



The Alkalinity Blue Carbon Pump: Tidal pumping drives long-term CO₂ sequestration as alkalinity in salt marshes

Paul Kanfer¹, Céline Charbonnier¹, Sabine Schmidt¹, Bruno Deflandre¹, Aurélia Mouret², Eric Bénéteau², Martin Danilo¹, Dominique Poirier¹, Jimmy Desnues¹, Gérard Lebreton¹, Pierre Anschutz¹

5 ¹ Univ. Bordeaux, CNRS, Bordeaux INP, EPOC, UMR 5805, F-33600 Pessac, France

² Université d'Angers, Nantes Université, Le Mans Université, CNRS, LPG UMR 6112, 49000 Angers, France

Correspondence to: Pierre Anschutz (pierre.anschutz@u-bordeaux.fr)

Abstract. Tidal salt marshes are increasingly recognised as significant blue carbon ecosystems. However, their total carbon sequestration capacity remains underestimated because dissolved carbon outwelling and alkalinity export are rarely considered alongside sediment burial. Here, we combine tidal creek time series measurements, open incubations, and sediment core analyses to quantify all the main carbon flux pathways in a temperate macrotidal salt marsh of the Arcachon Bay (SW France). A binary salinity mixing model coupled to direct discharge measurements allowed us to isolate the net tidal pumping fluxes of total alkalinity (TA), dissolved inorganic carbon (DIC), dissolved organic carbon (DOC) and methane (CH₄) from conservative mixing. DIC outwelling ($5.3 \pm 1.7 \text{ mmol m}^{-2} \text{ d}^{-1}$) was the largest form of dissolved carbon export but it does not fully represent net sequestration as part of it consists of CO₂ that returns to the atmosphere via air-water exchange. Net TA lateral fluxes averaged $3.9 \pm 1.4 \text{ mmol m}^{-2} \text{ d}^{-1}$, driven by sulphate reduction coupled to pyrite formation and aerobic respiration in marsh sediments, as supported by TA:DIC ratios and $\delta^{13}\text{C}$ -DIC. DOC outwelling ($3.4 \pm 3.3 \text{ mmol m}^{-2} \text{ d}^{-1}$) was on same magnitude but its sequestration potential depends on lability. High tide benthic incubations revealed significant TA and DIC fluxes from sediments to the water column, higher in vegetated than in bare mud areas, driven by benthic processes and root-network macroporosity. Organic carbon burial rates estimated from ²¹⁰Pb_{xs} profiles were around $80 \text{ mmol m}^{-2} \text{ d}^{-1}$ in vegetated zones. However, these rates likely represent an overestimated gross burial rather than net sequestration given the dynamic and erosive nature of this macrotidal system. When integrating all measured carbon flux, net TA outwelling emerges as the most chemically conservative and durable sequestration mechanism, as it is resistant to remobilisation and can remain stable over millennial timescales once exported to the coastal ocean. We propose the term Alkalinity Blue Carbon (ABC) to describe this pathway, and call for its systematic inclusion alongside sedimentary C burial in future blue carbon inventories and climate mitigation frameworks.



1 Introduction

Among the different types of ecosystems that sequester carbon (C), coastal environments play a key role. These interfaces between the continent and the ocean are subject to significant continental inputs that result in high biological production, making these ecosystems among the most productive on the planet. In particular, mangroves, tidal salt marshes, seagrasses and brackish reed wetlands sequester and store large amounts of carbon, known as Blue Carbon (Lovelock and Duarte, 2019). As a result, a range of nations have included coastal wetlands in their climate change mitigation or adaptation activities following the 2016 Paris Agreement (Intergovernmental Panel on Climate Change (IPCC), 2014).

The total worldwide area of coastal wetlands is small compared with other ecosystems, and consequently, the total amount of C stored in coastal wetlands is relatively small with 10–25 Pg compared to 1500 Pg in soils worldwide (Serrano et al., 2019). However, the carbon stored in coastal wetlands plays an essential role, because it is stored over a longer timescale than in most terrestrial soils. C storage in upland soils is limited by the high availability of oxygen, allowing for aerobic microbial carbon oxidation and its subsequent release back into the atmosphere. In contrast, the soil in coastal wetlands is saturated with water, keeping it in an anaerobic state, resulting in potential storage of carbon (Chmura et al., 2003), provided there is sufficient accommodation space for vertical sediment accretion and sediment availability (Fagherazzi et al., 2012; Amann et al., 2024). For example, in northern New England saltmarsh sediments are 3–5 meters thick, and are composed of up to 40% organic carbon (Johnson et al., 2007). However, the overall global amount of C stocks is only known with a large margin of uncertainty, mostly because of the high spatial variability of C stocks (McLeod et al., 2011). On average, the global C budget was imbalanced by -0.3 GtC yr^{-1} over the period 2011–2020 (Friedlingstein et al., 2022). A better estimate of C sequestration from coastal wetlands would help to rebalance the C budget, thereby reducing the missing C.

The proportional contribution of soil carbon to aboveground living biomass is often higher than 98% for tidal salt marshes (Johnson et al., 2016) and seagrasses (Fourqurean et al., 2012). It is also suspected to be as important for microalgal biofilms inhabiting mud and sandflats (i.e. microphytobenthos) although their actual contribution remains unknown. These ecosystems are highly productive, with high rates of photosynthesis, but they also release methane (CH_4) and carbon dioxide (CO_2) through aerobic and anaerobic mineralisation processes in soils that can offset part of their C sequestration capacity (Alongi, 2020; Yau et al., 2023).

The relative contribution of autochthonous C produced within the ecosystem and allochthonous C imported from external sources vary substantially among blue carbon ecosystems. Significant allochthonous contributions have been reported in salt marshes and seagrass meadows, derived from adjacent terrestrial or marine ecosystems. For example, marine macroalgae and microphytobenthos are key primary producers in coastal ecosystems (Bordeyne et al., 2017), but they are often overlooked in blue carbon inventories. Large quantities of macroalgal detritus, green algae and biofilms may accumulate on intertidal marshes or mud flats (Krause-Jensen and Duarte, 2016; Ortega et al., 2019). If most of the autochthonous and allochthonous organic C is ultimately remineralised back to CO_2 , its contribution to long-term net C sequestration may be limited. However, total alkalinity produced by organic matter mineralisation may contribute to long-term C storage in the ocean if at least a fraction



of the metabolic products (e.g., reduced Fe and S) is permanently removed by through burial (Burdige, 2007; Maher et al., 2013; Reithmaier et al., 2023). Yet, the transformation pathways and ultimate fate of C stored in wetland soil, as well as the role of pore water exchange in C transport to adjacent coastal waters remain poorly constrained (Perkins et al., 2022).

- In particular, the tidal cycle and the activity of benthic organisms promote the export dissolved carbon species into the ocean (Krauss et al., 2022). The export of total alkalinity represents sequestered carbon over very long periods of time, with a residence time up to a millennium. By contrast, some dissolved inorganic carbon can be rapidly released as CO₂ into the atmosphere, while some portion can be sequestered over longer timescales (Reithmaier et al., 2023). In some coastal ecosystems, the processes of calcification, associated to shell formation, can consume large amounts of alkalinity, leading to significant CO₂ emissions that can offset the carbon sequestration (Van Dam et al., 2021).
- Local C density, net CO₂ sequestration, CH₄ emissions, alkalinity production and its export to the ocean are controlled by a wide range of environmental and ecological factors including salinity, nutrient loading, sediment supply, climate, species composition, and tidal range. Therefore, it is important to accurately measure and understand what factors regulate C cycling processes in order to better characterising these ecosystem functioning, calculate robust C budgets and predicting how these systems may respond to climate change.
- The study aimed to quantify production, fluxes and export of alkalinity newly produced from the sediments of an intertidal salt marsh. These fluxes were compared with the sedimentary blue carbon burial in order to assess CO₂ sequestration pathways by this type of environment. This study contributes to three research questions raised by Santos et al. (2021): (1) how comparable the different methods used to estimate physical transport and carbon outwelling are, (2) what the main natural drivers of carbon outwelling are, and (3) what the contribution of TA (as bicarbonate) makes to blue carbon budgets.

2 Material and Methods

2.1 Study Site

- The Arcachon Bay is a semi-enclosed lagoon located on the west coast of the Nouvelle-Aquitaine region, France. The area is subject to a tidal range between 1 and 4.9 meters. The Arès and Lège-Cap-Ferret Salt Marshes National Nature Reserve, the study site, is located north of the Arcachon Bay (Fig. 1). The 3.3 km² reserve is predominantly characterised by a 2.0 km² vegetated salt marshes area. The vegetation forms a continuous gradient from the upper levels, which are dominated by *Halimione-juncetum maritimae* and *Halimione portulacoides*, down to the lower levels, which are characterised by *Spartina spp.* and *Salicornietum spp.* and, further down, by bare mud areas. This shift reflects an increase in the frequency of tidal inundation, with the transition from areas flooded only during spring tides to areas drained twice a day after each tide (Brun, 2011). The upstream region is characterised by the presence of an artificially constructed canal dating back to the 19th century. This canal provides a continuous source of freshwater, and acts as a freshwater endmember within the ecosystem. The canal intersects the reserve and contributes 15% of the freshwater inflows to the Arcachon Bay (Blanchet et al., 2005). The main freshwater source to the bay is the Leyre river, the mouth of which is located in the southern part of the bay.

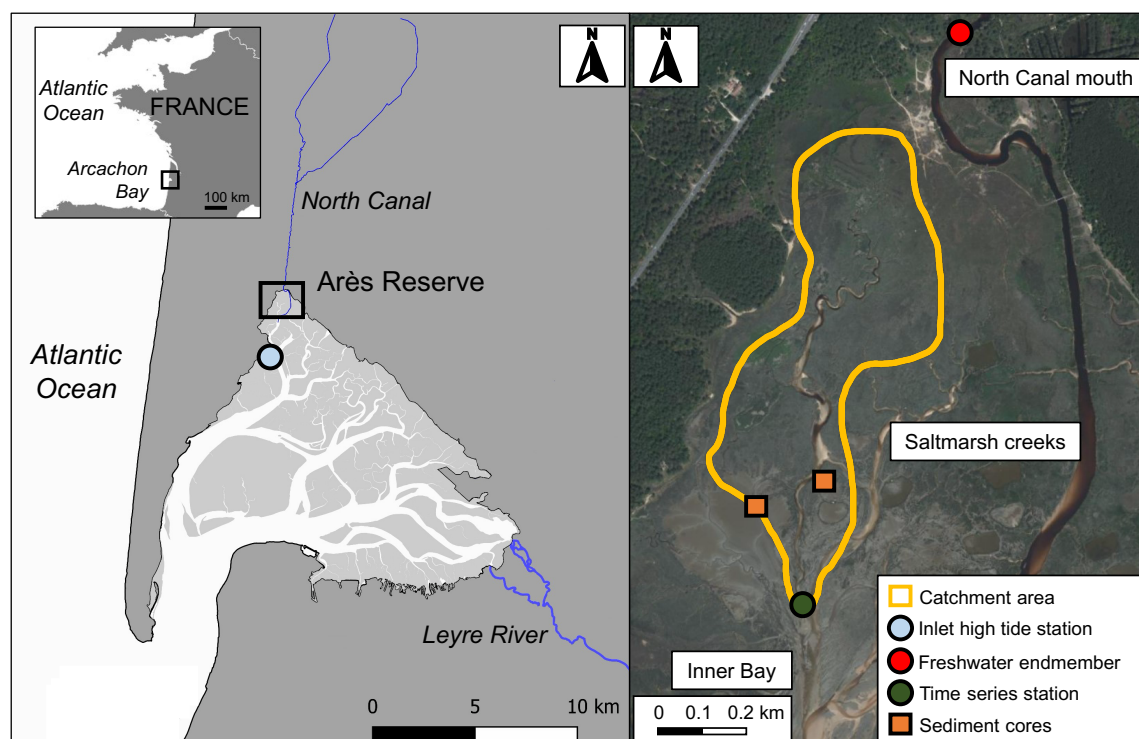


Figure 1 : Map and satellite view of the study site. The light and dark grey areas correspond respectively to intertidal zones and land. The blue lines are the main freshwater input in the Bay, the Leyre River in the south, and the North Canal, which is directly connected to the Arès Reserve. The basemap was designed by OpenStreetMap contributors, (2017), and the aerial photograph by Carte.gouv.fr. For more information on this photograph visit [https://cartes.gouv.fr/explorer-les-cartes/?c=-1.166708,44.768080&z=16&l=ORTHOIMAGERY.ORTHOPHOTOS&GEOPORTAIL:OGC:WMTS\(1;1;1;0\)&permalink=yes](https://cartes.gouv.fr/explorer-les-cartes/?c=-1.166708,44.768080&z=16&l=ORTHOIMAGERY.ORTHOPHOTOS&GEOPORTAIL:OGC:WMTS(1;1;1;0)&permalink=yes)

2.2 Water Sampling

The sampling focused on a tidal creek that drains water during low tide. This tidal creek runs from north to south across the salt marsh and drains 30 hectares of the intertidal zone. It passes through the different areas of vegetation. The creek is directly accessible on foot. The creek is not connected to the canal that brings fresh water from upstream. A total of 12 diurnal ebb tide time series were conducted in the tidal creek from September 20th 2024 to September 29th 2025 (44°45'53.0"N 1°10'05.0"W). The natural variations of biogeochemical and hydrological processes were considered by targeting different tide conditions and seasons. Water samples in the tidal creek were collected over a period of about six hours bracketing low tide. The first sample collected was used as a reference point for water least affected by tidal pumping of pore water. This point was collected as soon as the creek was accessible, which corresponded to the moment when the mudflat begun to be exposed. Due to the site elevation, this time coincided with the middle of the Arcachon Bay ebb tide. Subsequently, sampling was performed at discrete intervals ranging from 25 to 60 minutes over a 6 hour period, until the current reversed direction with the arrival of the incoming tide. An actual high-tide sample was also collected during several sampling campaigns. This sampling could not be carried out at exactly the same location, as it would have required a boat. Instead, it was conducted on a beach accessible at high tide



located 1.5 km south of the creek sampling point. The freshwater endmember (canal) was sampled once on each sampling date.

