortunately, there are no direct measurements of
149 river discharges into Nokoué lagoon, and estimates of peak fluxes during the rainy
150 season vary widely, ranging from ~400 m³ s⁻¹ to ~1,000 m³ s⁻¹ (Djihouessi et al., 2019;

151 Lawin et al., 2019; Le Barbé et al., 1993). However, a recent study by Morel et al.
152 (2022) proposed a method to estimate the net fluxes to the lagoon using
153 observations of water level variations and tide amplitude. Applying this method to
154 data acquired in 2018, they confirmed a maximum net flux of $\sim 1,100 \text{ m}^3 \text{ s}^{-1}$.

155 The prevailing winds in the study area are characterized by southwesterly
156 winds from the ocean throughout the year, except during the long dry season when
157 northeasterly winds called Harmattan can episodically occur from December to
158 March (Bajamgnigni Gbambie & Steyn, 2013; Guedje et al., 2019). The monthly
159 average wind speed varies between 2 and 5 $\text{m}\cdot\text{s}^{-1}$ with a main peak in August and a
160 secondary one in March (Guedje et al., 2019).

161 **2.2. SYMPHONIE model**

162 **2.2.1. Model description**

163 Numerical simulations were carried out using SYMPHONIE model (Marsaleix
164 et al., 2008), a three-dimensional (3D) free surface hydrodynamical model, solving
165 the Navier-Stokes equations using the Boussinesq and hydrostatic approximations.
166 The free surface variations at high frequency (mostly associated with surface gravity
167 waves) are taken into account using a time splitting method (Marsaleix et al., 2008).
168 The state variables (current, temperature, salinity, surface elevation) are calculated
169 on an Arakawa C-grid (Arakawa & Lamb, 1977) and a generalised sigma vertical
170 coordinates (terrain following coordinate), using an energy-conserving finite
171 difference method (Marsaleix et al., 2008, 2012, 2019). We have chosen a free slip
172 bottom boundary condition, for which the terrain following coordinate ensures a
173 perfect representation. This is of particular importance for such a shallow lagoon,
174 since spurious effects at the bottom (such as staircase representation for z Cartesian

175 coordinate) can strongly influence the dynamics over the whole water column. The
176 turbulence variables (turbulent vertical mixing coefficients, turbulent kinetic energy
177 and its dissipation rate) are calculated by the K-epsilon scheme as in Costa et al.,
178 (2017).

179 **2.2.2. Model set up**

180 For our study, the configuration used extends between longitudes 2.26°E and
181 2.76°E and between latitudes 6.11°N and 6.55°N (Fig. 1) with 563×646 horizontal
182 grid points. To correctly represent the dynamics in the Cotonou Channel and the
183 lagoon, a curvilinear bipolar grid is used with horizontal resolution varying from ~40 m
184 in the Cotonou Channel to ~250 m at the open ocean boundary (Fig. 1c). To correctly
185 resolve the hydrodynamic conditions of this shallow lagoon, 10 vertical levels were
186 used.

187 The bathymetry used merges over 123,000 depth soundings in the lagoon and rivers
188 (Chaigneau et al., 2022) complemented by GEBCO (GEBCO_08 Grid, version
189 20090202, <http://www.gebco.net>) in the coastal ocean.

190 The model is forced by large-scale ocean conditions, ocean tide, freshwater
191 inflow from rivers, wind, and atmospheric conditions.

192 Oceanic forcing (temperature, salinity, Sea Surface Height, currents) at the open-
193 ocean boundary (Marsaleix et al., 2006) are based on daily outputs from the
194 MERCATOR-Ocean (Lellouche et al., 2013) global configuration at 1/12° spatial
195 resolution (version PSY4QV3R1).

196 Tidal forcing is performed at the ocean boundary using the FES2014 solution
197 (Carrere et al., 2016), and 9 (M2, N2, S2, K2, K1, O1, P1, Q1, M4) main tidal
198 harmonic constituents are used.

199 Meteorological forcing fields were obtained from ECMWF three-hourly and $1/8^\circ$
200 spatial resolution forecasts (wind speed at 10m, temperature and specific humidity at
201 2m, surface pressure, precipitation, solar fluxes and thermal radiative fluxes) using
202 COARE bulk formulae (Fairall et al., 2003).

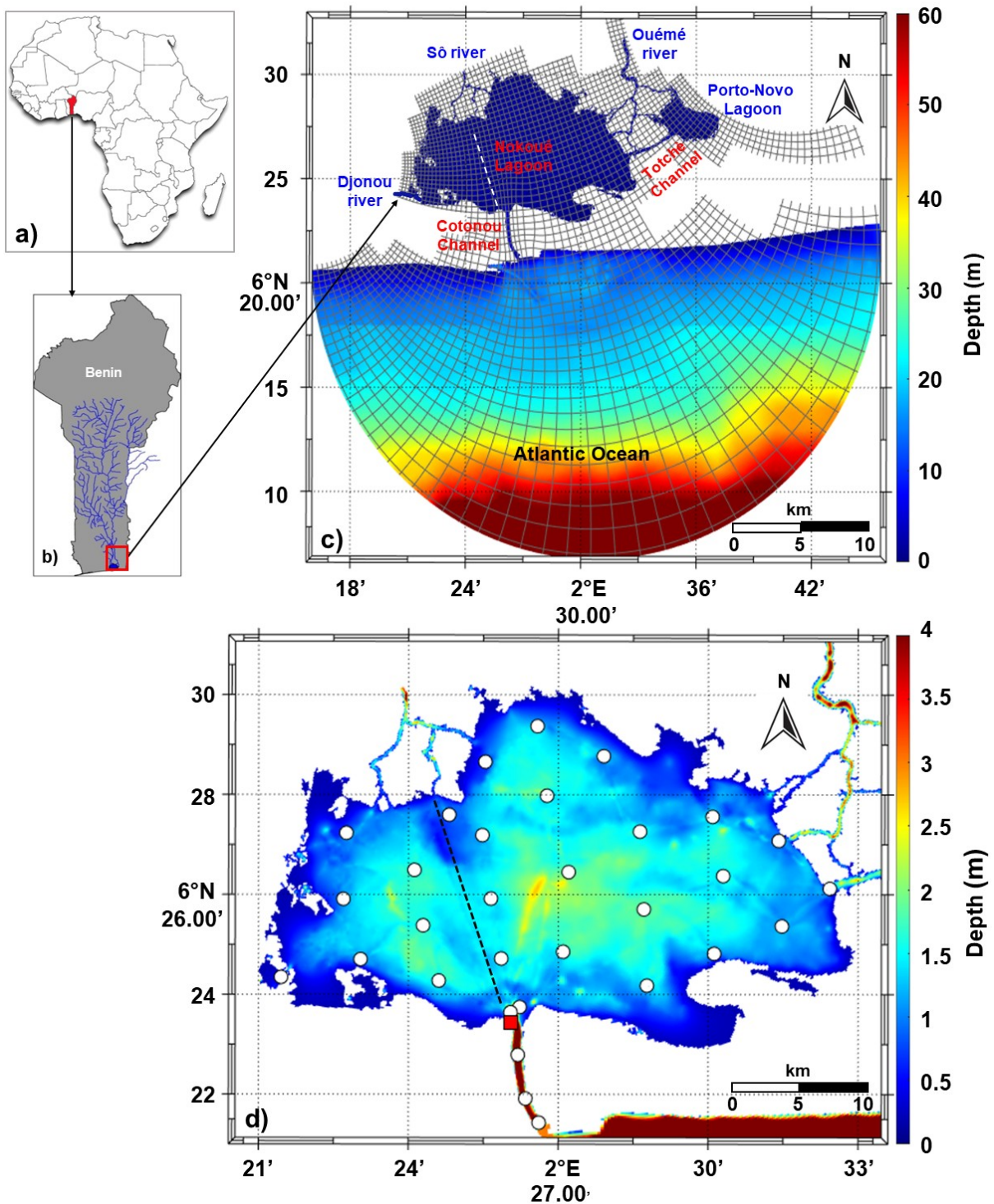
203 River discharges entering the lagoon during flood periods were estimated by
204 Morel et al. (2022). Fluxes during low water periods are more difficult to estimate and
205 required calibration. To account for this, we conducted sensitivity analyses and
206 determined that a minimum flux of $7.5 \text{ m}^3 \text{ s}^{-1}$ was necessary to realistically simulate
207 lagoon desalination near the rivers.

208 Three freshwater inputs were considered: the Djonou River, the Sô River and the
209 Ouémé River (Fig. 1). The net river discharge into the lagoon varied from $\sim 7.5 \text{ m}^3 \text{ s}^{-1}$
210 in the dry season to a maximum of $\sim 1100 \text{ m}^3 \text{ s}^{-1}$ during the wet season (Chaigneau et
211 al., 2022; Morel et al., 2022). Historical data from the SIEREM database
212 (<http://www.hydrosciences.org/spip.php?article1236>) indicated that the Ouémé and
213 Sô rivers contributed approximately to 70% and 30%, respectively, of the total river
214 discharge feeding into the lagoon from the northern side. Although the Djonou river
215 lacked measurement gauges, recent estimations suggest that this smaller river might
216 contribute approximately 2% to the overall flow entering the lagoon. Without any
217 complementary information or observations, we assumed these percentages
218 remained correct.

219 Our simulations cover the period between October 2016 and December 2018,
220 with a 12-month spin-up period from October 2016 to September 2017. Therefore,
221 the retained simulation extended from October 2017, when the lagoon was devoid of
222 salt, to December 2018. A computational time step of $\sim 30 \text{ s}$ was determined to

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223 satisfy the CFL condition (Blumberg & Mellor, 1987; Marsaleix et al., 2012) for our
224 simulations.



225

226 **Fig. 1.** Study area in southern Benin (West Africa), showing (a) the location of Benin, (b) the
227 hydrographic network of the Sô and Ouémé catchments in Benin, (c) the SYMPHONIE model grid
228 (each cell representing 10 grid points) and bathymetry of model domain, d) bathymetry of the Nokoué

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229 lagoon. The 31 CTD observation stations performed monthly between December 2017 and December
230 2018 are marked by white dots in d), the water level data logger position is shown by a red square,
231 and the black dashed line represents the location of south-north high CTD resolution section.

232 **2.2.3. Numerical simulations**

233 In this study, we construct a reference simulation (*Ref*) as the most realistic
234 representation of the Nokoué Lagoon's salinity dynamics, which considers the
235 physical drivers and calibrations previously described. To understand the importance
236 of each driver, we conducted three additional simulations:

237 - *NoTide*, in which the tidal forcing was removed while keeping all other forcings as in
238 the reference simulation.

239 - *NoSal*, in which the effect of salt in the density equation of state was removed within
240 the lagoon, to isolate the baroclinic effects associated with salinity gradients between
241 the ocean and the lagoon. However, the pressure gradient effects of salt were
242 retained in the ocean, maintaining the equation of state's integrity. Along the
243 Cotonou channel, which connects the two zones, a linear adjustment was
244 implemented to smoothly transition between the two regimes, accommodating both
245 the inclusion and exclusion of salt in the equation of state. This simulation was
246 designed to discern the specific impact of baroclinic effects within the Nokoué
247 Lagoon, a factor estimated previously using a rudimentary box model to account for
248 an average transport of approximately $40 \text{ m}^3 \text{ s}^{-1}$ (Okpeitcha et al., 2022). By isolating
249 these effects, the *NoSal* scenario provides valuable insights into the role of salinity
250 gradients and baroclinic effects in shaping lagoon dynamics and salinity fluxes.

251 - *NoWind*, in which the wind stress was removed from the momentum forcing but
252 kept in the mixing closure scheme.

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253 These 3 simulations were conducted from October 2017 to December 2018 and were
254 used to investigate the impact of each forcing on the salinization phase of the lagoon
255 observed between November 2017 and February 2018. To further study the
256 sensitivity of the lagoon's salinity to variations in river inflow, we conducted 13
257 additional simulations during the desalinization phase (May to September 2018).
258 These *RunOffs* simulations involved the addition of constant river discharge
259 increments, ranging from $30 \text{ m}^3 \text{ s}^{-1}$ to $900 \text{ m}^3 \text{ s}^{-1}$, to the dry season outflow of the *Ref*
260 configuration ($\sim 7.5 \text{ m}^3 \text{ s}^{-1}$), starting from May 1, 2018 (beginning of the desalinization
261 phase). The distribution of flux rates between the Djonou, the Sô and the Ouémé
262 rivers was kept constant across all simulations. A description of each simulation is
263 provided in Table 1.

264

265 **Table 1.** Summary of Reference and sensitivity simulations

Numerical simulation	Tidal Forcing	Wind Forcing	Baroclinic Forcing	River Forcing	Other Forcings
<i>Ref</i>	Yes	Yes	Yes	Yes	Yes
<i>NoTide</i>	No	Yes	Yes	Yes	Yes
<i>NoSal</i>	Yes	Yes	No	Yes	Yes
<i>NoWind</i>	Yes	No	Yes	Yes	Yes
<i>RunOffs</i>	Yes	Yes	Yes	Constant value	Yes

266

267 **2.3. In-situ data**

268 In order to evaluate the model results, we used measured salinity and water-
269 level data. Salinity data were collected during 13 monthly surveys between
270 December 2017 and December 2018, with each survey including approximately 31
271 vertical temperature and salinity profiles taken over Nokoué Lagoon using a
272 Conductivity Temperature Depth (CTD) probe (Valeport MIDAS CTD+ 300, white
273 dots in Fig. 1c). The CTD was sampled at a rate of 8 Hz and parameters were
274 recorded every ~10 cm from the surface to the bottom. Only the descent CTD profiles
275 were used. An objective interpolation method (Bretherton et al., 1976; McIntosh,
276 1990; Wong et al., 2003) was used to spatially interpolate the 31 irregularly
277 distributed salinity data onto a regular grid of ~100 m × 100 m resolution.
278 Additionally, high spatial resolution CTD data were collected over a south-north
279 section during each survey, consisting of 24 stations spaced 250 m apart, from the
280 Cotonou channel entrance to the mouth of the Sô river (white dashed line in Fig. 1c).
281 Further information regarding the salinity dataset can be found in Okpeitcha et al.
282 (2022).

283 Water-level measurements were acquired by a submerged pressure sensor
284 (HOBO U20L-01, (Abimbola et al., 2020; Guragai et al., 2018) installed in the
285 southwestern part of the lagoon, where it connects with the Cotonou channel (red
286 square in Fig. 1c). This sensor recorded the absolute water pressure every 20
287 minutes since February 2018, providing a continuous measurement of water level
288 variations in the lagoon. The water level data are considered representative of the
289 overall water level changes in the lagoon; a detailed analysis of the dataset is
290 available in Chaigneau et al. (2022).

291 **2.4. Mean salt flux decomposition**

292 In the following section, the model results will be analysed to study the
 293 mechanisms that are responsible for the salt variations in the lagoon during the
 294 salinization phase. In the model, the equation for the evolution of salinity is a simple
 295 advection diffusion equation and can be written in terms of fluxes

$$296 \partial_t S = -\vec{U} \cdot \nabla S + \partial_z (K_z \partial_z S)$$

297 where S is salinity, \vec{U} the 3D velocity field and K_z the diffusion coefficient. Integrating
 298 over a time period $[0, T]$ the net variation of salinity can be related to mean fluxes
 299 over the time period, as:

$$300 S(T) - S(t=0) = -\int_0^T (\vec{U} \cdot \nabla S) + \partial_z (\overline{K_z \partial_z S}).$$

301 We concentrate on the contribution of horizontal salt fluxes at different timescales.
 302 The mean horizontal salt flux \overline{US} over a period $\Delta T < T$ (denoted $\overline{US}^{\Delta T}$) is decomposed
 303 into an average term and a fluctuation term:

$$304 \overline{US}^{\Delta T} = \overline{U}^{\Delta T} \overline{S}^{\Delta T} + \overline{U' S'}^{\Delta T} \quad (1)$$

305 where $\overline{U}^{\Delta T}$ et $\overline{S}^{\Delta T}$ represent the average values of velocity and salinity respectively
 306 over a period ΔT , U' et S' are the fluctuations around the mean. Thus, the term

$$307 \overline{U' S'}^{\Delta T} = \overline{US}^{\Delta T} - \overline{U}^{\Delta T} \overline{S}^{\Delta T} \quad (2)$$

308 represents the effect of processes with periods less than ΔT , on the total fluxes
 309 during ΔT .

310 Now, over $[0, T]$, the total salt flux can be obtained by integrating (1):

$$311 \overline{US}^T = \overline{\overline{U}^{\Delta T} \overline{S}^{\Delta T}} + \overline{\overline{U' S'}^{\Delta T}} \quad (3)$$

312 which leads to

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$$\overline{U' S'^{\Delta T T}} = \overline{U S^T} - \overline{U^{\Delta T} \overline{S}^{\Delta T T}} \quad (4)$$

314

315 $\overline{U' S'^{\Delta T T}}$ here represents the net effect of processes with period less than ΔT on the
316 total salt fluxes over the whole period $[0, T]$.

317 By distinguishing two periods $\Delta T_1 < \Delta T_2$, we deduce from equation (4) that:

$$\overline{U' S'^{[\Delta T_1, \Delta T_2] T}} = \overline{U' S'^{\Delta T_2 T}} - \overline{U' S'^{\Delta T_1 T}} = \overline{U^{\Delta T_1} \overline{S}^{\Delta T_1 T}} - \overline{U^{\Delta T_2} \overline{S}^{\Delta T_2 T}} \quad (5)$$

318 where $\overline{U' S'^{[\Delta T_1, \Delta T_2] T}}$ represents the net effect of processes associated with time periods
319 between $[\Delta T_1, \Delta T_2]$ on the total salt fluxes over the period $[0, T]$.