For all samples, water was collected using a bucket rinsed with in situ water. Temperature, pH, salinity, and dissolved oxygen saturation were recorded using a WTW multi-parametric probe (Multi 3630 IDS, Xylem Inc., Germany) that was calibrated prior to each sampling campaign. The Sentix® 940 pH probe was calibrated using WTW buffer solutions. Salinity probe (TetraCond® 925) was calibrated with 1413µS/cm Hanna Conductivity buffer and OSIL Atlantic Seawater was used as a standard to verify the calibration ($S_{\text{measured}} = 35.0$). The FDO®925 oxygen probe was calibrated in the calibration cell before every measurement. When appropriate, oxygen saturation values were corrected for the influence of salinity. These probes had a precision of ± 0.2 °C for temperature, ± 0.004 for pH, ± 0.1 for salinity and ± 0.5 % for oxygen saturation.

Water samples were collected from the bucket using a 50 ml syringe and immediately filtered through a 0.2 µm cellulose acetate syringe-filter (Sartorius Minisart®). Then, 15 mL were frozen and 15mL were acidified using 20 µL of 37 % HCl respectively for nutrients (DIN) and dissolved ions analysis. Water samples for TA measurements were sampled using a 50 mL plastic syringe, filtered through the 0.2 µm cellulose acetate syringe-membrane into 150 mL polypropylene bottles. Waters samples for DOC analysis were collected with a clean 50 mL glass syringe, filtered through pre-combusted (550°C) GF/F filters (porosity of 0.7 µm) into detergent-washed and pre-combusted Pyrex vials (30 mL), acidified with 50 µL of a 37 % HCl solution to pH 2, and stored at 4 °C. Methane water sample was transferred into a 100 mL glass bottle with a gas-tight septum, immediately sealed and poisoned with 300 µL of HgCl₂ until 17/12/2024 and 300 µL of 10 % Benzalkonium Chloride (Osaka et al., 2024) after this date. The second poison is less dangerous to manipulate than HgCl₂. Poisoning comparison was conducted resulting in same values with the same reliability for the two methods. We collected samples to analyse $\delta^{13}\text{C}$ -DIC for 6 out of the 18 sampling dates. Water was transferred into 200 mL glass bottles with a gas-tight septum, immediately sealed and poisoned with a saturated HgCl₂ solution.

2.3 Discharge measurements

For each time series, discharge measurements were performed alternating with water sampling. Velocities were carried out with an OTT MF Pro electromagnetic current meter equipped with a depth probe. The device provides values with an accuracy of 0.001 m s^{-1} at a time step of 0.2 s. Velocity measurements were carried out along vertical profiles separated from each other by 50 cm along the 4 to 8 m wide creek section. The current measurements were taken at the bottom, a few cm from the bottom, then every 5 cm up to the surface. Water depth ranged from 2 to 50 cm, depending on tide and the position along the section. The probe was placed at each point for several seconds until the integrated value was stable. In total, between 17 and 115 points were measured per section within 3 to 10 minutes, depending on the shape and depth of the creek. The current velocity and depth values were recorded manually and entered on a spreadsheet for processing. In view of the variations in current, the accuracy of the speed measurements was 0.02 m s^{-1} , representing a relative error of 5 % for high discharge and 20 % for low discharge.



150 2.4 Core sampling and open incubations

In May and September/October 2025, sediment cores were sampled from both bare mud and vegetated zones close to the tidal creek. For both sampling periods, 40 cm long cores were collected at each station, sliced and freeze-dried for organic carbon concentration and sedimentation rates determination.

Open incubations were conducted using an additional set of four cores for each date and site. Sediment cores were collected
155 in 9.6 cm diameter polycarbonate tubes to measure benthic fluxes across the sediment-water interface during simulated immersion. In the laboratory, the upper 8 cm of each tube were filled with 580 mL of in situ waters collected at high tide on the same day. The cores were disposed placed under dark condition in temperature-controlled regulated incubators adjusted set to the water temperature measured at high tide. One control tube containing only high-tide water was incubated under the same conditions. The overlying water was continuously stirred (at 30 rpm) using a Teflon-coated magnetic stir bars. Cores
160 were allowed to stabilize at least 6 hours before the first sampling. Then small, volumes of overlaying water volumes (15 mL) were collected every 24 h for 4 days (T_0 to T_4) for TA and dissolved nutrients. Temperature, pH, salinity, and dissolved oxygen were monitored with the same WTW probes as those already used in the field.

2.5 Analyses

Frozen samples were used to measure dissolved inorganic nitrogen (DIN) compounds including ammonium, nitrite and nitrate
165 ($\text{DIN} = \text{NH}_4^+ + \text{NO}_2^- + \text{NO}_3^-$). NO_2^- and NO_3^- were analysed according to (Schnetger & Lehnert, 2014). NH_4^+ was measured by colourimetry with a reaction waiting time > 8 hours (Koroleff, 1976b). Dissolved iron (Fed), dissolved manganese (Mnd) and sulphates (SO_4^{2-}) were analysed on the acidified sample to prevent dissolved Fe(II) oxidative precipitation and subsequent post-sampling phosphate adsorption on Fe(III) phases. Fed was analysed by colourimetry by adding a ferrozine solution in an aliquot (Stookey, 1970). Mnd was determined colourimetrically using the Cd-TCPP complex (Madison et al., 2011;
170 Charbonnier and Anschutz, 2019). DIP was measured by colourimetry in 5 cm long cells (Murphy and Riley, 1962). Measurements for sodium, calcium and sulphur interpreted as sulphate following Metzger et al. (2007) were performed by an ICP-OES ICAP 6300 Thermo-Fischer. Acidified samples were diluted 1:50 with 1 % HCl. Calibration was performed with seven multi-element standards, covering the concentration range of the samples, prepared by properly diluting single-element stock solutions ($1000 \mu\text{g g}^{-1}$). All standards and the blank were measured at the beginning, midpoint, and end of the analytical
175 run; three replicate readings were performed on standards, blank and samples. Sulphate was also analysed by nephelometry in water samples by analytical triplicates (Rodier, 1976). Accuracy was controlled with standard sea water. All analysis has an accuracy from 1 to 8 % according b the references.

Total alkalinity of tidal creek water samples was analysed on 50 mL sub-sample by automated potentiometric titration with 0.1N HCl using a Metrohm 888 Titrand equipped with an UnitrodePlus probe controlled with Tiamo 3.0 software. The
180 equivalence point was determined using the inflection point of the titration curve, identified from its first derivative. Analytical precision was better than $5 \mu\text{M}$. For open incubations, the small volumes available ($< 5 \text{ mL}$) were insufficient to measure TA



by automated potentiometric titration. These TA samples were analysed by manual potentiometric titration using same equipment and 0.01N HCl as titrant. A 1 mL aliquot of sampled water was diluted in 20 mL Milli-Q water prior to titration. Precision was better than 10 μ M according to replicate samples. Calibration was performed using an internal 2 mM NaHCO₃ standard.

Dissolved organic carbon (DOC) concentrations were determined by the high temperature catalytic oxidation method using a Shimadzu TOC-L analyser after removing DIC by bubbling oxygen through the acidified samples (Sharp, 1993). The instrument was calibrated using potassium phthalate standards solution covering the concentrations range expected for the samples. The precision of these measurements was about 10 μ M, based on replicate analyses of the same samples.

Dissolved CH₄ was analysed by gas chromatography using a Shimadzu GC-2010 Pro. A headspace was created in the vials by injecting N₂ and the volume of water and gas in each vial was determined gravimetrically. After equilibration between the liquid and gas phases for a several hours at ambient temperature, duplicate gas samples were injected in duplicate into the Gas Chromatograph. Gas standards (Messer, pCH₄ = 10.04 ppm) were used for calibration. In situ CH₄ concentrations were calculated using CH₄ solubility coefficient as a function of salinity and temperature (Yamamoto et al., 1976).

$\delta^{13}\text{C}$ -DIC measurements were performed according to the method described by Gillikin and Bouillon, (2007). Briefly, a headspace was created in sealed 100 mL glass bottles by injecting a volume of helium equivalent to approximately 30 % of the total vial volume. Samples were then acidified with 0.3 mL of 85 % phosphoric acid to convert all DIC species in CO₂. After shaking, the samples were allowed to equilibrate over-night at ambient room temperature of the mass spectrometer laboratory. They were subsequently analysed using a dual inlet EA-IRMS (ThermoFisher Delta Q + Flash 2000). About 0.5 mL of headspace gas was injected through an injection port that was located upstream of the EA water trap. The $\delta^{13}\text{C}$ values were calibrated with a homemade standard prepared by reacting 45 mg of Na₂CO₃ dissolved with 3 mL of 85 % phosphoric acid in a 100 mL helium flushed vial. Its isotopic composition had previously been determined as $4.5 \pm 0.2\text{‰}$ (Polsenaere et al., 2013b). The $\delta^{13}\text{C}$ -DIC were computed from these corrected $\delta^{13}\text{C}$ values using the equation of Miyajima et al., (1995), which accounts from the partitioning of CO₂ between the headspace and the water phase of the sample as a function of temperature and salinity. Analytical precision was estimated at approximately $\pm 0.1\text{‰}$, based on replicate analyses of the same samples.

Particulate Organic Carbon (POC) was measured in the freeze-dried sediments by infrared spectroscopy using an Eltra ELEMENTRAC CS-i after carbonate removal with 2 N HCl (Etcheber et al., 1999). Radionuclide activities (⁷Be, ²¹⁰Pb, ²²⁶Ra) were measured using a high efficiency, broad energy gamma detector equipped with a Cryo-Cycle II (@Mirion). The gamma detector was calibrated using IAEA certified materials (RGU1). Errors on activities were based on standard deviation counting statistics. Excess ²¹⁰Pb (²¹⁰Pb_{xs}) was calculated by subtracting the parent isotope ²²⁶Ra from the total ²¹⁰Pb, both of which were measured in the sediments.



2.6 Calculations

Oxygen concentration (μM) was deduced from oxygen saturation with correction of salinity, temperature and pressure influences using empirical equations and hourly measured atmospheric pressure (Garcia and Gordon 1992). From laboratory TA measurements, and in situ salinity, temperature and pH, whole carbonate system (DIC , pCO_2 , HCO_3^- , CO_3^{2-}) was calculated using (Millero et al., 2006) equations the dissociation constants K_1 and K_2 for carbonic acid from (Mehrbach et al, 1973), readjusted by (Dickson and Millero, 1987), and the solubility of CO_2 (K_0) from (Weiss, 1974).

For each sampling date, the salinity (S) and the concentration in a given species (C) of the freshwater (A) and high tide (B) endmembers were used in a mixing model to calculate the influence of tidal pumping on the water chemistry of tidal creek samples (e.g. Zhang et al., 2026) The model assumes a linear relationship between salinity and the concentration of each dissolved species, as defined by the two endmembers. The salinity measured in each sample (S_{meas}) was used to calculate the theoretical concentration expected from conservative mixing (C_{theo}), which was then compared with the measured values (see Eq. 1). The differences from conservative mixing were defined as excesses when $(C_{\text{meas}} - C_{\text{theo}}) > 0$, as deficits $(C_{\text{meas}} - C_{\text{theo}}) < 0$, and consistent with the conservative when $(C_{\text{meas}} - C_{\text{theo}}) = 0$, within the analytical precision.

$$C_{\text{theo}}(\text{at } S_{\text{meas}}) = \frac{S_{\text{meas}} - S_A}{S_B - S_A} (C_B - C_A) + C_A \quad (\text{Eq. 1})$$

The theoretical concentrations of calcium and sulphate concentrations were calculated using dissolved Na (Nad) rather than salinity. Indeed, C_{theo} could be estimate more precisely because Cad , Nad and elemental S from dissolved sulphate were measured in the same aliquot by ICP-AES.