320 This approach is similar to the method of coarse grain decomposition (see Aluie et
321 al., 2018, and references therein) but applied to time instead of space.

322 In our study, we first select a global time period T , which corresponds to the
323 salinization period from mid-November to mid-February (see Section 3.2 and
324 Okpeitcha et al., 2022). From equation (5), we then decomposed the total salt flux
325 over this salinization period into a sum of fluxes over sub-periods smaller than T .
326 While the semi-diurnal tide dominates, the O1 and K1 component, and other forcings
327 -such as wind- are associated with significant diurnal variability (see Chaigneau et
328 al., 2022). The basic period retained in this study is thus twice the semi-diurnal lunar
329 period, related to all high frequency processes, $[0, T_{HF} = 24 \text{ h } 50]$. We then consider
330 the periods which is specific to spring/neap tides $[T_{HF}, T_{15d} = 14 T_{HF}]$ and the low
331 frequency $[14 T_{HF}, T]$.

332 Thus, the contributions of these different temporal sub-periods to the total flux over
333 the salinization period can be written respectively:
334

$$\overline{U' S'^{[0, T_{HF}] T}} = \overline{U S^T} - \overline{U^{T_{HF}} \overline{S}^{T_{HF} T}} \quad (6)$$

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$$\overline{U^i S^i [T_{HF}, T_{15d}]^T} = \overline{U^T}_{HF} \overline{S^T}_{HF} - \overline{U^T}_{15d} \overline{S^T}_{15d} \quad (7)$$

$$\overline{U^i S^i [T_{15d}, T]^T} = \overline{U^T}_{15d} \overline{S^T}_{15d} - \overline{U^T} \overline{S^T} \quad (8)$$

338 By combining the different terms of equations (6), (7) and (8) member by member,
339 we obtain:

$$\overline{US^T} = \overline{U^T} \overline{S^T} + \sum_n \overline{U^i S^i [T_n, T_{n+1}]^T} \quad (9)$$

341 with n the number of sub-periods (3 in our case). Note that each term in equations (6-
342 9) can be calculated from the fields $\overline{US^T}_{HF}$, $\overline{U^T}_{HF} \wedge \overline{S^T}_{HF}$, which are stored during the
343 simulation.

344 **3. Results and discussion**

345 **3.1. Model validation**

346 **3.1.1. Salinity distribution**

347 Figure 2 presents a comparison between the modeled monthly salinity
348 distribution and the observed values, taken both at the surface and bottom of the
349 lagoon. In situ salinity data were collected on a monthly basis within a relatively short
350 time period of 2 days. Consequently, to validate the model, the salinity outputs were
351 averaged over the 2-day duration of each field campaign. It is worth noting that
352 conducting a more precise extraction of the model's salinity fields at the closest time
353 of each CTD measurement did not yield any significant deviations in the results.

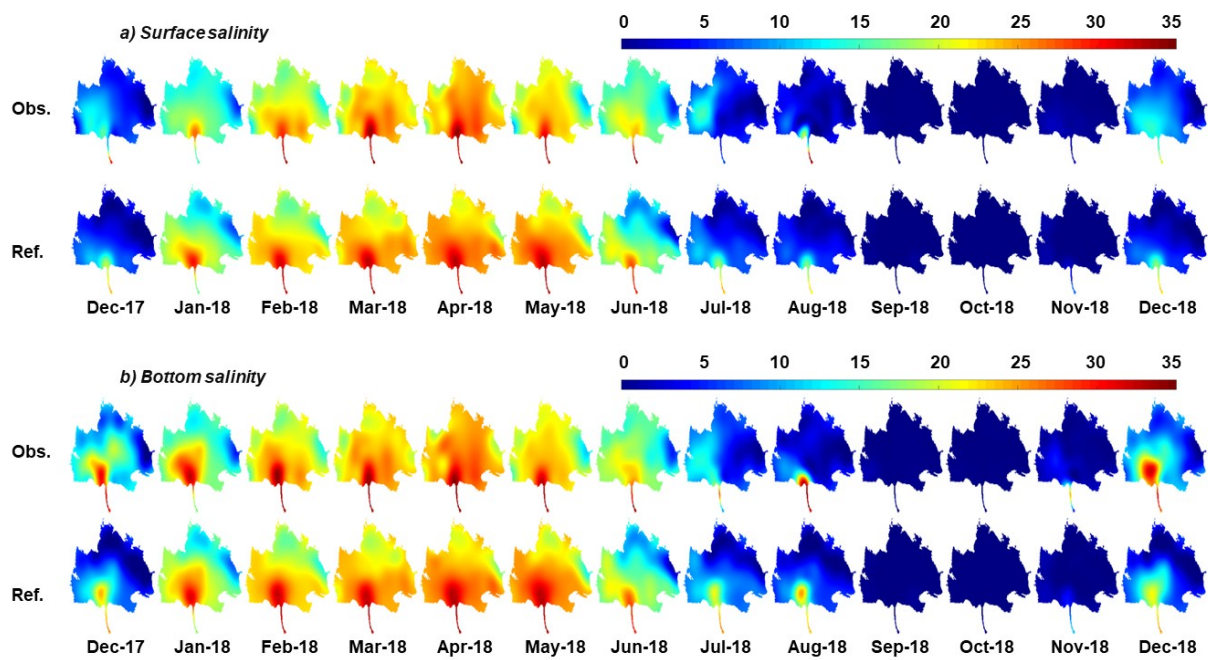
354 The salinization of the lagoon starts at the bottom layer in the southwest during
355 December (Dec-2017 and Dec-2018), at the beginning of the dry season (Fig. 2b).
356 From January to April, salinity gradually increases, with the lagoon bottom always

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357 saltier than the surface (Fig. 2a and b). The maximum salinity observed in April
358 reaches up to 35 in the South of the lagoon and 25 in the North near the river
359 mouths. From May to August, during the main wet season in southern Benin, salinity
360 decreases rapidly, with the western part of the lagoon being saltier than the eastern
361 part. The difference in salinity between the West and East in August is ~ 4 . During the
362 flood period (September-November), which is the second wet season, the entire
363 lagoon is filled with fresh water, and no salt is observed in either the model or the
364 observations.

365 The model realistically reproduces the spatial and seasonal distributions of the
366 observed salinity. In particular, it reproduces the gradual increase in salinity observed
367 between December and April, both at the surface and at the bottom, as well as the
368 important differences between surface and bottom, indicating marked vertical
369 stratification, especially in December-January (see also Section 3.1.2). The model
370 also captures the desalinisation phase observed between May and August at the end
371 of the main wet season. The spatial salinity patterns observed in the lagoon,
372 including the gradient from the Cotonou channel towards the North, where the lowest
373 salinity values are observed at the river mouths, are qualitatively reproduced by the
374 model. However, it should be noted that the model shows slight discrepancies in
375 salinity levels compared to the observations. Specifically, the model shows lower
376 salinity than observed in December and slightly higher salinity in May. These
377 disparities could be linked to the model's empirical estimation of river fluxes, detailed
378 in Section 2.2.2. Variations in both climatological data driving the model's river
379 discharge and the actual 2017-2018 river discharge, as well as differences in
380 distribution among contributing rivers (Sô, Ouémé, Djonou), contribute to

381 discrepancies between observed and modeled salinity, particularly noticeable during
382 salinization and desalination phases.



383

384 **Fig. 2.** Comparison of monthly (a) surface and (b) bottom salinity in the Nokoué lagoon between
385 December 2017 and December 2018, as observed (top row) and modelled (bottom row).

386 Figures 3a and 3c demonstrate the strong agreement between the modelled
387 salinity fields, extracted at the CTD measurement locations, and the corresponding
388 in-situ data, both at the surface (Fig. 3a) and the bottom (Fig. 3c). Notably, a nearly
389 linear relationship is observed between the modelled and observed data, with a
390 global correlation coefficient of 0.94 at the surface and 0.93 at the bottom. The lowest
391 correlation coefficients are found during the desalination phase, with values of 0.3
392 (0.6, respectively) at the surface (bottom) in July (represented by blue dots in Fig. 3a
393 and 3c), and 0.8 (0.7) at the surface (bottom) in August (represented by black dots in
394 Fig. 3a and 3c). Notably, discrepancies observed during the transitional phases of
395 salinization/desalination, especially in regions of relatively low salinity near the river
396 mouths, may be linked to subtle differences between the mean seasonal cycle of

397 river discharge used to drive the model and the actual 2017-2018 river discharge.
398 Slight variations in the volume and timing of river discharge between the datasets are
399 likely contributing to localized discrepancies in salinity.