To calculate discharge, the weighted average of current velocity was calculated for each vertical profile. For that, the average velocity (v) between two successive points (i) was attributed to the proportion represented by the vertical distance (p) of these two points relative to the total height of the profile (h). This current profile was assigned to the cross-section area (A) represented by the trapezoid limited by the bottom, the surface water and the mid-distance between the profile located at a distance (d_n) from the edge and its neighbours or the edge of the stream bed (Eq. 3). The total discharge (Q , Eq. 4) was calculated as the sum of the flows (q_n) through the n trapezoidal sections.

$$q_n = [A] \times \sum \left[\left(\frac{v_i + v_{i+1}}{2} \right) \times \frac{(p_i - p_{i+1})}{h_n} \right] \quad (\text{Eq. 2})$$

$$A = \frac{\left[\left(\frac{d_{n+1} - d_n}{2} \right) - \left(\frac{d_n - d_{n-1}}{2} \right) \right] \times \left[\left(\frac{h_n + h_{n-1}}{2} \right) - \left(\frac{h_{n+1} + h_n}{2} \right) \right]}{2} \quad (\text{Eq. 3})$$

$$Q = \sum q_n \quad (\text{Eq. 4})$$

Fluxes of dissolved compounds were calculated by multiplying the tidal creek discharge by the measured concentration (gross fluxes), or by the excess concentrations relative the mixing line (net fluxes). Because water samples were collected was at t ,



245 t+30 min, t+60 min, etc., whereas discharge measurements at t+15 min, t+45 min, etc., the mean value of two successive discharge measurement was multiplied by the concentration at the intermediate sampling time.

The estimate of CO₂ and CH₄ degassing rate at the water-atmosphere interface aw estimating from the concentration gradient between the dissolved gas concentration and the equilibrium concentration at the atmosphere [gas]_{sat}, together with the gas transfer rate (k):

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$$Flux = k * ([gas]_{water} - [gas]_{sat}) \quad (Eq. 5)$$

Numerous empirical formulations for estimating the gas transfer rate, because the geometry of a tidal creek is more similar to that of a stream than to that of a lentic water body or the open ocean; we selected formulations designed for rivers to estimate k in the tidal creek. In rivers, value of k (m d⁻¹) depends on temperature and on several parameters related to water turbulence, including current velocity, slope, and water depth. Values of k were estimated using empirical formulas that into account for these parameters (see Appendix A). The creek gas flux (in mol m⁻² h⁻¹) was then calculated on the assumption that, on average, a parcel of water travels an average of 500 meters through the creek before reaching the sampling site.

To separate the biogeochemical addition of DIC from physical mixing during tidal exposure, we built a two-endmember conservative mixing model between high tide creek water and the dominant freshwater source, using salinity as the conservative tracer for DIC and δ¹³C-DIC values (Eq. 6) in the same way than Eq.1. Because δ¹³C is a ratio rather than a conservative quantity, the isotopic composition of this excess pool cannot be obtained by simple subtraction; it was instead calculated by isotopic mass balance:

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$$\Delta\delta^{13}C = \frac{(DIC_{meas} \times \delta^{13}C_{meas} - DIC_{theo} \times \delta^{13}C_{theo})}{\Delta DIC} \quad (Eq. 6)$$

The isotopic signature of the additional carbon source was then estimated from the excess DIC pool using two complementary approaches: a Keeling-like plot (Δδ¹³C vs. 1/ΔDIC; Pataki et al., 2003), in which the intercept represents δ¹³C signature of the source; a Miller-Tans plot (Δδ¹³C × ΔDIC vs. ΔDIC; Miller and Tans, 2003), in which the slope represents the δ¹³C signature source). For the latter Miller-Tans approach, the regression was also forced through the intercept to 0 since ΔDIC is zero by construction of excess carbon.

An objective was to calculate sediment and mass accumulation rates (SAR and MAR, respectively) from the exponential curve that usually describes the decrease of ²¹⁰Pb_{xs} (T_{1/2} = 22.3 years) with depth in sediment and cumulative mass, respectively, using the CF:CS (constant flux and constant sedimentation) model. Carbon accumulation rates (CAR), were calculated as the product of the POC content where the profile was stable at lower sections and MAR derived from ²¹⁰Pb_{xs} profiles.

275 Benthic fluxes (μmol m⁻² h⁻¹) from open incubations were calculated with the slope of the linear regression of the stock against time. Stocks were defined as the total number of moles of a given dissolved species in the overlying water. At each sampling



time, it was calculated by multiplying the measured concentration by volume of overlaying water obtained by the water height and the surface area of the core tube.

To test the correlation between the measured variables and the excess/deficit over time during the series of low tides, the non-parametric Spearman correlation was used with the “R” stats package (version 4.3.1; R Core Team, 2023). Kruskal-Wallis test was also used with the same package to evaluate the significance of the difference of organic carbon content in sediment cores.

3 Results

3.1 Endmember characteristics

The waters sampled in the tidal creek were located along a mixing continuum between the freshwater canal endmember and marine waters from Arcachon bay (Fig. 1). The canal endmember was characterised by salinities between 0 and 4.4 (Fig. 2). The highest values were observed during periods of high spring tides and low canal discharge in summer and early autumn. During these periods, the lower section of the canal was influenced by the tide, resulting in partial mixing between freshwater and marine water. Dissolved oxygen was always undersaturated ($85 \pm 9\%$). The pH of the canal water samples was 6.51 ± 0.23 , while TA in the least saline samples ranged from 260 to 400 μM . The pCO_2 values calculated from these parameters and temperature were between 1200 and 12000 ppm. The $\delta^{13}\text{C}$ -DIC was only measured in 5 samples and averaged $-16 \pm 1\%$. Freshwater DOC and CH_4 concentrations averaged $697 \pm 112\ \mu\text{M}$ and $2.7 \pm 1.5\ \mu\text{M}$, respectively. Nitrate concentrations ranged between 7 and 21 μM in winter and spring when canal discharges were highest, but decreased close to the detection limit (0.2 μM) in summer. Ammonium concentrations remained low, between 5 and 21 μM . Dissolved iron had a mean concentration of $13 \pm 5\ \mu\text{M}$.

The salinity of samples collected at high tide on a beach located to the south of the tidal pumping monitoring station ranged from 8.5 to 32.9. The lowest values were related to the winter high freshwater discharge of rivers and canals flowing into the Arcachon lagoon. TA increased linearly with salinity from 815 to 2265 μM following the equation $\text{TA} = 60\ \text{S} + 353$. The intercept was roughly consistent with the TA of freshwater endmember of the canal. Dissolved oxygen (O_2) concentrations were slightly lower than those measured in the freshwater endmember, with a mean O_2 saturation of $91 \pm 6\%$. The average pH was 7.6 ± 0.1 . The pCO_2 was close to the atmospheric equilibrium, with values ranging from 360 to 685 ppm. The $\delta^{13}\text{C}$ -DIC measured in two samples was $-0.6 \pm 0.1\%$. DOC concentrations averaged $300 \pm 112\ \mu\text{M}$ and varied with salinity. The highest salinity waters, collected in summer and autumn had DOC concentrations between 185 and 300 μM . CH_4 concentrations were much lower than those measured in the freshwater endmember, averaging $0.1 \pm 0.05\ \mu\text{M}$. Ammonium concentrations remained low, below 9 μM for all the sampling dates. Nitrate concentrations showed a marked seasonal variation, ranging from 0 μM in summer to 21 μM in winter and spring. These nitrate concentrations were similar to those in the canal. The high nitrate concentrations in winter were probably due to the influence of the nitrate-rich Leyre River, the main freshwater source to Arcachon lagoon (Canton et al., 2012). Except in the least saline sample, dissolved iron concentrations were always below 1 μM .



3.2 Ebb tides series observations

310 Measurements carried out in the tidal creek showed a significant variability in the measured parameters at both seasonal and tidal timescales (Fig. 2). Discharge dynamics (Fig. 2.f) followed a recurrent pattern throughout the monitoring periods with maximum values occurring at the start of the ebbing tide (from 0.7 to 2.3 m³ s⁻¹). Minimum values were always observed at the end of low tide, a relatively low background discharge between 0.02 and 0.10 m³ s⁻¹ was measured. At any given tidal phase, discharge was systematically higher during spring tides than during the neap tides discharge. The total volume of water
315 discharged through the tidal creek over the six-hour ebb and low water period ranged between from 2000 m³ during neap tides periods and 10000 m³ during spring tide periods. Over the annual cycle, water temperature varied from 3.2 to 29.7 °C according to the local seasonal climate, with a relatively mild winter conditions (> 10 °C in January) and a warm summer condition (June and July > 25 °C). Water temperatures systematically increased over the 6 hours sampling period, as all sampling campaigns started in the morning and ended in the afternoon, when air temperatures were higher. Salinity was mainly controlled by the
320 freshwater from the canal located north of the studied salt marsh. Salinity was between 18 and 31.6 during the summer and autumn whereas saltmarsh waters were more diluted by the canal water during winter and spring, with minimum salinity values below 10 in March 2024 and February 2025. During the sampling period, salinity generally decreased when the initial value was above 20, whereas it increased slightly when salinity was low at the start. Oxygen changed from undersaturation conditions (61.9 – 96.5 %) at high tide to values close or above saturation as tide ebbed. This increase could not be explained solely by
325 changes in O₂ solubility as a result of temperature and salinity variations, because the O₂ concentrations increased by 50 to 100 μM, suggesting net O₂ production. Field observations showed that the sandy surface of the creek bed became greenish during the day, consistent with microphytobenthos photosynthetic activity. The highest O₂ saturation reached 163 % during a sunny day of September 2025. Concentrations of most carbon-related parameters (TA, DIC, pCO₂, DOC, CH₄) as well as those of reduced redox sensitive species increased over the 6 hours sampling period. TA ranged between 850 and 2379 μM. Within the
330 tidal creek, the lowest TA values were measured when marine water was diluted by the low alkalinity freshwater endmember. During the sampling chronicles, TA either remained relatively stable or increased (Fig. 3.g) even though salinity decreased on most sampling dates. Similarly, DIC concentrations, which ranged from 892 to 2493 μM showed non-linear behavior with respect to salinities. The isotopic composition of DIC (δ¹³C-DIC) was measured at five periods when DIC concentrations remained relatively constant. Despite the limited changes in DIC concentrations, δ¹³C-DIC decreased from a value close to -5
335 ‰ to a value close to -10 ‰. The concentrations of CH₄ ranged between 0.17 and 4.72 μM and DOC concentrations from 316 to 1411 μM, systematically increased during the monitoring. The pCO₂, ranged between 1008 and 7038 ppm and typically increased by 2000 to 5000 ppm during the 6 hours sampling period. The concomitant decrease in pH from ~7.4 to ~6.9 was more pronounced during spring tides. Dissolved Fe (0 – 45.8 μM) and Mn (0 – 5.2 μM) concentrations increased significantly during low-tide cycles, with concentrations at the end of the period 2 to 10 times higher than the start. Ammonium
340 concentrations also tended to increase during the ebb tide period with values averaging approximately 15 μM. Nitrate

concentrations were generally below 20 μM with a decreasing trend during the low tide period. Sulphates (4538 – 24546 μM) and dissolved calcium (8 – 12 mM) concentrations followed the freshwater-seawater dilution pattern related to salinity.

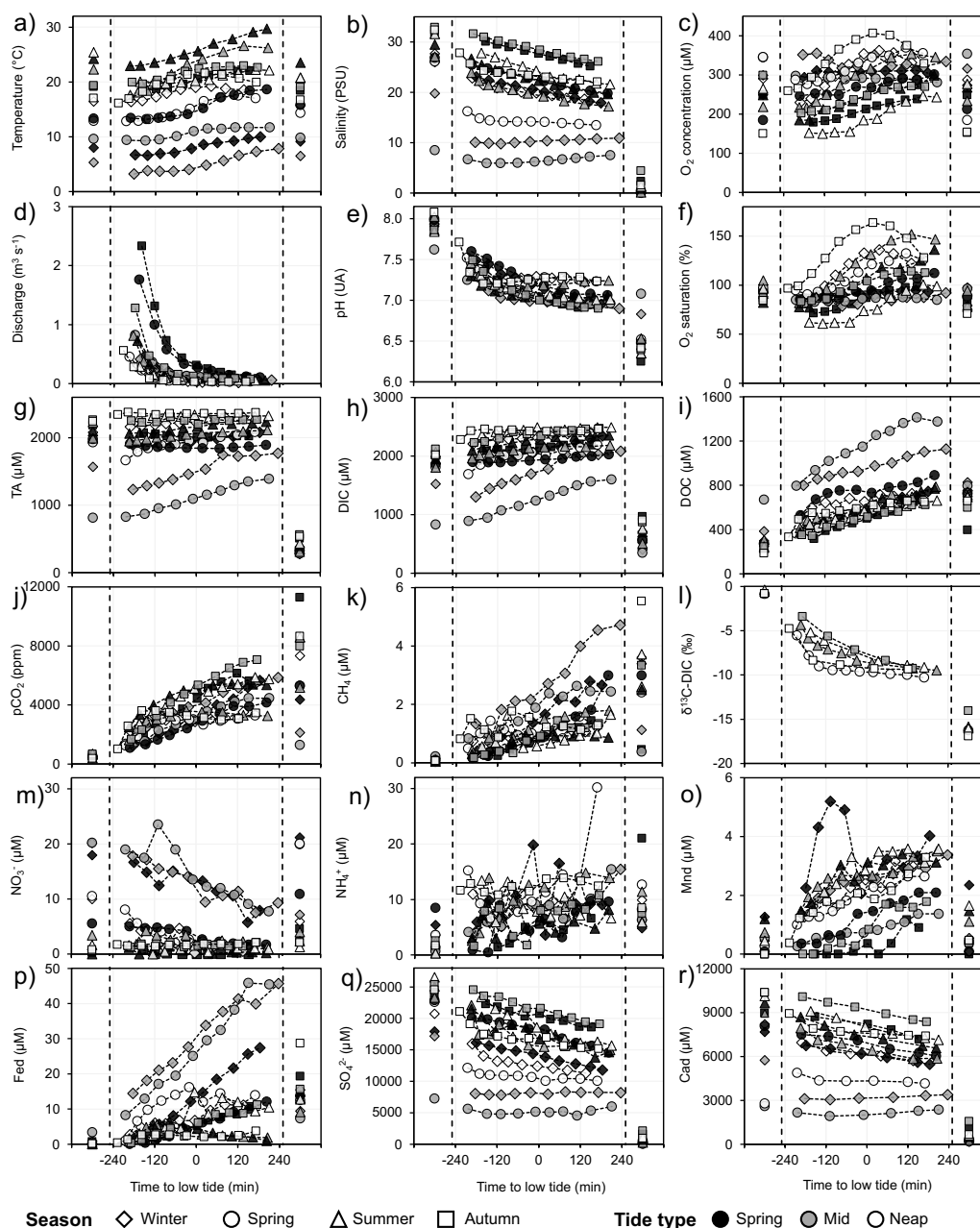


Figure 2: Evolution of parameters measured in the tidal creek during low tide periods. Time 0 corresponds to the time of low tide at the Arcachon tide gauge. The site is accessible approximately 240 minutes before low tide, and the flowing tide reaches the site approximately 240 minutes after low tide. a) temperature, b) salinity, c) oxygen concentration, d) oxygen saturation, e) pH, f) alkalinity, g) dissolved inorganic carbon, h) pCO_2 , i) dissolved organic carbon, j) methane, k) nitrates, l) dissolved manganese, m)