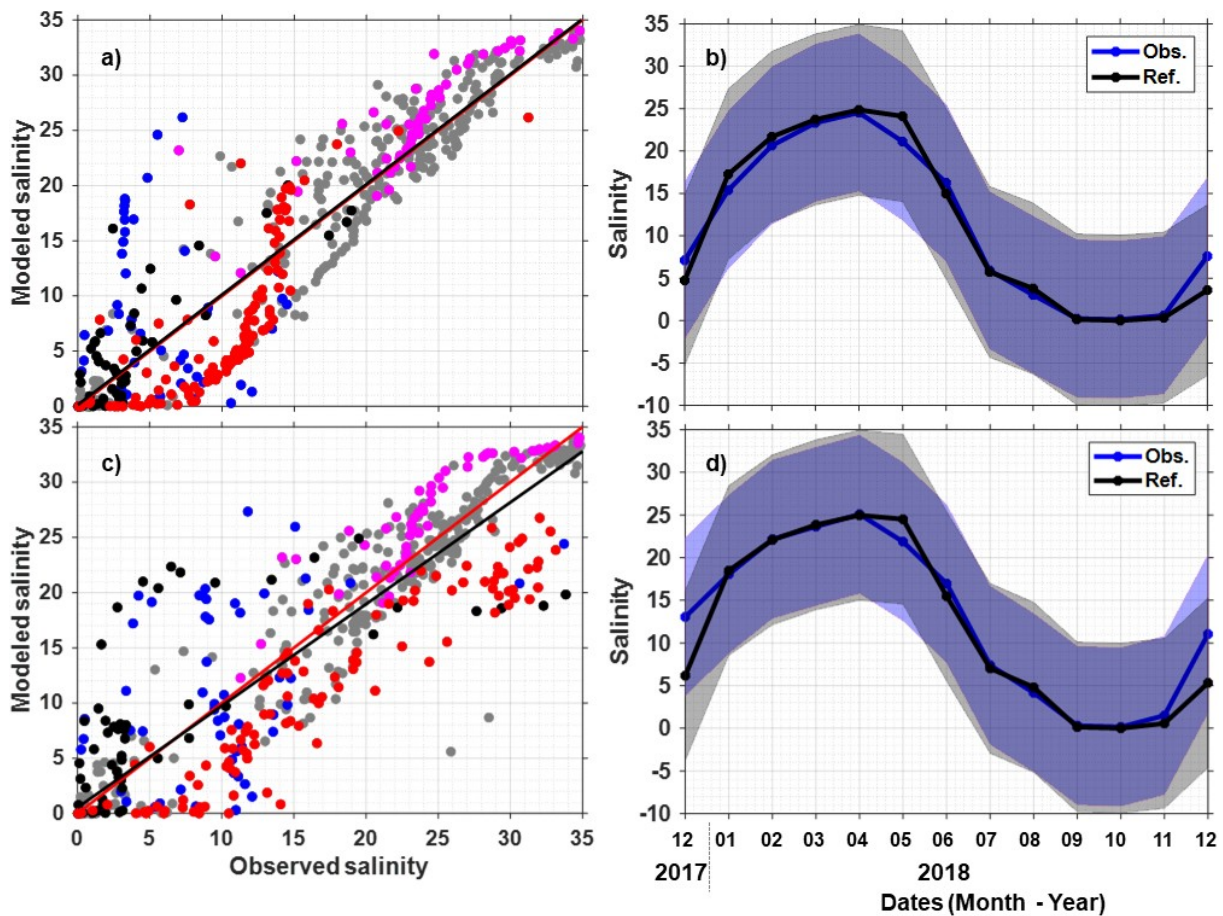
400 Figures 3b and d display the monthly average surface and bottom salinities
401 across the entire Nokoué lagoon, with each monthly value representing a 2-day
402 average. As shown, these salinities exhibit large seasonal variations. Specifically,
403 from December to April, both surface and bottom salinities experience a rapid
404 increase, peaking at 25 in April. Subsequently, from May to August, salinity
405 decreases, reaching a minimum of 5 at both the surface and bottom. From
406 September to November, the lagoon becomes filled with freshwater, rendering
407 salinity negligible.

408 The model performs well in capturing the observed global average salinity
409 variations for both the surface (Fig. 3b) and bottom (Fig. 3d) of the lagoon. However,
410 some discrepancies are evident. In December (at the onset of salinization), the
411 model tends to underestimate the observations, as also shown in Fig. 3a and 3c (red
412 dots), with a difference of approximately 5. Additionally, in May (at the onset of
413 desalinization), the model slightly overestimates the observations, with a difference of
414 approximately 2 (see also magenta dots in Fig. 3a,c).

415 The disparities observed between the modeled and observed salinity fields might
416 result from differences in the time periods of data acquisition between the
417 observations and the model, particularly considering the significant salinity variability
418 near the channel occurring at a semi-diurnal scale. Additionally, uncertainties in the
419 model drivers, especially concerning river discharge, could contribute to these
420 discrepancies. Moreover, the relatively smooth bathymetry within the model may
421 influence salinity outcomes; a shallower bathymetry could restrict salt intrusion into

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422 the lagoon during the salinization phase. Despite these considerations, the model
423 demonstrates highly satisfactory statistical performance. The Nash-Sutcliffe
424 efficiency coefficient (Nash & Sutcliffe, 1970), the ratio of standard deviations, and
425 root-mean-square error, for the surface (bottom, respectively) salinity of both the
426 model and observations in the monthly mean are 0.96 (0.91), 0.92 (0.93), and 1.70
427 (2.64), respectively.



428
429 *Fig. 3. Comparison of modeled and observed salinity fields. a, c) Scatter plots of modelled vs*
430 *observed surface (a) and bottom (c) salinity at the CTD stations for the 13 field campaigns conducted*
431 *from December 2017 to December 2018. The black line represents the linear relationship, while the*
432 *red line depicts the 1:1 relationship. The magenta, blue, black, and red dots correspond to the months*
433 *of May, July, August, and December, respectively, while the grey dots represent the rest of the year.*
434 *b, d) Monthly mean surface salinity (d) and bottom salinity (b) between December 2017 and*

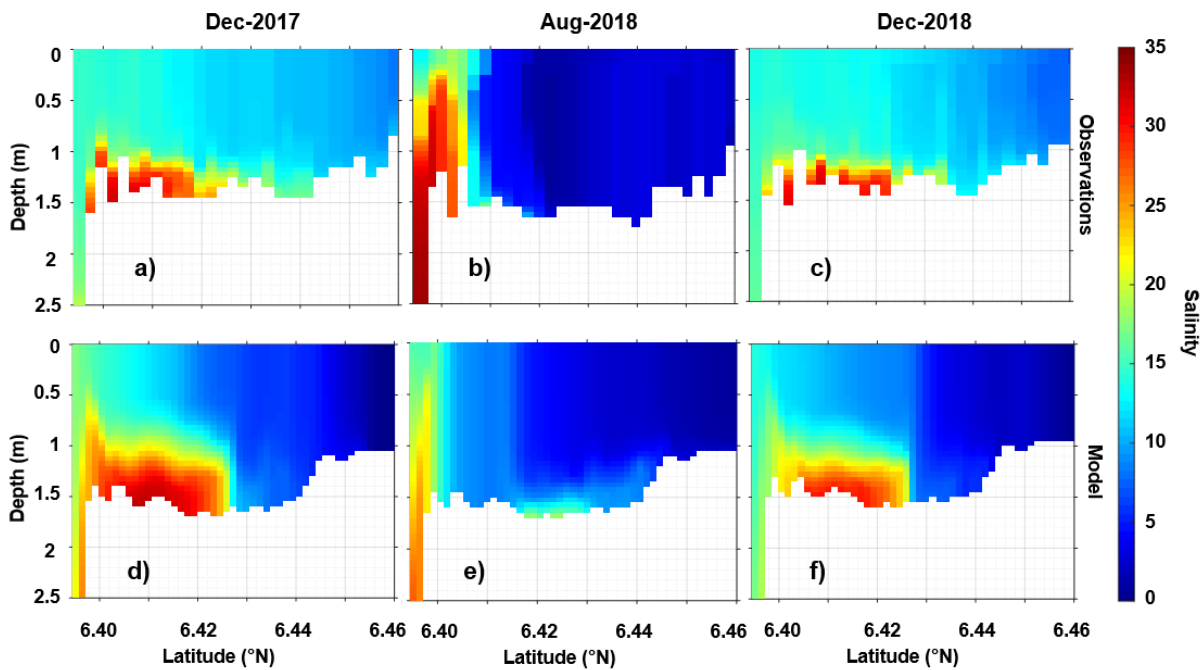
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435 *December 2018. The color shading indicates the range of ± 1 standard deviation around the monthly*
436 *mean.*

437 **3.1.2. Lagoon stratification**

438 Figure 4 shows the observed and modeled vertical structure of salinity along the
439 high-resolution section between the entrance of Cotonou channel (South) and Sô
440 river mouth (North) (white dashed line in Fig. 1c). In December 2017 and 2018, the
441 southern lagoon's observed bottom salinity reaches 30, with a salinity difference of
442 about 15 between the bottom and the surface salinity, while the stratification remains
443 weak near the river mouth. In August 2018, the bottom salinity of the lagoon has
444 increased, but the salinity difference between bottom and surface remains important
445 (~10), despite the lagoon being shallow. Apart from this limited area in the south, the
446 rest of the lagoon is filled with fresh water.