dissolved iron, n) sulphates, o) dissolved calcium and p) dissolved inorganic carbon isotopic composition. Concentrations are in micromoles per kg of solution (μM). Each dashed curve represents a sampling date, and each point denotes a sample. Vertical dashed lines separate on the left part external high tide point, in the middle the ebb tides series and on the right the freshwater endmember. Time scale only applies to ebb tide series in the middle between vertical lines.

The parameters related to carbon cycle did not follow a simple mixing line between the freshwater endmember and the water present at high tide at the site. For example, for TA (Fig. 2), the values measured during the low-tide monitoring periods were consistently plotted above the mixing line between freshwater endmember and the water sampled at high tide near the salt marshes (Appendix B). This deviation suggests that the water flowing through the tidal creek was influenced by an additional alkalinity enriched endmember. The alkalinity excess relative to the conservative mixing line is hereafter referred to as ΔTA . For a given salinity, ΔTA was calculated as the difference between the measured value and the value on the mixing line (Appendix B). The reference mixing line used for this calculation was defined between the freshwater endmember and the first sample collected in the tidal creek just before ebbing tide. This sample was considered as being the most representative of the *in situ* high tide water prior to any potential contribution of interstitial water through tidal pumping.

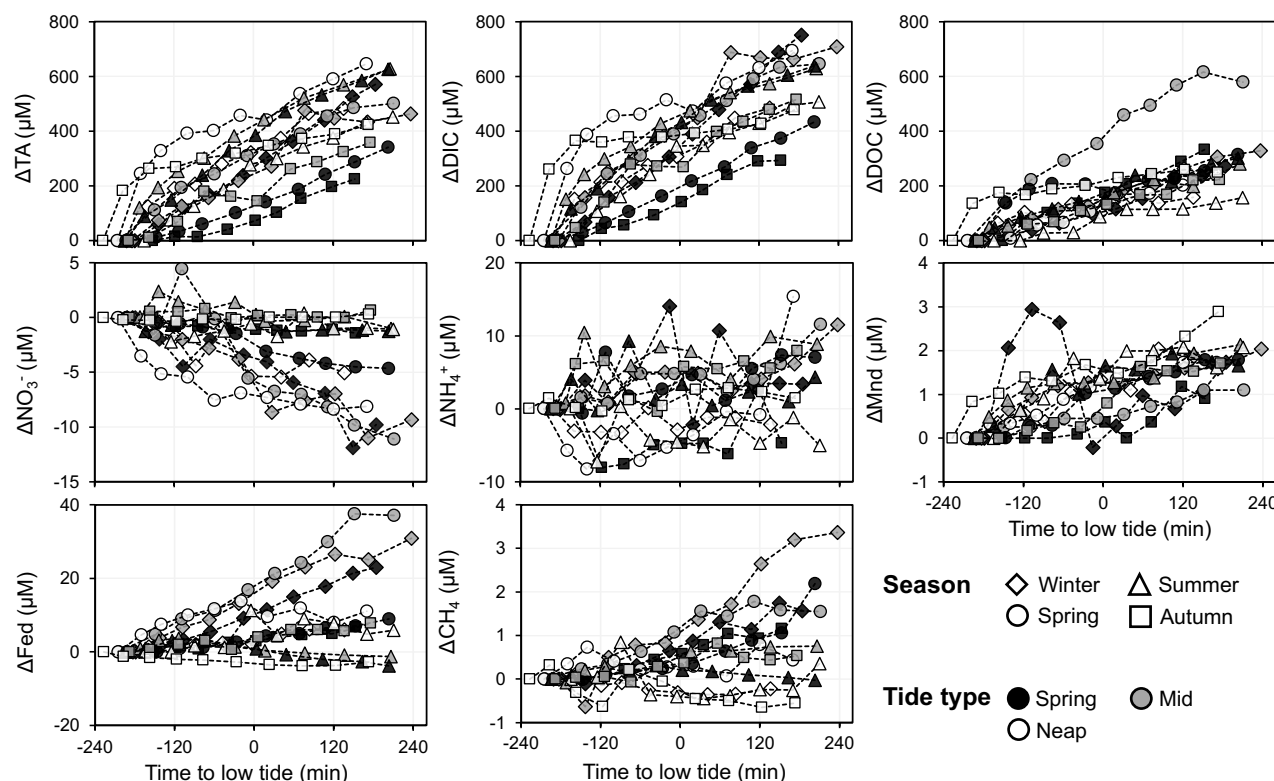


Figure 3 : Evolution of excess or deficit of biogeochemical parameters in tidal creek water during the ebb tide cycles. Concentrations are expressed as deviations from a salinity dilution model (Δ = measured value – theoretical value).



Δ TA increased by 225 to 645 μ M throughout each measurement, reaching maximum values at the end of ebb tide, just before the current reversal (Fig. 3). Δ DIC and Δ DOC followed the same trend with maximum values also reached at the end of ebb tide. Maximum Δ DIC values were 70 to 106 μ M above Δ TA values. Maximum Δ DOC value were between 156 and 340 μ M except during one campaign when the maximum value was close to 600 μ M. In all seasons, Δ TA and Δ DIC values were greater during neap tides than during spring tides. Mnd, Fed and CH_4 were also present in excess in the most campaigns. Dissolved manganese was systematically in excess during the 12 campaigns. This was also the case for 11 of the 12 campaigns for Fed and in 10 of the 12 campaigns for CH_4 . Generally, Δ Mnd, Δ Fed and ΔCH_4 increased over the tidal cycle, reaching their maximum positive values at the end of ebb tide (up to +3 μ M for Δ Mnd, +38 μ M for Δ Fed and +3 μ M for ΔCH_4). NO_3^- showed the opposite trend. Concentrations remained close to zero when both the freshwater and the high-tide endmembers were less than 2 μ M. In winter, when nitrate was present in the water column, concentrations in the tidal creek were below the conservative mixing line, with ΔNO_3^- reaching as low as -11 μ M. Ammonium concentrations fluctuated around the mixing line and showed no specific pattern. Dissolved Ca and sulphate concentrations were consistent with the conservative mixing line within analytical error (2 %). Deviations from the mixing line in the tidal creek waters ranged from a few tens of μ M up to 200 μ M. However, because sulphate concentrations ranged between 5000 and 25000 μ M and Cad concentrations between 2000 and 9000 μ M, the measured concentrations differed from those predicted by the conservative mixing equation by only 0 to 2 %. Consequently, these apparent excesses or deficits could not be quantified. Nevertheless, sulphate values were predominantly below the mixing line and the Cad values were above it, suggesting an weak addition of Ca and a net removal of sulphate.

By multiplying the concentration differences relative to the conservative mixing model by the discharge measured in the tidal channel and integrating these values over the low-tide period, we calculated the net mass flux associated with tidal pumping for each tidal cycle. The daily flux was estimated assuming 2 tides per day. Positive values indicate a net source, whereas negative values represent a net sink or a dilution. TA, DIC and DOC fluxes were always positive (Table 2). Over all campaigns, the net TA fluxes averaged $1149 \pm 412 \text{ mol d}^{-1}$ and was always lower than net DIC fluxes, which averages $1548 \pm 486 \text{ mol d}^{-1}$. Net TA and DIC fluxes were of the same order of magnitude all throughout the year. The lowest values ($< 1000 \text{ mol d}^{-1}$) occurred mainly in summer and autumn, when the high tide water salinity was the highest. The mean net DOC fluxes were $815 \pm 705 \text{ mol d}^{-1}$. Because deviations of sulphate concentrations from the value predicted by the conservative mixing line were within 1 %, and thus within the analytical uncertainty of measurements, it was impossible to provide a precise value for the sulphate deficit in the tidal creek waters. Cad concentrations were also very close to the mixing line, either directly on the mixing line or slightly above it. However, the apparent Cad excess calcium fell within the 2 % analytical uncertainty of the measurements. Net Fed and Mnd fluxes were 1 to 3 orders of magnitude lower than TA or DIC fluxes. They reached their highest values during the colder months, between December and March, when the water was the least saline. Net CH_4 fluxes were very low, with an average value of $2 \pm 2 \text{ mol d}^{-1}$. Net NO_3^- fluxes were negative or close to zero with a mean of $-10 \pm 10 \text{ mol d}^{-1}$.

Tidal creek waters always exhibited dissolved CO₂ and CH₄ concentrations above atmosphere equilibrium values. The tidal creek therefore acted as a net source of CO₂ and CH₄ to the atmosphere during all sampling campaigns (Table 1). The flux of CO₂ was between 0.5 and 5 μmol m⁻² s⁻¹ depending on waters supersaturation. Assuming a travel distance of 500 meters before the water reached the sampling point, the estimated flux of CO₂ ranged from 90 mol d⁻¹ during a neap tide in November to 304 mol d⁻¹ during a spring tide in September 2024 with a mean of 196 ± 70 mol d⁻¹. The CO₂ flux increased throughout each low-tide monitoring sequence. This increase primarily reflected the progressive rise in CO₂ concentrations, and thus the increasing concentration gradient between water and the atmosphere. The relative impact of CO₂ loss on concentrations was greater at the end of the sequence than at the beginning, because the longer residence time of water in the creek and the shallower water depths. At the beginning of the monitoring, the CO₂ loss corresponded to nearly 7 to 20 % of the dissolved CO₂. By the end of the monitoring, this proportion exceeded more than 25%, reaching up 70 %. Methane fluxes were substantially lower, than CO₂ fluxes, but they followed a similar pattern. Estimated CH₄ losses to the atmosphere ranged from 0.84 to 2.83 mol d⁻¹ with a mean value of 1.81 ± 0.56 mol d⁻¹. These fluxes represented a loss of methane ranging from 10% to 80% of the amount dissolved CH₄, depending on the tidal stage, in particular, on water discharge in the tidal creek.

Net		Water	TA	DIC	DOC	CH ₄	NO ₃ ⁻	NH ₄ ⁺	Fed	Mnd	SO ₄ ²⁻	Cad	FCO ₂	FCH ₄
Date	Tide	m ³	mol d ⁻¹											
20/9/24	Spring	8813	690	1560	2181	5.1	-18	-130	24	1.5	-869	315	304	1.85
7/11/24	Neap	1405	570	600	240	-0.3	-6	-6	6	1.8	-914	127	90	0.84
17/12/24	Spring	3039	1110	1440	480	2.7	-21	24	42	9.6	-1144	73	146	1.55
13/1/25	Mid	2805	1230	2040	750	4.5	-21	18	69	6.0	-309	58	210	2.58
12/2/25	Mid	4369	1890	2370	2310	5.4	-18	24	105	3.6	-427	-60	206	2.83
21/3/25	Neap	1629	1260	1410	180	1.5	-18	-18	27	2.4	-811	38	92	1.12
31/3/25	Spring	9342	1071	1538	882	4.2	-16	57	32	8.5	-866	380	186	1.90
14/5/25	Mid	2880	1740	1950	630	1.5	3	31	9	6.3	-438	375	186	2.49
26/6/25	Spring	3384	1560	1800	750	0.9	-6	18	6	6.0	-516	737	287	1.73
21/7/25	Neap	1929	750	900	210	0.0	-3	-12	12	4.5	-605	350	253	1.55
12/9/25	Mid	4206	980	1730	540	1.5	6	50	21	3.6	29	1401	272	1.50
29/9/25	Neap	1808	960	1260	630	-0.3	0	4	-6	4.5	-680	361	124	1.76
mean			1149	1548	815	2.2	-10	5	29	4.9	-629	349	196	1.81
± SD			412	486	705	2.1	10	48	31	2.5	319	416	70	0.56

Table 2 : Net fluxes of carbon and dissolved species into surface water of the tidal creek draining the salt marsh at low tide calculated using the saltmarsh high tide point. A positive value indicates a local export whilst a negative value indicates a local sink. FCO₂ and FCH₄ are creek water-air fluxes.



3.3 Benthic fluxes and OC burial

Benthic fluxes at the sediment water interface measured by using open sediment incubations, represent fluxes occurring during at high tide, when the mudflat is covered by water. During spring tides, the mudflat is submerged for about 4 hours per cycle. During neap tides, the flat is not submerged. On an annual basis, the average elevation of the tidal flat means that it is covered for about 3 hours per tide, corresponding to an average inundation time of 6 hours per day. The watershed of the studied creek comprises 2 ha of bare mudflats and 28 ha of vegetated sediment. Benthic TA fluxes were measured in May and in October using sediment cores collected in both vegetated and bare mud (Fig. 4). Fluxes from vegetated sediment showed much higher variability than those from bare mud sediments, with values between 0 and 0.70 mmol m⁻² h⁻¹ for vegetated sediments and between 0.09 and 0.36 mmol m⁻² h⁻¹ for bare sediments. Overall mean fluxes were 0.18 ± 0.09 mmol m⁻² h⁻¹ (n=8), and 0.29 ± 0.29 mmol m⁻² h⁻¹ (n=8) for bare and vegetated sediments, respectively. At the scale of the tidal creek watershed, these values corresponded to a mean benthic alkalinity flux of 509 mol d⁻¹, which was approximately half the mean advective TA flux from tidal pumping. Benthic DIC fluxes were of similar magnitude to the TA flux values (Fig. 4). A few values were slightly lower, likely due to CO₂ degassing from the overlaying water during the sediment core incubation.

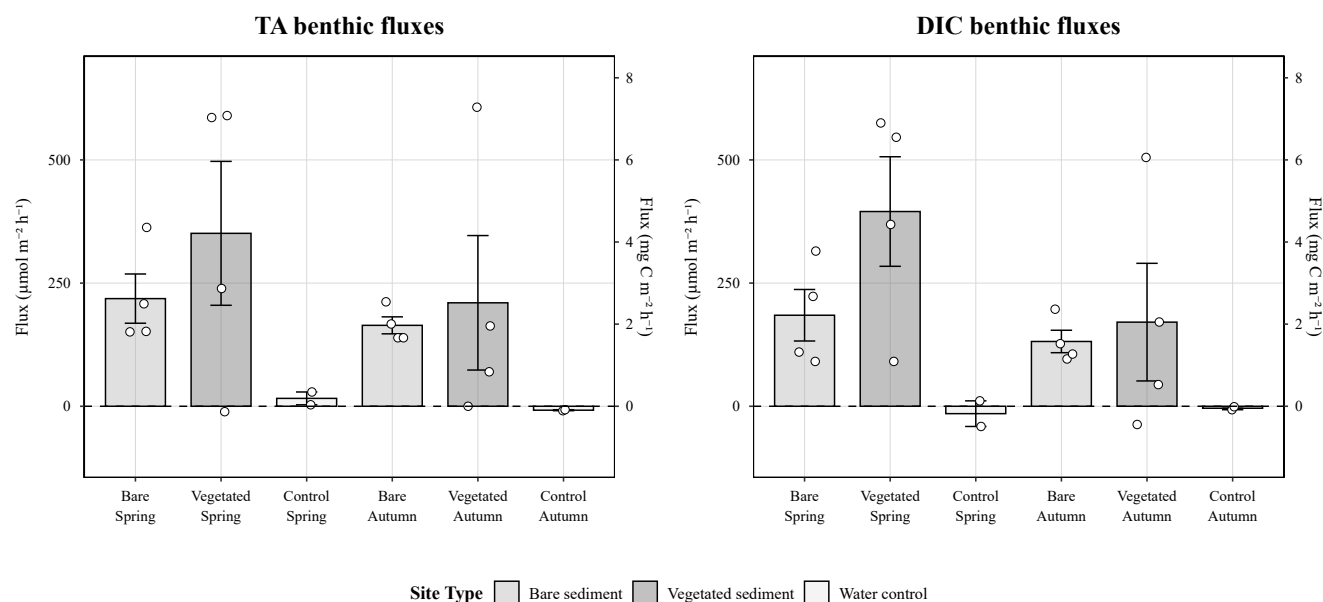


Figure 4 : TA and DIC benthic fluxes. Each point represents an incubated core, bar plot top symbolizes mean and bars SE of benthic fluxes (μmol m⁻² h⁻¹). Data are grouped by season and sediment type.

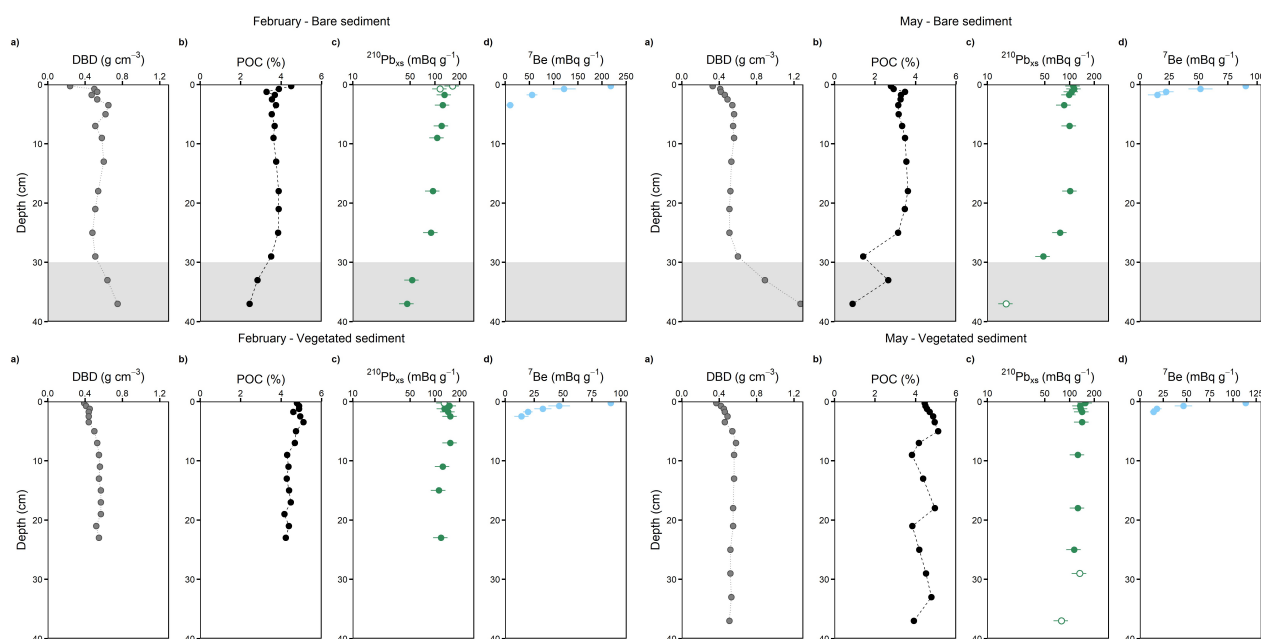
The particulate organic carbon (POC) content in bare sediments was between 1.4 and 4.5 %, with a mean concentration of 3.4 % (Fig. 5). The POC was lower below the sandy layer (25 cm depth) when present in bare muds. Carbon content in vegetated sediments was between 3.8 and 5.1 %, with a mean of 4.5 %. This difference of POC content between sediment type was



statistically significant: KW tests showed that POC was significantly higher in vegetated than in bare sediments ($p < 0.001$), independently of season ($p = 0.10$). The ^7Be activities ranged from 91 to 218 mBq g^{-1} in surface sediments (Fig. 5) and reached negligible values at depths ranging between about 3 to 8 cm. Considering the short half-live of ^7Be , this suggests important mixing of the sedimentary column.

445 $^{210}\text{Pb}_{\text{xs}}$ activities in bare muds and vegetated cores were between 130 and 164 mBq g^{-1} in the uppermost layer. Values were lower at the bottom of cores, but this is mainly related to dilution by sand and not to time. The rest of the profiles shows almost constant values, confirming the occurrence of a strong mixing of sediments as suggested from ^7Be profiles (Arias-Ortiz et al., 2018). The strong mixing and the limited decrease of $^{210}\text{Pb}_{\text{xs}}$ with depth implies to consider bioturbation when calculating sedimentation rates. Therefore, we applied the equation that combines bioturbation and sediment rates to the layers below the

450 ^7Be penetration and using the bioturbation rates obtained from ^7Be profiles (Smoak and Patchineelam, 1999; Isla and DeMaster, 2021). Values of bioturbation mixing were between 2.4 and 15.3 $\text{cm}^2 \text{yr}^{-1}$, and mean SAR values were 16 mm yr^{-1} in vegetated sediment, and 4 mm yr^{-1} in bare sediments. The respective carbon accumulation rates were 355 and 84 $\text{g m}^{-2} \text{y}^{-1}$.



455 **Figure 5 : Depth profiles of a) dry bulk density (DBD), b) particulate organic carbon (POC), c) $^{210}\text{Pb}_{\text{xs}}$ and d) ^7Be activity for sediment cores collected in Arès salt marshes in February (left) and May (right) 2025 in bare (top) and vegetated (bottom) sediments. The shaded area indicates the position of sandy mud layer. Error bars correspond to 1 SD.**

4. Discussion

The data presented here were based on tidal creek monitoring, incubation experiments, geochemical tracers and sediment core analyses. They provide new insights on TA and dissolved carbon fluxes in a tidal creek draining a salt marsh in the Arcachon



460 lagoon. The one-year time-series allowed us to obtain results across different seasons tidal amplitudes. This approach allowed
us to quantify the relative contributions of dissolved carbon outwelling, outgassing, organic carbon burial, and alkalinity export
to the net carbon budget. Our results demonstrate that the salt marsh acts as a net source of alkalinity. Lateral fluxes due to the
discharge of saline groundwater triggered by tidal pumping at low tide are a major source of alkalinity and DOC. Diffusive
fluxes at the sediment-water interface during immersion represent another process of alkalinity flux, although of lower
465 amplitude. In contrast, the POC accumulation rate could not be accurately quantified.

4.1 Methodological comparison to Existing Approaches

Estimating carbon and TA exports from coastal wetlands is essential quantifying their potential contribution to blue carbon
sequestration and their role in the marine carbon cycle (Santos et al., 2021). Here, we chose to directly measure the water
discharge in the tidal creek and to compare the concentrations measured in the collected samples with those predicted by a
470 two-endmember mixing model, using the water at high tide and the connected surface freshwater as the endmembers.
Concentrations above the mixing line are interpreted as reflecting the contribution of a third endmember enriched in the
component considered, in this case sediment pore water. Conversely, the values below the mixing line may result either from
dilution by less concentrated interstitial water or from degassing or biogeochemical processes. The contribution of interstitial
water was then calculated directly from the deviation from the mixing line at a given salinity and from the discharge
475 measurement in the channel. This approach allows the gross flux, defined as the concentration multiplied by the discharge, to
be separated from the net flux from interstitial water, defined as the difference in concentration relative to the conservative
mixing multiplied by the discharge. Importantly, this approach did not require direct characterization of the interstitial water
source in order to assess its impact on the tidal creek. Measurement within the tidal creek were sufficient and integrated the
full range of spatial and temporal variability in the groundwater and pore water sources that feeds the tidal creek.

480 Natural isotopic tracers, such as Ra isotopes or ^{222}Rn , have frequently been used to infer porewater exchanges rates and the
associated dissolved species fluxes in mangroves and coastal wetlands (Burnett and Dulaiova, 2003; Faber et al., 2014; Santos
et al. 2019). However, the activities in the groundwater endmembers are frequently identified as the main source of uncertainty
in natural tracer mass balance approaches (Burnett and Dulaiova, 2003). In particular, the use of the radioisotope ^{222}Rn to
estimate porewater fluxes relies on a sequence of successive assumptions and calculations (Rodellas et al., 2021). Each step
485 introduces sources of uncertainty, and the cumulative uncertainty may be comparable to, and even exceed the natural variability
of the signal, especially at the seasonal timescale. The porewater ^{222}Rn endmember can be highly variable in both space and
time within a given site, resulting in a poor control on this parameter. In many ^{222}Rn mass balances, the parameterisation of
atmospheric escape also represents a major sink. Its quantification relies on empirical formulations and may account for 50 %
of the variability, depending on hydrodynamic conditions and surface roughness. In mangrove and salt marsh ecosystems, the
490 natural spatial heterogeneity further complicates this estimation (Santos et al., 2019). Moreover, steady-state assumptions may
not be valid in systems subject to strong tidal and diel forcing. The uncertainties associated with such assumptions are rarely
quantified exhaustively, but they may be a similar or larger source of error than analytical uncertainties (Rodellas et al., 2021).