447 The model accurately reproduces the observed salt-wedge at the beginning of
448 the salinization (Fig. 4a and 4d) and desalinization periods (Fig. 4b and e), which
449 suggests that the density difference between salty ocean water and fresh river water,
450 along with associated gravity currents, may play a significant role in salt penetration
451 in the lagoon (Okpeitcha et al., 2022). The presence of this fine-scale vertical salinity
452 structure provides further confidence in the model's ability to reproduce the lagoon
453 hydrodynamics and the dispersion of salt.



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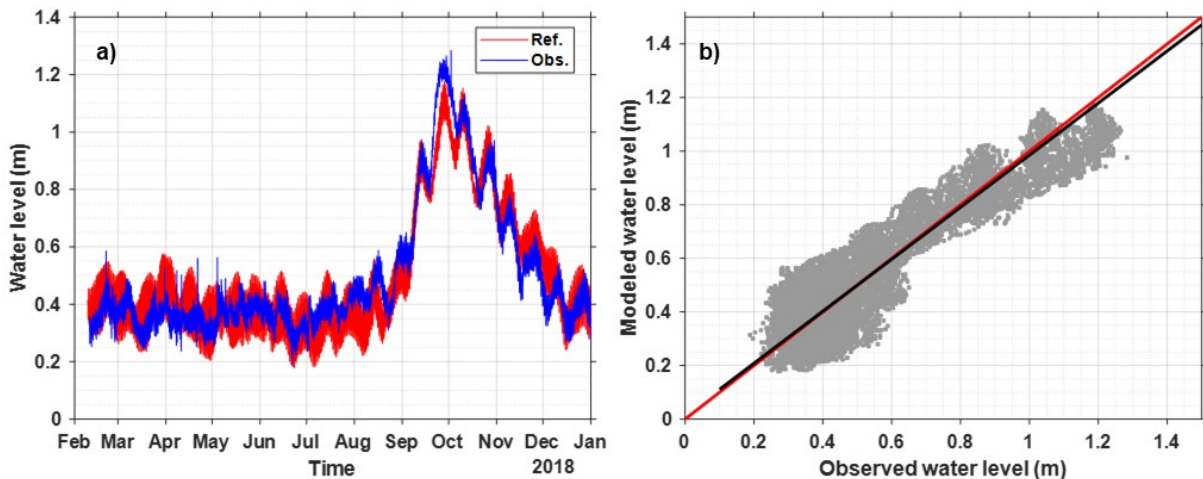
455 *Fig. 4. Vertical salinity structure of the Nokoué Lagoon along a south-north section from the entrance*
 456 *of the Cotonou Channel (south) to the mouth of the Sô River (north) during the salinization*
 457 *(December) and desalination (August) phases. The upper panel shows observations, while the lower*
 458 *panel shows model results. The location of the south-north section can be found in Fig. 1.*

459 3.1.3. Water-level variation

460 We present a comparison of the water level observed at a specific point located
 461 south of the Nokoué lagoon (marked by a red square in Fig. 1) with the
 462 corresponding model results. The model successfully simulates the different scales
 463 of water level variability observed by Chaigneau et al., (2022), including the seasonal
 464 variations and the 14-day cycle associated with spring and neap tides (Chaigneau et
 465 al., 2022; Morel et al, 2022) (Fig. 5a). However, the maximum water level during the
 466 peak of the wet season is underestimated by the model by ~5 cm compared to the
 467 observations. Additionally, the tidal amplitudes are slightly overestimated in the
 468 model (~6-7 cm) compared to the observations (~3-5 cm) at the location of the
 469 observation, but the model reproduces the tidal phase consistently with the

470 observations. We note that the peaks in the observations are related to a seiche
471 phenomenon (Chaigneau et al., 2022) which are not represented in the model due to
472 the absence of high-frequency winds and squall lines in the ECMWF fields used for
473 atmospheric forcing.

474 Despite the discrepancies between observations and model results at high
475 frequencies, the modelled water-level variations match the observed variations well
476 (Fig. 5b). The Nash–Sutcliffe efficiency coefficient, ratio of standard deviation and
477 root-mean-square error for both model and observations are 0.86, 0.94 and 0.08 m,
478 respectively.



479

480 *Fig. 5. Comparison of modelled and observed water-level time series from February 2017 to*
481 *December 2018. a) Time series variation of water-level. b) Scatter plot showing the relationship*
482 *between modeled and observed water-level datasets. The black line represents the linear relationship,*
483 *while the red line depicts the 1:1 relationship.*

484 In summary, the good qualitative and quantitative agreement between the
485 simulated and observed salinity and water-level variations demonstrate that the
486 model accurately reproduces the lagoon dynamics. Hence in the following, we use
487 the various simulations to understand the physical processes and mechanisms
488 driving the observed salinity variations in the lagoon. We note that the long and

489 narrow Cotonou channel makes this lagoon a strongly choked system, and the
490 differences in the observed and modelled water-level variations at both high and low
491 frequencies are consistent with a lack of dissipation in the channel, possibly related
492 to the presence of large pylons of three bridges and rocky blocks obstructing the
493 channel entrance, which are not considered in the current model configuration.

494 **3.2. Main drivers of the Nokoué lagoon salinity**

495 In this section, we aim to gain a deeper understanding of the individual
496 contributions of the main forcings - tide, baroclinic fluxes related to salt gradients,
497 winds, and freshwater fluxes - to the salinity dynamics of Nokoué lagoon. To do so,
498 we analyze the results of the simulations, including the *NoTide*, *NoSal*, *NoWind*, and
499 *RunOffs* scenarios. Note that in this analysis, results from surface and bottom level
500 simulations are consistent, and therefore, we focus on the surface results only in the
501 following discussion.

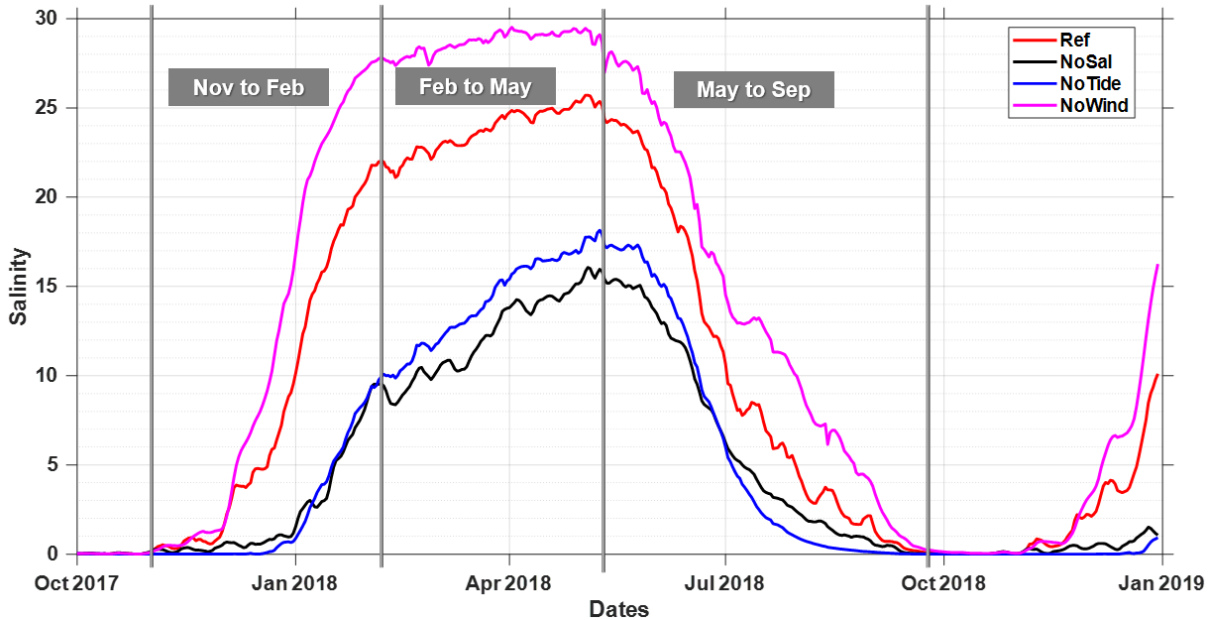
502 **3.2.1. Global impact of the main forcings on seasonal salinity variations**

503 Fig. 6 shows the daily mean surface salinity of the *Ref* simulation (red curve)
504 from November 2017 to December 2018. From November to February, the salinity
505 increases progressively from 0 to 22, marking the salinization phase. From mid-
506 February to May, the salinity remains quasi-stable at around 22-25, representing the
507 quasi-equilibrium phase. From May to October, the salinity decreases rapidly from 25
508 to 0, marking the desalination phase. We will focus on the salinization and
509 desalination phases in the following analysis.

510 Global impact of specific forcings can be inferred from the simulations removing
511 driving processes. This demonstrates the importance of the tide but also of the wind
512 and the baroclinic effects in determining the global salinity level of the lagoon. The

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513 mean surface salinity in *NoTide* (blue curve) and *NoSal* (black curve) follows a
514 similar variability to the *Ref* simulation, but with significantly lower intensities. The
515 salinity in *NoTide* varies from 0 to 10 from November to February, from 10 to 18 from
516 February to May and 18 to 0 from May to October. In the *NoTide* and *NoSal*
517 experiments, quasi-equilibrium is not reached, and the variation in salinity from
518 February to May is much greater than in *Ref*. *NoSal* remains less salty than *NoTide*
519 with a difference of ~ 2 in salinity between February and June, highlighting the crucial
520 role of the baroclinic effect in the salinization mechanism of the lagoon.
521 In the *NoWind* simulation, the salinity is greater than in *Ref* but follows a similar
522 variation. The difference in salinity between the *Ref* and *NoWind* simulations is ~ 5 .
523 The impact of the wind forcing on the salinity dynamics will be further discussed in
524 the following analysis.



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526 Fig. 6. Daily mean lagoon surface salinity variations for the *Ref* (red), *NoTide* (blue), *NoSal* (black) and
527 *NoWind* (magenta) simulations, and selected phases.

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528 **3.2.2. Tidal, baroclinic and wind effects during the salinization phase**

529 Here we aim to assess the total change in salinity during the salinization
530 phase (November-February) for each simulation. To achieve this, as explained in
531 Section 2.4, we analyze the total salt fluxes throughout the entire salinization phase,
532 as well as the average fluxes at different temporal scales. However, we observed
533 that fluxes at intermediate frequencies (24h50 to 15 days) are significantly weaker.
534 Hence, we focus on high frequency (i.e., periods <24h50) and low frequency (i.e.,
535 periods >15 days) scales.

536 During the salinization phase (Fig. 7), the *Ref* simulation shows that the mean
537 horizontal salt fluxes structure (Fig. 7e) is similar to that of the high-frequency fluxes
538 (Fig. 7i), but the latter are more intense. The high intensity of salt flux at high
539 frequency in the southern part of the lagoon (Fig. 7i) indicates the important role of
540 the tide in supplying salt to this region. For the rest of the lagoon, high-frequency
541 processes play important role in spreading the saltwater from the southern input to
542 the entire lagoon. The low-frequency horizontal mean circulation (Fig. 7m), on the
543 other hand, has the opposite effect and is mainly directed from the rivers towards the
544 Cotonou Channel. This is due to the inflow of freshwater from the rivers, which tends
545 to 'push back' some of the salt water from the river mouths, leading to weaker salinity
546 variations in these regions during the salinization phase. Therefore, low frequency
547 fluxes are related to freshwater inflows and reduce salinization by high frequency
548 fluxes, related to tidal and other high-frequency processes.

549 In the *NoTide* simulation, salinity variations are significantly lower compared to
550 the *Ref* simulation (Fig. 7b vs. Fig. 7a, also Fig. 6). Nevertheless, the spatial
551 distribution of salinity in the two simulations shows strong similarities (Fig. 7a and b),
552 with high values in the southeast and low values in the north. High-frequency

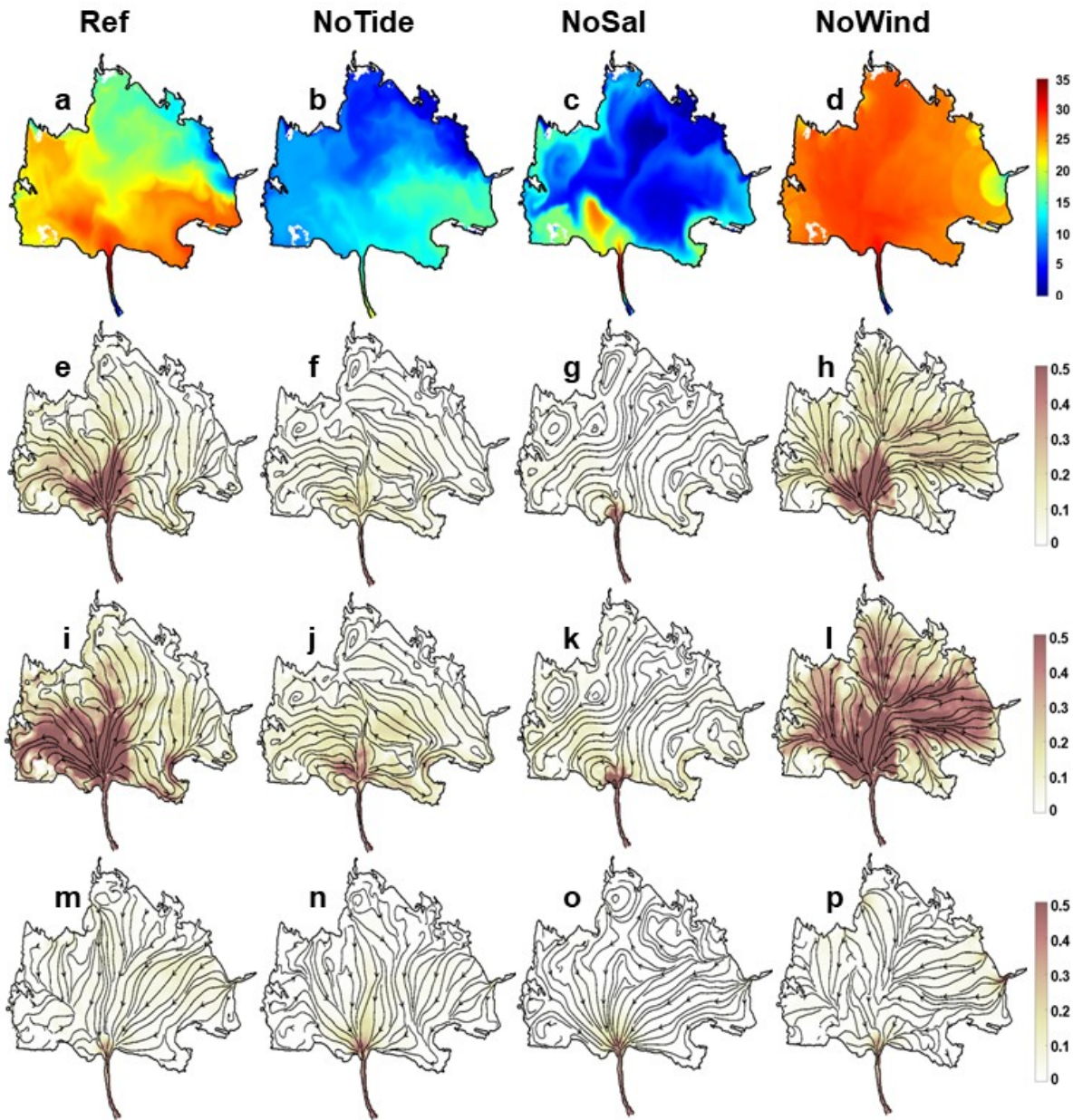
553 processes drive salinization in both *Ref* and *NoTide* simulations (Fig. 7f and j), while
554 low-frequency fluxes (Fig. 7n) limit it. Similar to the *Ref* simulation, the *NoTide*
555 simulation show similarities at low-frequencies (Fig. 7m and n), confirming that tidal
556 effects are not involved in low-frequency variability, which is mostly associated with
557 river discharges. The major differences between the *Ref* and *NoTide* simulations are
558 at high frequency, mostly at the channel mouth, where the flux intensity is
559 significantly lower in the *NoTide* simulation (Fig. 7i and j). Thus, the *NoTide*
560 simulation confirms that the tide plays a major role in the salt input to the south of the
561 lagoon, which then spreads throughout the lagoon through other high-frequency
562 processes limited by low-frequency fluxes generated by the rivers. Overall, this
563 confirms that the tide is crucial for the overall salt input into the lagoon, but other
564 processes such as baroclinic flows and winds also play an important role in the
565 redistribution of southern salt input within the lagoon.