A well-constrained interstitial water flow by radon does not guarantee proper allocation of carbon fluxes without additional constraints on the carbonate system (Santos et al., 2019). Other studies have estimated TA and DIC outwelling using a mixing curve plot, but without direct measurements of creek discharge (e.g. Zhang et al., 2026). Fluxes were then estimated from the mean residence time of water. This approach assumes that the system is at steady state. Although, it can provide estimations of average flows, but it does not consider of short-term variations, such as those that affecting changes in freshwater discharge endmember or the tidal range. By contrast, the combination of discharge measurements with a linear salinity mixing model does not rely on a long inference chain. Since the error in the measured flow rates is less than 20 % and while analytical uncertainty is much smaller, the instantaneous fluxes of carbon species estimated here are relatively accurate. The error—which is more difficult to estimate—is greater when extrapolating our data to an annual timescale, because field measurements were conducted approximately once per month. However, this sampling strategy required considerable fieldwork and combined both tidal cycles with the seasonal variability over an entire year. This approach allowed for a more accurate mass balance of the fluxes than some previous studies based on a different sampling strategy (e.g., Bai et al., 2026; Zhang et al., 2026), which focused on achieving good spatial coverage of the sites.

4.2 Description of endmember and salt marsh chemistry

This study demonstrates the high temporal variability of dissolved carbon and redox species that can occur over a 6-hour low tide cycle in a salt marsh tidal creek. By accounting for both seasonal and tidal variabilities, the sampling strategy, provides new insights in salt marshes carbon cycle. The mixing line model approach, based on two identified hydrological endmembers, enabled the calculation of the net DIC and TA fluxes without requiring additional proxies. The two endmembers influenced salt marsh water chemistry in different ways. Their characterisation is therefore essential to correctly describe the processes occurring within in the salt marsh. The freshwater or continental endmember reflects the chemistry of the watershed, which is in this region mainly covered by pine forests growing on acidic sandy podzols (Jolivet et al., 2007; Polsenaere et al. 2013a). The water in the watershed streams originates directly from the aquifer within the sandy substrate. It is enriched in compounds resulting from the mineralisation of soil organic matter, including CO₂ and methane (Deborde et al., 2010; Deirmendjian et al., 2018) as well as reduced compounds such as Fed and Mnd (Deirmendjian et al., 2019). Portions of the watershed are also covered by agricultural land, which are a source of nitrate. Water from the watershed flows into the coastal lakes of Carcans-Hourtin and Lacanau. These lakes retain a significant amount of the nutrients (Buquet, 2017), and allow partial exchange of dissolved gases with the atmosphere (Mayen et al., 2025). The freshwater source influencing the salt marsh investigated here originates from the 18-km-long canal connecting the lake to Arcachon Bay. This canal also collects water from tributaries that flow directly into it without passing through the lakes. Consequently, the freshwater source is supersaturated with CO₂ and CH₄ compared to atmospheric equilibrium. Because of the acidic nature of the soils, alkalinity in region's watercourses is very low, often below 200 μM. The water residence time in the lakes and the biogeochemical processes occurring there generate alkalinity (Buquet, 2017). Nevertheless, the freshwater endmember sampled at the entrance to the salt marshes had an alkalinity below 300 μM, indicating that freshwater is not a major TA source to the salt marsh. The seawater endmembers were



characterised by their salinity and the associated chemical species as SO_4^{2-} , TA, DIC, Cad and Nad that are major components of seawater. The high-tide water collected at the tidal creek sampling site was not sampled precisely at high tide, but rather during the early ebb tide, when the salt marsh had begun to be exposed while only the channel remained submerged. The water was more diluted by the freshwater source than the water collected at high tide, which occurred 1.5 km from the site. In both cases, salinity was below 35, reflecting the estuarine nature of the bay. Since the samples taken at high-tide can be influenced by other sources of freshwater, particularly the main river discharging into the Arcachon Lagoon, we chose to use the tidal creek water sampled at the beginning of the ebb tide as the most representative of the local marine point. This will provide, more accurate identification of changes associated with biogeochemical processes within the salt marsh.

4.3 Carbon outwelling

Oxygen concentrations, which increased during tidal monitoring, with reached values close to or even exceeding saturation, suggesting that photosynthetic activity occurred in the tidal creek. A clear difference was also observed between TA fluxes and DIC fluxes. At the pH values measured in the creek waters, this difference is primarily due to the contribution of dissolved CO_2 to DIC, in addition to HCO_3^- . In other words, CO_2 concentrations are higher than those resulting from the mixture of freshwater and high-tide endmembers, despite CO_2 degassing and photosynthetic uptake. This suggests that the channel is being supplied with water that is enriched in both TA and CO_2 . The steady lightening of $\delta^{13}\text{C}$ -DIC during low tide further supports that the DIC accumulation was largely sustained by benthic organic matter mineralisation rather than by water mass mixing alone. Creek waters also accumulated reduced species (Fed, Mnd). Given the rapid kinetics of Fed oxidation in oxygenated waters, with a half-life of only a few minutes at circumneutral pH (Millero et al., 1987), the persistence of dissolved Fe under oxygenated conditions indicates a recent and continuous supply of benthic-derived reduced iron. This observation is consistent with active tidal pumping of sediment pore waters. Similarly to Fed and Mnd, creek waters at low tide were enriched in TA, DIC, and dissolved species typically found in reduced porewaters of Arcachon Bay sediments (Deborde et al., 2008b; Delgard et al., 2016; Mouret et al., 2016).

The mean net TA flux was $1149 \pm 412 \text{ mol d}^{-1}$ in the tidal creek draining the 0.3 km^2 saltmarsh area. Assuming a TA:C molar ratio of 1:1 and normalizing the annual carbon-equivalent flux to the drainage area, this flux corresponds to $17 \pm 6 \text{ gC m}^{-2} \text{ yr}^{-1}$. This net TA flux is considered as a long-term carbon sequestration pathway (Thomas et al., 2009; Hu and Cai, 2011; Krumins et al., 2013). DIC fluxes were higher. At pH values measured in the channel waters—which were closer to pK_1 than to pK_2 —DIC essentially represented the sum of bicarbonate and dissolved CO_2 , whereas carbonate ions accounted for less than 2 % of DIC. During low-tide, pCO_2 increased significantly (Fig. 2j), reaching values well above atmospheric equilibrium, despite efficient CO_2 degassing and, during certain seasons, photosynthetic CO_2 uptake in the creek bed, accompanied by O_2 production. The increase in pCO_2 therefore also reflects an additional of an internal source of CO_2 within the channel, consistent with tidal pumping of porewaters. The DIC flux of $1548 \pm 486 \text{ mol d}^{-1}$ can therefore be considered as the sum of the alkalinity flux and the CO_2 flux. In contrast to alkalinity export, this CO_2 excess does not contribute to long-term carbon sequestration because it can be released to the atmosphere. We estimated a mean degassing rate ranging from 0.5 to $5 \mu\text{mol}$



$\text{m}^{-2} \text{s}^{-1}$, comparable to the CO_2 efflux of $3.6 \pm 0.63 \mu\text{mol m}^{-2} \text{s}^{-1}$ reported for a temperate salt marsh tidal creek by Trifunovic et al. (2018). It is easy to imagine that at high tide, when the exchange surfaces between the water and the atmosphere are larger, the pCO_2 of the water tends to quickly re-equilibrate with that of the atmosphere, which is approximately what we observed at high tide (Fig. 2 j).

Methane fluxes were very low, averaging 1.8 mol d^{-1} or $6 \mu\text{mol m}^{-2} \text{d}^{-1}$ or $0.9 \text{ gCO}_2\text{eq m}^{-2} \text{yr}^{-1}$. This result is consistent with several studies conducted in intertidal environments and saltmarsh. In the Arcachon Bay, a water-air methane flux of 0.7 to $2.8 \mu\text{mol m}^{-2} \text{h}^{-1}$ was measured in a tidal creek (Deborde et al., 2010). Poffenbarger et al. (2011) demonstrated that CH_4 production is inhibited when salinity exceeds 18, which was typically the case for most of the sampling dates of this work. Neubauer and Megonigal (2015) also reported negligible CH_4 fluxes in salt marshes compared to freshwater marshes, with values often below $50 \mu\text{mol m}^{-2} \text{d}^{-1}$. Magenheimer et al. (1996) reported that CH_4 emissions remain very low relative to CO_2 emissions in saline marshes (molar ratio < 0.05).

The high frequency approach used in this study captured only ebb tides variability. No clear spring/neap tidal pattern emerged, suggesting that other environmental factors (temperature, local primary production, microbial processes, freshwater or nutrient inputs) primarily control these fluxes at the marsh outlet. The FCO_2 and FCH_4 measured in this study represent net emissions to the atmosphere, and can be interpreted as a component of ecosystem respiration. However, they are not equivalent to net ecosystem exchange (NEE) because they do not integrate the full balance between gross primary production and ecosystem respiration across the entire salt marsh.

DOC fluxes were highly variable but in the same order of magnitude than DIC fluxes with an average $12 \pm 10 \text{ gC m}^{-2} \text{yr}^{-1}$. Unlike TA, the DOC fluxes may be further transformed, depending of its own lability. DOC may be mineralised or photo-oxidised and ultimately converted into CO_2 (Tzortziou et al., 2011). Net CH_4 lateral fluxes were on average $2.2 \pm 2.1 \text{ mol d}^{-1}$. This value represents the residual fraction of CH_4 exported through tidal pumping, because CH_4 from interstitial waters can be rapidly oxidised in oxygenated waters (Deborde et al., 2010). This residual flow is of the same order of magnitude as the transfer rate to the atmosphere calculated based on a 500-meter path through the creek. This suggests that methane is released very rapidly into the atmosphere. Considering a 100-year global warming potential of 28 relative to CO_2 (Intergovernmental Panel on Climate Change (IPCC), 2014), the CH_4 flux expressed as CO_2 equivalent was on average 10 to 20 times lower than lateral TA, DIC and DOC fluxes. Thus, net lateral methane fluxes in this salt marsh do not represent a major greenhouse-gas source as it was considered in some other salt-marshes (Santos et al., 2019; Yau et al., 2022; Liu et al., 2025).

The measured net carbon outwelling was $3.8 \text{ mmol m}^{-2} \text{d}^{-1}$ for TA and $5.2 \text{ mmol m}^{-2} \text{d}^{-1}$ for DIC. These values are much lower than the rates reported for other sites. For example, DIC exports of $319 \pm 177 \text{ mmol m}^{-2} \text{d}^{-1}$ were measured in an estuarine tidal creek at Evans Head (Santos et al., 2019), $35 \pm 13 \text{ mmol m}^{-2} \text{d}^{-1}$ at the mouth of a drainage channel of a coastal wetland (Sadat-Noori et al., 2024), or $42 \pm 31 \text{ mmol m}^{-2} \text{d}^{-1}$ in a macrotidal salt marsh (Yau et al., 2022b). However, these fluxes corresponded to gross fluxes. In the present study, the gross flux of DIC was $51 \pm 38 \text{ mmol m}^{-2} \text{d}^{-1}$. A significant fraction of the gross flux corresponds to the conservative transport of DIC and alkalinity associated with the seawater mixing endmember. Gross fluxes primarily depend on the tidal exchange, with a larger volume of water flushing out during spring tides than during neap tides.



Thus, gross fluxes are not direct measures of the outwelling but mainly the reflect of the retreat of the tide, which a small portion is pore-water outwelling. The binary mixture model using salinity as a conservative tracer of the relative proportions between high tide water (marine endmember) and freshwater (continental endmember), allows to extract the signature of non-conservative processes as TA tidal pumping.