566 In the *NoSal* simulation, salt is a passive tracer inside the lagoon, and its
567 influence on dynamics is eliminated (see Section 2.2.3). The mean flux (Fig. 7g) and
568 high-frequency flux (Fig. 7k) are similar but very different from their counterparts in
569 the *Ref* simulation (Fig. 7e and i). Baroclinic effects related to salt gradients increase
570 salt supply at the mouth of the channel in the south of the lagoon, but mainly allow for
571 its redistribution throughout the lagoon by high frequency processes. The low-
572 frequency salt fluxes are again similar to *Ref* and *NoTide* simulations, confirming
573 their association with inflow of fresh water from the rivers. Overall, without the effect
574 of salt gradients and under tidal conditions, the lagoon becomes slightly less salty
575 than in the *NoTide* experiment (see also Fig. 6). Only the southwestern region of the
576 lagoon near the mouth is saltier than the *NoTide* experiment (Fig. 7c).

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577 In the *NoWind* simulation, the mean fluxes are similar to *Ref*, except in the
578 North-East of the lagoon, where the mean and high-frequency salt fluxes are much
579 stronger. The penetration of salt is associated with high-frequency fluxes, while low-
580 frequency fluxes exhibit some differences. This is expected as a baroclinic circulation
581 associated with the mean wind is anticipated, with water pushed in the direction of
582 the wind at the surface and the opposite effect at the bottom (Bernardi et al., 2012;
583 Cerralbo et al., 2016; Lacy et al., 2003; O’Callaghan & Stevens, 2011). However, the
584 most striking effect is that high-frequency variability associated with the wind keeps
585 the saltwater from penetrating further to the North-East.

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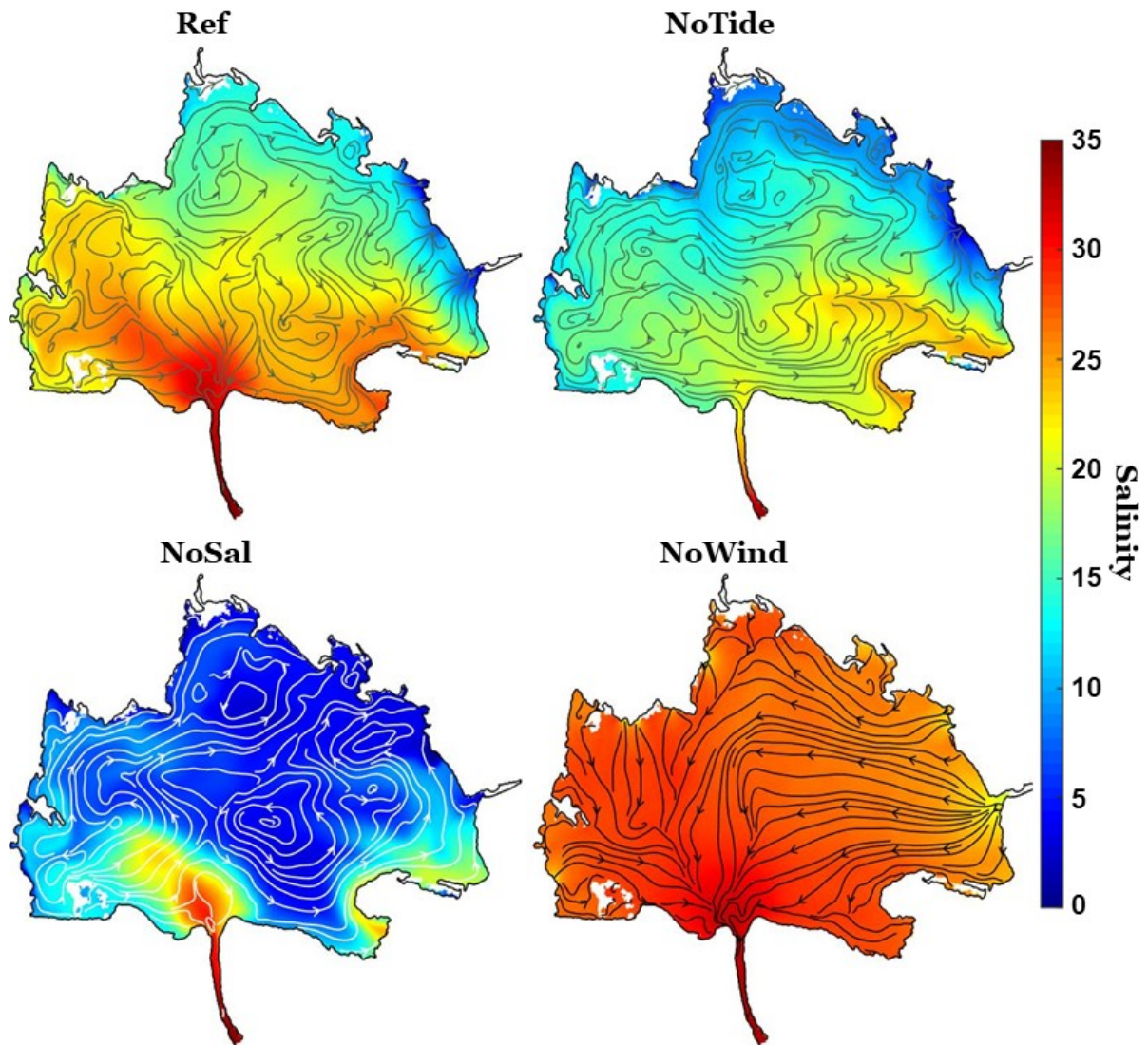
587 *Fig. 7. Salinity and salt flux evolution during the salinization period (November to February). Upper*
 588 *panel: Total surface salinity variability (difference between February and November surface salinity)*
 589 *for a) Ref, b) NoTide, c) NoSal and d) NoWind simulations. Second panel from the top: Total mean*
 590 *salt flux. Third panel from the top: high-frequency (<24h50) salt flux. Lower panel: low- frequency (>15*
 591 *days) salt flux. The salt flux unit is PSU m s⁻¹, and streamlines represent the direction of the salt flux*
 592 *propagation.*

593 To analyse this effect, an additional simulation (not shown here) was
 594 conducted with seasonal (low frequency) wind forcing. This simulation resulted in a

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595 salinity structure and fluxes that were very close to the *Ref* simulation, showing that
596 the high-frequency variations of the wind itself plays no role in limiting the salt
597 penetration to the North-East. Our interpretation of this counterintuitive result is that
598 the permanent winds generate small scale and high-frequency circulations, either
599 through shear instabilities or under the influence of topography. Indeed, as shown by
600 Fig. 8, daily outputs of the current and salinity fields confirm the generation of
601 horizontal recirculation patterns at small-scale and high-frequency when the wind is
602 taken into account and that these features disappear in the NoWind configuration.
603 The instantaneous (daily) salinity distribution in the lagoon is also strongly correlated
604 with the streamlines, in particular at small scales, showing the influence of eddies in
605 the redistribution of salt. Further studies are required to confirm this interpretation
606 and analyse the details of the processes responsible for the generation of these
607 small-scale recirculation patterns.

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609 *Fig. 8. Instantaneous surface salinity and current streamlines on the 27th of February during the*
610 *salinization phase for Ref, NoTide, NoSal and NoWind simulations. Note the strong correlation*
611 *between salinity distribution and small scale circulation patterns.*

612 To conclude, the experiments conducted reveal that tide-driven high-
613 frequency fluxes transport a significant amount of salt from the ocean to the southern
614 part of the lagoon through the Cotonou channel. Although the initial salt input at the
615 mouth of the lagoon is critical, other processes are involved in the salinization of the
616 entire lagoon. The high frequency baroclinic fluxes intensify the salinity input at the
617 south and promote the redistribution of salt by gravity throughout the lagoon. Despite

618 its shallowness, the pressure gradient associated with the salt plays important role in
619 the penetration of saltwater in the lagoon. The rivers supply a permanent flow of
620 fresh water that limits the salinization of the entire lagoon. Lastly, wind patterns alter
621 the low-frequency fluxes but their primary effect is to generate small-scale and high-
622 frequency currents, that partially inhibit the salinization of the north-eastern part of
623 the lagoon.

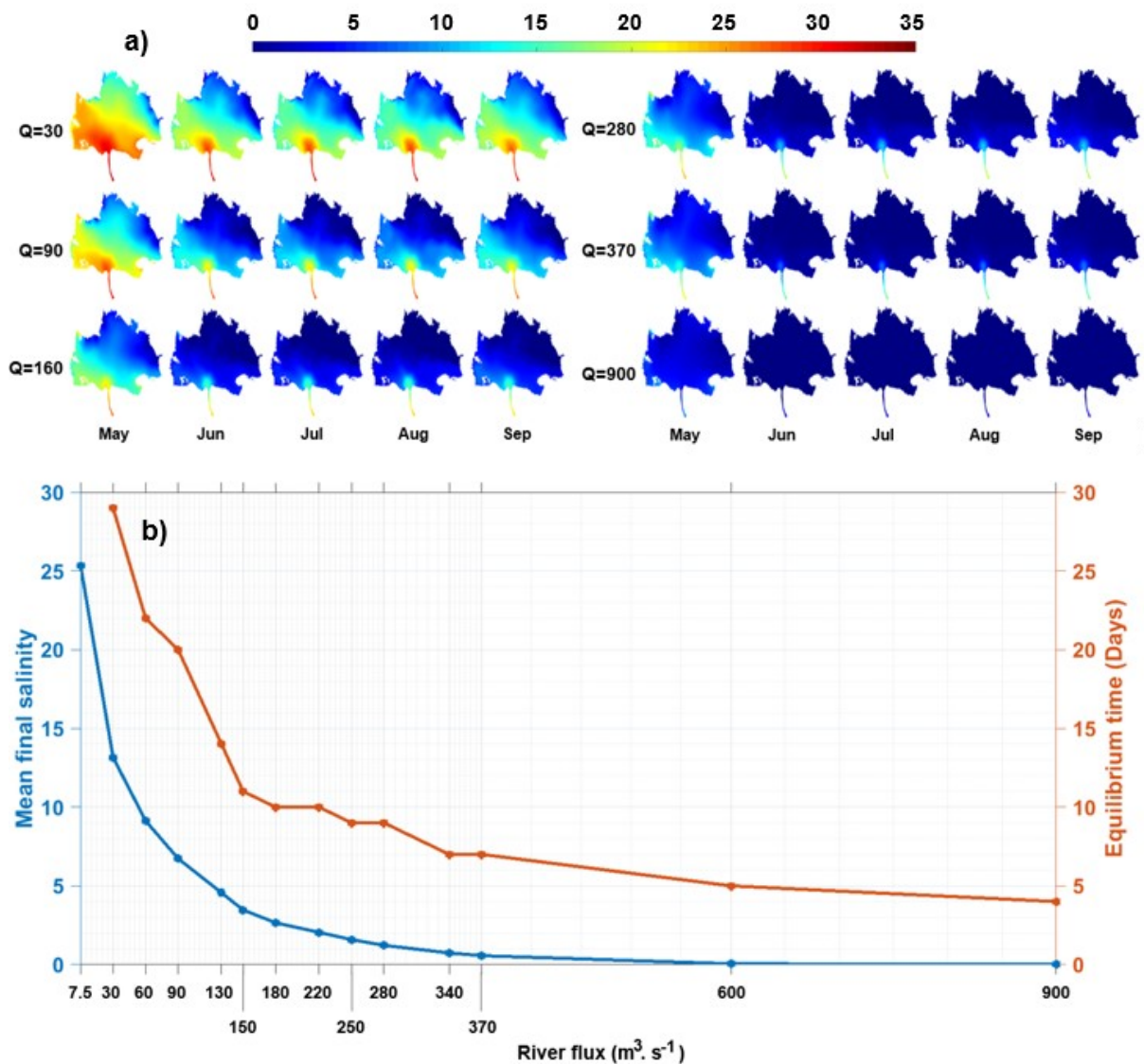
624 3.2.3. River discharge effects

625 In this section we investigate the impact of river inflow rates on the equilibrium
626 salinity structure in the lagoon and the threshold at which complete desalinization can
627 be achieved. Figure 9a shows an overview of the monthly salinity evolution in 6 of the
628 13 *RunOffs* simulations, all starting with a same condition from May 1, 2018, of the
629 *Ref* simulation. As expected, the more intense the river inflow, the faster the lagoon
630 desalinizes. There is a clear salt gradient between the north-east and south-west of
631 the lagoon, with the southern part remaining saltier due to the presence of brackish
632 water near the mouth of the Cotonou channel, which gradually disappears as the flow
633 increases. However, desalinization in most parts of the lagoon occurs much more
634 rapidly.

635 Figure 9b summarizes the results of the 13 scenarios and shows the average
636 value of the final surface salinity attained by the lagoon at equilibrium as a function of
637 the prescribed total river discharge (blue curve). At the end of the dry season (May),
638 the river flux in the *Ref* experiment is $7.5 \text{ m}^3 \text{ s}^{-1}$, and the mean value of the surface
639 salinity is about 25 (see Fig. 3). This final average salinity decreases rapidly with
640 higher river fluxes, varying from 13 for river flows of $30 \text{ m}^3 \text{ s}^{-1}$ to 7 for flows of $90 \text{ m}^3 \text{ s}^{-1}$,
641 which is consistent with the findings of Okpeitcha et al (2022) using a simplified box
642 model. The present experiment confirms this (Fig. 9b -blue curve-) by demonstrating

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643 that increasing the river discharge by $\sim 15 \text{ m}^3 \text{ s}^{-1}$ is required to decrease the average
644 salinity from 25 to 16. Above $100 \text{ m}^3 \text{ s}^{-1}$, the final average salinity in the lagoon falls
645 below ~ 5 and varies less rapidly due to the maintenance of a salinity bulge at the
646 lagoon entrance associated with the tide. When the flow rate reaches $500 \text{ m}^3 \text{ s}^{-1}$ and
647 above, the entire lagoon becomes desalinated. It is also apparent from Fig. 9a that
648 regardless of the river flux value, the salinity of the lagoon reaches a state of
649 equilibrium after a certain time interval. The red curve in figure 9b shows the time
650 taken for the lagoon's salinity to reach this equilibrium, which decreases as the river
651 flow increases, ranging from 30 to 9 days for flows between 30 and $280 \text{ m}^3 \text{ s}^{-1}$. For
652 flow values above $280 \text{ m}^3 \text{ s}^{-1}$, this time is almost constant at ~ 5 days.



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654 Fig. 9. a) Spatiotemporal surface salinity variability in response to river fluxes (Q , in $m^3 s^{-1}$). b)
655 Relationship between final mean surface salinity (blue curve) and time required to reach equilibrium
656 (red curve) of the lagoon under different river flows.

657 4. Conclusion

658 We developed a 3D numerical configuration, based on the SYMPHONIE
659 model, to study the salinity variations of the Nokoué lagoon. The model accurately
660 reproduced various features of the observed salinity and water level of the lagoon. To
661 study the processes responsible for salt dispersion in the lagoon, we performed
662 sensitivity studies by removing key drivers (tide, wind, influence of salinity on density)
663 or modifying rivers inflow.

664 We first analyzed the evolution of the mean lagoon salinity for three
665 experiments: *NoTide*, *NoSal*, and *NoWind*. Although these experiments showed a
666 similar pattern to the reference simulation, the intensities were significantly different.
667 The maximum average salinity reached in April in the reference simulation was 25,
668 whereas in *NoTide* and *NoSal*, it was only 17 and 15, respectively, and was reached
669 in May. This result highlights the strong influence of the tide in salt penetration into
670 the lagoon but also underlines the important role played by the baroclinic effect
671 despite the lagoon's shallowness. In the *NoWind* simulation, the salinity was ~5
672 greater than in reference simulation.

673 Our analysis of the role of tide, salt gradient and wind on the lagoon
674 salinization revealed that high-frequency fluxes associated with the ocean tide import
675 a significant amount of salt from the ocean via the Cotonou channel to the South of
676 the lagoon, and that this initial salt input is crucial for the salinization of the rest of the
677 lagoon. High-frequency baroclinic fluxes increase this initial local input of salt in the

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678 South and redistribute salt throughout the whole lagoon via gravity and baroclinic
679 pressure gradients. Finally, low-frequency winds generate high-frequency
680 recirculation that partially prevent the salinization of the North-East of the lagoon.

681 The rivers provide a permanent flux of fresh water, which limits the salinization
682 of the lagoon. The equilibrium salinity of the lagoon decreases with increasing river
683 flux and reaches a steady state within 2 months of the simulations' start, with the
684 presence of a layer of brackish water in the southwestern part of the lagoon. Above
685 $100 \text{ m}^3 \text{ s}^{-1}$, the final average salinity remains below 5, and above $500 \text{ m}^3 \text{ s}^{-1}$, the
686 entire lagoon is desalinated. At low flow conditions (equivalent to average salinity
687 above 5) Nokoué lagoon is highly sensitive to even small changes in river discharge.
688 The model developed is a valuable tool for managing the lagoon, such as evaluating
689 the impact of a dam construction, which regulates flows entering the lagoon, on the
690 lagoon salinity and therefore on its ecosystem. The study improves our
691 understanding of the Nokoué lagoon salinity dynamics and West African coastal
692 lagoons more generally.

693 Future research on the Nokoué lagoon should account for the three bridges
694 crossing the Cotonou channel and the rocky blocks to more accurately represent the
695 dissipation that influences the salinity initial input in the south of the lagoon and its
696 distribution within the lagoon. Additional studies are also needed to investigate
697 further the development of high-frequency recirculation patterns associated with the
698 permanent wind, which may be related to bathymetry gradients or shear instabilities
699 of the mean currents.

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