4.4 Processes involved in TA and DIC production

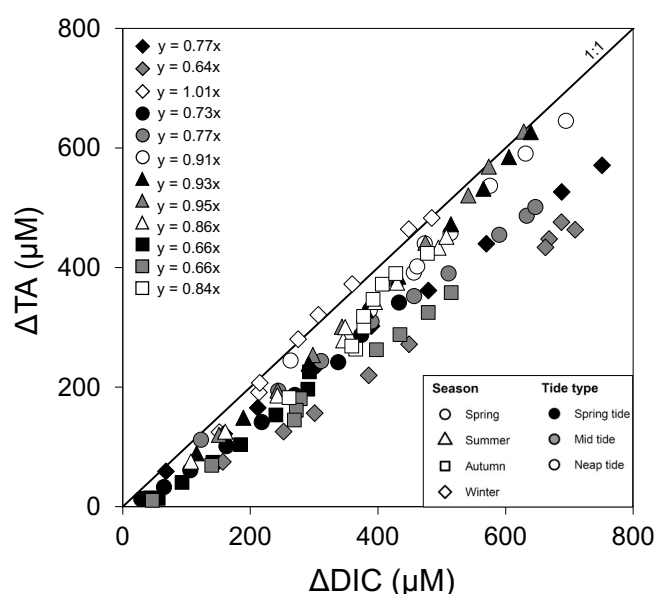
Atmospheric CO₂ is captured by photosynthesis. The organic matter produced in this way represents a storage of atmospheric carbon if it is buried in sediments over the long term. In coastal environments, this is known as blue carbon. However, a proportion sedimentary organic matter is mineralised, meaning that carbon storage may only be temporary. If organic is converted back into CO₂ through aerobic respiration, the net carbon balance will be zero. However, if organic carbon mineralisation results in the production of bicarbonate, carbon can be trapped in the ocean over long timescales. This process is associated with the alkalinity production during anaerobic organic matter mineralization. Denitrification, reduction of iron and manganese oxides, and sulphate reduction are these major anaerobic organic matter mineralization processes (Kempe, 1990; Soetaert et al., 2007; Thomas et al., 2009). The slope of the $\Delta\text{TA}:\Delta\text{DIC}$ relationship helps to distinguish among these processes. This slope reflects the stoichiometry of the predominant reactions contribute to the observed changes in the dissolved carbonate chemistry and thus can be used to infer which reactions control dissolved carbonate chemistry (Cai et al., 2003; Liu et al., 2025). For example, this slope would be 2 if the increase in alkalinity was due to the carbonate dissolution. A slope close to 1 is consistent with denitrification and sulphate reduction processes. The values obtained for the 12 dates ranged from 0.64 to 1.01 (Fig. 6). This is consistent with a combination of sulphate reduction ($\Delta\text{TA}:\Delta\text{DIC} = 1$), denitrification ($\Delta\text{TA}:\Delta\text{DIC} = 0.8$) and aerobic respiration ($\Delta\text{TA}:\Delta\text{DIC} = 0.15$). Denitrification was probably not a major source of alkalinity, because nitrate concentrations in the high tide water were always much lower than the DIC in excess measured at low tide. The reduction of Fe(III) and Mn(III, IV) oxides is a net source of alkalinity, with $\Delta\text{TA}:\Delta\text{DIC}$ ratio is 8 and 4 for these processes, respectively. The obtained values were well below 4, suggesting that metal oxide reduction was not a major source of alkalinity in Arès salt marsh. The redox cycle of iron is known to be an important process governing the geochemistry of phosphate in Arcachon Bay sediments (Deborde et al., 2008a). However, Fe(III) and Mn(III,IV) oxide reduction is a net source of alkalinity only if the resulting reduced Fe(II) and Mn(II) are not subsequently reoxidized and precipitated again and instead accumulate in water, which is generally not the case in sediment pore waters (Burdige, 2007). The oxidation of dissolved iron and manganese consumes TA (Wolf-Gladrow et al., 2007; Hu and Cai, 2011). We observed that tidal pumping generated a flux of dissolved reduced Fe and Mn in the tidal creek. However, given the low concentrations measured in the oxic creek waters, their stoichiometric contribution to TA consumption following complete oxidation remained negligible relative to the amount of TA released.

Among the secondary processes that can produce TA and DIC, calcium carbonate dissolution can be major process in some coastal environments (e.g. Tutiyaarn et al., 2025) producing a $\Delta\text{TA}:\Delta\text{DIC}$ ratio of 2. Here, the calcium concentration remained



on the conservative mixing line within analytical uncertainty. Most of the calcium excess measured in this study were within the analytical uncertainty (5 %). This indicates that calcium carbonate dissolution did not significantly contribute to TA or DIC production. Moreover, this process is not consistent with the observed increase in $p\text{CO}_2$ and decrease in $\delta^{13}\text{C}\text{-DIC}$ (Fig. 2). The addition of DIC to creek water shifted its isotopic signature to negative values, suggesting that this additional
630 DIC originated from a lighter carbon source. Marine carbonates have a relatively heavy signature, typically close to 0 ‰, whereas organic matter is characterised by more negative $\delta^{13}\text{C}$ values.

Sulphate reduction probably represents the main source of excess TA in creek waters, as indicated by sulphate deficits in creek waters. This process produces two moles of TA per mole of sulphate reduced to S(-II) species, which may either escape to the atmosphere when sediments are exposed at low tide (Kristensen et al., 2000), or be buried as Fe-sulphides (Reithmaier et al.
635 2021). However, $\Delta\text{TA}:\Delta\text{DIC}$ values were generally below 1 and the observed increase in $p\text{CO}_2$ suggests that an additional process producing more DIC than TA also influenced the carbonate chemistry of the salt marsh. The regression results therefore indicate that the observed TA and DIC production reflects a combination of aerobic respiration, denitrification and sulphate reduction.



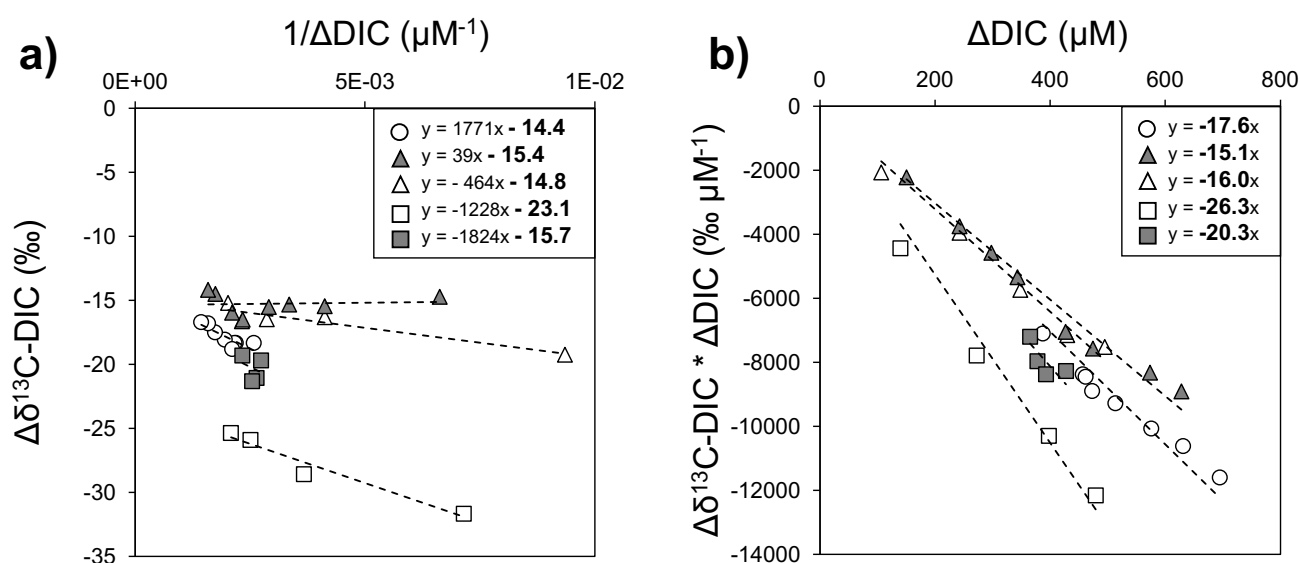
640 **Figure 6 : ΔTA vs. ΔDIC plot.** Each point position represents the excess of Alkalinity (ΔTA) and Dissolved Inorganic Carbon (ΔDIC) in samples taken in the creek during ebb tides sampling. The filling of the point represents the tide type and forms represent the season. The $\Delta\text{TA}:\Delta\text{DIC}$ slope of each sampling date is resumed as: $y = ax$ with its marker on the upper left section. Black line represents the 1:1 position of the ratio.

Keeling intercepts and origin-forced Miller-Tans slopes (Miller and Tans, 2003) yielded consistent source signature estimates
645 for most campaigns differing by only 0.3 to 4.6 ‰ (Fig. 7). This supports the overall robustness of the extrapolated values once conservative mixing is accounted for. Source signatures, estimated from the origin-forced Miller-Tans regressions, ranged from approximately from -15 ‰ to -26 ‰ among campaigns. These values are markedly less negative than the uncorrected



estimates obtained by applying the Keeling approach directly to total DIC and $\delta^{13}\text{C}$ -DIC without isolating the conservative mixing component (Appendix C). This discrepancy is expected because the system involves three isotopically distinct end-
650 members: high tide marine water, freshwater input, and the biogeochemical DIC source, whereas a standard Keeling plot assumes only two components. When applied directly to total DIC and $\delta^{13}\text{C}$ -DIC, the regression cannot distinguish the physical mixing between the marine and freshwater end-members from the biogeochemical addition of DIC. Consequently, the intercept represents an unresolved composite of these contributions rather than the isotopic signature of the biogeochemical DIC source alone. The unresolved contributions of the freshwater and marine end-members bias the extrapolated intercept toward more
655 depleted $\delta^{13}\text{C}$ values relative to the biogeochemical source, thereby explaining the systematic offset between corrected and uncorrected estimates.

The corrected signatures are consistent with mineralisation of a mixture of marine phytoplankton-derived organic matter ($-22.7\text{‰} \pm 0.8$; Dubois et al., 2012) and terrestrial/riverine particulate organic matter ($-28.0\text{‰} \pm 1.3$; Dubois et al., 2012), with a variable contribution of marsh angiosperm-derived organic matter ($-11.9\text{‰} \pm 1.4$; Dubois et al., 2012) among campaign.
660 These results suggest that the organic carbon mineralised in salt marsh sediments originated from a mixture of marine, terrestrial and local marsh sources. Importantly, this invalidates the need for an additional biogenic methane oxidation pathway as invoked in the classical Keeling plot to explain the intercept values below -55‰ . These values are no longer supported once conservative mixing is properly accounted for (see Appendix C).



665 **Figure 7 : a) Keeling plot of $\delta^{13}\text{C}$ -DIC, ‰ vs. VPDB) as a function of the inverse of DIC concentration and b) Miller-Tans regressions, excess $\delta^{13}\text{C}$ -DIC multiplied by excess DIC vs. excess DIC) in tidal creek waters across seasons and tidal conditions. Dashed lines show seasonal linear regressions whose y-intercepts for Keelin and slope for Miller Tans reflect the isotopic signature of the sediment-derived DIC end-member. Legend is the same as Fig. 6.**



4.5 Carbon fluxes of Arès salt marshes

670 The data collected during this study provide a robust estimate of the annual lateral carbon fluxes in a salt marsh. These fluxes include DIC and DOC exported through the tidal creeks as a result of tidal pumping. In particular, our sampling strategy enables to quantify the lateral export associated with benthic organic matter mineralisation, and to assess the fraction of this export that may contribute to long-term sequestration of atmospheric CO₂.

DIC fluxes essentially consisted of bicarbonate and CO₂. We showed that CO₂ concentrations in the channel waters were 675 significantly above equilibrium with atmospheric CO₂. Part of this excess is therefore expected to degas during transport through the creek, and particularly at high tide. Methane is also supersaturated and is released at the water–atmosphere interface. However, the gas fluxes calculated from measurements represent only a part of the gas exchanges between the salt marsh ecosystem and the atmosphere, as we did not measure the fluxes associated with photosynthesis and respiration at the of the salt marsh surface.

680 Upscaling our estimates to the worldwide surface of saltmarshes, which is estimated at 90000 km² (Murray et al., 2022), highlights the importance of fluxes identified in this work. Assuming that the tidal pumping fluxes measured at our study site are representative at the global scale, net lateral TA flux would correspond to about 1.5 TgC yr⁻¹, approximately 28 % of the global organic carbon burial rate in saltmarshes (5.4 TgC yr⁻¹; Chmura et al., 2003), suggesting that alkalinity outwelling may represent a significant, yet often overlooked, pathway of long-term carbon sequestration. The global DIC lateral export would 685 be about 2.0 TgC yr⁻¹, of the same order of magnitude as TA export. However, a fraction of this DIC may subsequently return to the atmosphere through air-sea gas exchange, thereby limiting its contribution to net long-term carbon sequestration. The global lateral DOC export is subject to considerable uncertainty with an estimated flux of 1.1 ± 0.9 TgC yr⁻¹, and highlighting the limitations observed at the study site, cautioning against direct extrapolation from a single dataset. Finally, water-air CH₄ fluxes remain negligible at the global scale (0.08 TgCO₂eq yr⁻¹), consistent with the potential inhibition of methanogenesis by 690 sulphate availability under high-salinity conditions in temperate salt marshes.

Lateral fluxes of dissolved carbon observed in tidal creek at low tide can also be compared with diffusive fluxes between water and sediment during inundation. Open-core incubations were designed to reproduce high tide conditions and to quantify benthic TA and DIC fluxes. These measurements confirmed that marsh sediments act as a significant source of inorganic carbon and alkalinity to the overlying water column throughout the year, with TA fluxes greater than 2 gC m⁻² yr⁻¹ in vegetated 695 and bare mud areas. The seasonal pattern was consistent with a temperature-driven intensification of anaerobic respiration during warmer months, as sulphate reduction rates in salt marsh sediments are known to peak in summer and decline as temperatures decreased in winter (Kostka et al., 2002; Wang et al., 2016). The higher and more variable benthic fluxes observed in vegetated cores likely reflect the influence root networks, which generates macropores and preferential flow paths that enhance porewater-water column exchange and generate strong spatial heterogeneity in TA and DIC production (Xin et al., 700 2022). The release of labile organic compounds from roots may further fuels microbial respiration in the rhizosphere, amplifying anaerobic organic matter mineralisation rates relative to bare mud. As vegetation senescens and root systems decay,



mean benthic fluxes in vegetated sediments may converge toward bare mud values, while their variability remains high, because of structural heterogeneity inherited from the decaying root and rhizome network (Kostka et al., 2002).

The particulate organic carbon profiles reveal contrasting sedimentary dynamics between habitat types. The bare mud core shows relatively uniform POC contents of about 3 to 4% throughout most of the profile, probably reflecting limited autochthonous organic matter inputs, followed by a sharp decrease below 30 cm associated with a sandy layer. In contrast, the vegetated core exhibits higher and more variable POC content ($> 4\%$), consistent with enhanced belowground organic matter production by roots and rhizomes (Chmura et al., 2003). Estimates of organic carbon burial in Arès Reserve should be interpreted with caution because $^{210}\text{Pb}_{\text{xs}}$ profiles in dynamic macrotidal environments can be affected by sediment mixing, erosion, accretion and bioturbation (Ouyang and Lee, 2014; Arias-Ortiz et al., 2018). The apparent OC burial rates of $84 \text{ gC m}^{-2} \text{ yr}^{-1}$ for bare mud and $355 \text{ gC m}^{-2} \text{ yr}^{-1}$ for vegetated zones fall in global salt marsh average value of $218 \pm 24 \text{ gC m}^{-2} \text{ yr}^{-1}$ (Chmura et al., 2003). Carbon accumulation rates in salt marshes range from 18 to $1713 \text{ gC m}^{-2} \text{ yr}^{-1}$ (McLeod et al., 2011), reflecting the extensive variability among sites driven by differences in tidal range, sediment supply, and vegetation. The nearly four-fold difference between bare muds and vegetated zones of the Arès saltmarsh underlines the critical role of vegetation in promoting both sediment trapping and organic matter production. Considering the surface areas of bare and vegetated muds, the storage rate of carbon in the creek catchment is approximately 100 tons yr^{-1} . However, this value should be treated with caution, as the obtained $^{210}\text{Pb}_{\text{xs}}$ depth profiles are relatively vertical, particularly in vegetated sediments. The apparent SAR value of 1.6 cm yr^{-1} suggests that the topographic level was 1 m lower in the 1960s than it is today. This would imply a major change in the salt marsh landscape—which is not at all the case, according to descriptions from local hunters and aerial photographs from that period (<https://cartes.gouv.fr/explorer-les-cartes/>). In this type of environment, sediment accommodation can be considered to be at least the mean sea level rise. The analysis of long-term sea level trends reveals an acceleration in the average rate of sea level rise from 1.2 mm yr^{-1} during the 20th century to 2.4 mm yr^{-1} over the last two decades (Dodet et al., 2019). These values are, in fact, an order of magnitude lower than those we estimated with radionuclides. Therefore, the carbon accumulation rate, which corresponds to the sedimentary component of blue carbon, is likely overestimated here.

Salt marshes and particularly macrotidal systems such as Arcachon Bay, are also highly dynamic environments subject to intense lateral erosion and creek migration (Fagherazzi et al., 2012). Organic carbon deposited in marsh sediments can be remobilised following habitat disturbance and erosion of marsh edges, releasing stored carbon back to the atmosphere or coastal ocean rather than being permanently buried (Macreadie et al., 2013). This implies that the apparent OC burial rates estimated here from $^{210}\text{Pb}_{\text{xs}}$ profiles likely represent gross rather than net sequestration, and the actual long-term carbon sequestration capacity of this salt marsh may actually be lower than suggested by sediment core-based estimates alone.

Integrating the measured carbon fluxes provides an overall comprehensive assessment of the carbon budget of the Arès salt marsh (Fig. 8). TA outwelling represented an important potential pathway for long-term carbon sequestration pathway, because alkalinity exported to the coastal ocean can increase the capacity of seawater to store inorganic carbon over very long timescales. Lateral DIC fluxes constituted the largest dissolved carbon export but cannot be considered entirely as net



sequestration because a fraction of the exported DIC may be subsequently released to the atmosphere through air-sea CO₂ exchange. DOC outwelling is of similar magnitude although its contribution to long-term carbon sequestration depends on its persistence and lability, with a significant fraction to be likely oxidised or remineralised in adjacent coastal waters. Air-water CO₂ fluxes and CH₄ fluxes represent non-negligible greenhouse gas emissions to the atmosphere. Sediment OC burial, constitutes the largest potential carbon sink, although its effective sequestration capacity is moderated by the dynamic nature of this tidal system and the associated risk of POC remobilisation. Overall, the salt marsh functions as a net carbon sink, primarily through alkalinity export and sediment carbon burial.

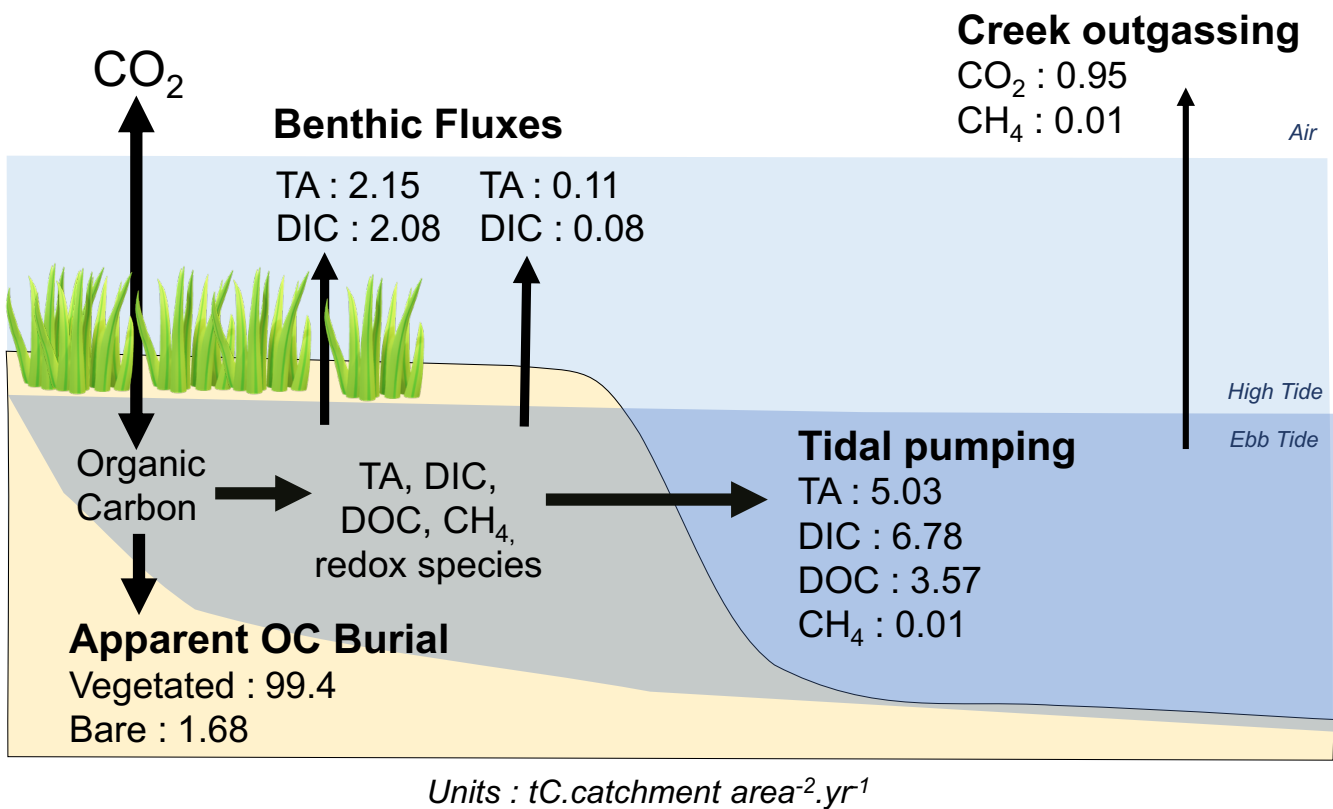


Figure 8 : Conceptual diagram of the carbon budget of the Arès salt marsh (Arcachon Bay, SW France), integrating all measured carbon fluxes expressed in tC y⁻¹ for the creek catchment area (0.3 km²). Organic carbon produced by marsh vegetation is either buried in sediments or mineralised through aerobic and anaerobic respiration reactions, generating dissolved species (TA, DIC, DOC, CH₄ and redox-sensitive solutes) that accumulate in sediment porewaters. During ebb tide, tidal pumping exports these dissolved species to the adjacent coastal ocean. A fraction of the dissolved carbon exported to the tidal creek is outgassed to the atmosphere as CO₂ and CH₄ through air-water gas exchange at the creek surface. High tide submersion drives benthic fluxes of TA and DIC from sediments to the overlying water column, with higher fluxes in vegetated areas than in bare mud. Benthic fluxes are normalised by the surface of the corresponding sediment nature.



Conclusion

755 This study demonstrates the high temporal variability of dissolved carbon and redox species that can occur over a single low tide cycle lasting about 6 hours in a salt marsh tidal creek. Over this time period, TA, DIC and DOC concentrations increased above values predicted by the conservative mixing line between high tide and freshwater endmember, resulting in substantial excesses of these dissolved species. Tidal pumping, which is defined here as the release of sediment porewaters during low tide due to difference in altitude and hydrologic pressure, is likely a major process driving these increases. Porewaters, enriched in dissolved carbon species, seep out from marsh sediment and feed the tidal creek during ebb tide. The methodological approach developed here, which combines direct discharge measurements with a binary salinity mixing model, provides a robust alternative to radon-based porewater tracing for quantifying net dissolved carbon outwelling. This approach does require the parameterisation of degassing terms or the use of poorly constrained subsurface endmembers.

760 Net TA outwelling, which is largely associated with sulfate reduction coupled to FeS/FeS₂ formation in marsh sediments, is a potentially stable and long-term pathway for carbon sequestration. We propose the term Alkalinity Blue Carbon (ABC) to describe the net TA outwelling relative to the mixing line or salinity reference. We suggest that this deserves recognition alongside the Burial of organic carbon in Blue Carbon (BBC) inventories. Unlike the burial of particulate organic carbon, which may be vulnerable to erosion and remobilization in dynamic macrotidal systems, alkalinity export to the coastal ocean can contribute to the formation of a chemically conservative and durable carbon sink over centennial to millennial timescales.

770 Sediment OC burial rates were important, particularly in vegetated areas, but should be interpreted as gross rather than net sequestration estimates because of the strong erosion/sedimentation dynamics characteristic of this macrotidal system. Air-water CO₂ and CH₄ fluxes were comparatively modest and consistent with the other processes highlighted in this work. Overall, this salt marsh functions as a net carbon sink with alkalinity outwelling and sediment organic carbon burial representing the dominant sequestration pathways, while greenhouse gas emissions provide only a minor offset. Upscaling these fluxes to the global salt marsh areas could substantially increase current estimates of blue carbon sequestration capacity. However, such extrapolation must explicitly account for spatial variability in hydrodynamics, sediment properties, vegetation and biogeochemical functioning among salt marsh ecosystems. These findings emphasize the importance of protecting and restoring these ecosystems in the context of ongoing global change, which are characterized by accelerating sea level rise, coastal erosion, and anthropogenic pressures.

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Appendix A

795 The estimate of CO₂ and CH₄ degassing at the water-atmosphere interface takes into account the gradient between the concentration of dissolved gas and the concentration in equilibrium with the atmosphere [gas]_{sat}, as well as the gas transfer rate (k):

$$Flux = k * ([gas]_{water} - [gas]_{sat}) \quad (A1)$$

The values of k are estimated using empirical formulas that take these parameters into account (Raymond et al., 2012), with:

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$$k = k_{600} * \frac{Sc}{600} - n \quad (A2)$$

In this context, k₆₀₀ refers to the value of k, which has been normalised at 20°C for a Schmidt number (Sc) of 600. The dimensionless Schmidt number is the ratio between the kinematic viscosity of water (m² s⁻¹) and the diffusion coefficient of gas (m² s⁻¹). The factor n depends on turbulence and can be considered equal to 0.5 for river water (Jähne et al., 1987; Raymond et al., 2012). For CO₂, two empirical formulas were used to estimate Sc as a function of temperature (T, in °C) (Wanninkhof, 1992; Raymond et al., 2012).

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$$Sc_1 = 1911 - 118.11T + 3.453T^2 - 0.0413T^3 \quad (A3)$$

or

$$Sc_2 = 1742 - 91.24T + 2.208T^2 - 0.0219T^3 \quad (A4)$$

The values range from 380 to 1567 for measured temperatures between 3 and 29°C. The two empirical equations give very similar values in this temperature range, with relative differences of 0 to 10%.

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For CH₄, the two empirical formulas given by Raymond et al., (2012) were used:

$$Sc_1 = 1824 - 98.12T + 2.413T^2 - 0.0241T^3 \quad (A5)$$

or

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$$Sc_2 = 1898 - 114.28T + 3.29T^2 - 0.0391T^3 \quad (A6)$$

Values computed for a given temperature were within 5% for the two equations.



Two empirical formulas for estimating k_{600} have been tested (Eq. A7 and Eq. A8). These equations take into account the flow velocity (V , m s^{-1}), the slope (S , unitless), and the depth (D , m), and are the most accurate for determining a gas-transfer velocity for small first order systems such as the tidal creek, and correspond to equations 1 and 2 of Raymond et al., (2012):

$$k_{600} = 5037 \times (VS)^{0.89} \times D^{0.54} \quad (\text{A7})$$

or

$$k_{600} = 5937 \times (VS)^{0.89} \times D^{0.58} \times \left(1 - 2.54 \times \frac{V^2}{gD}\right) \quad (\text{A8})$$

The parameters used correspond to the values measured at the sampling point for a given date. By combining the two Sc values and the two k_{600} values, we obtained four k values for each period of sampling.

Appendix B

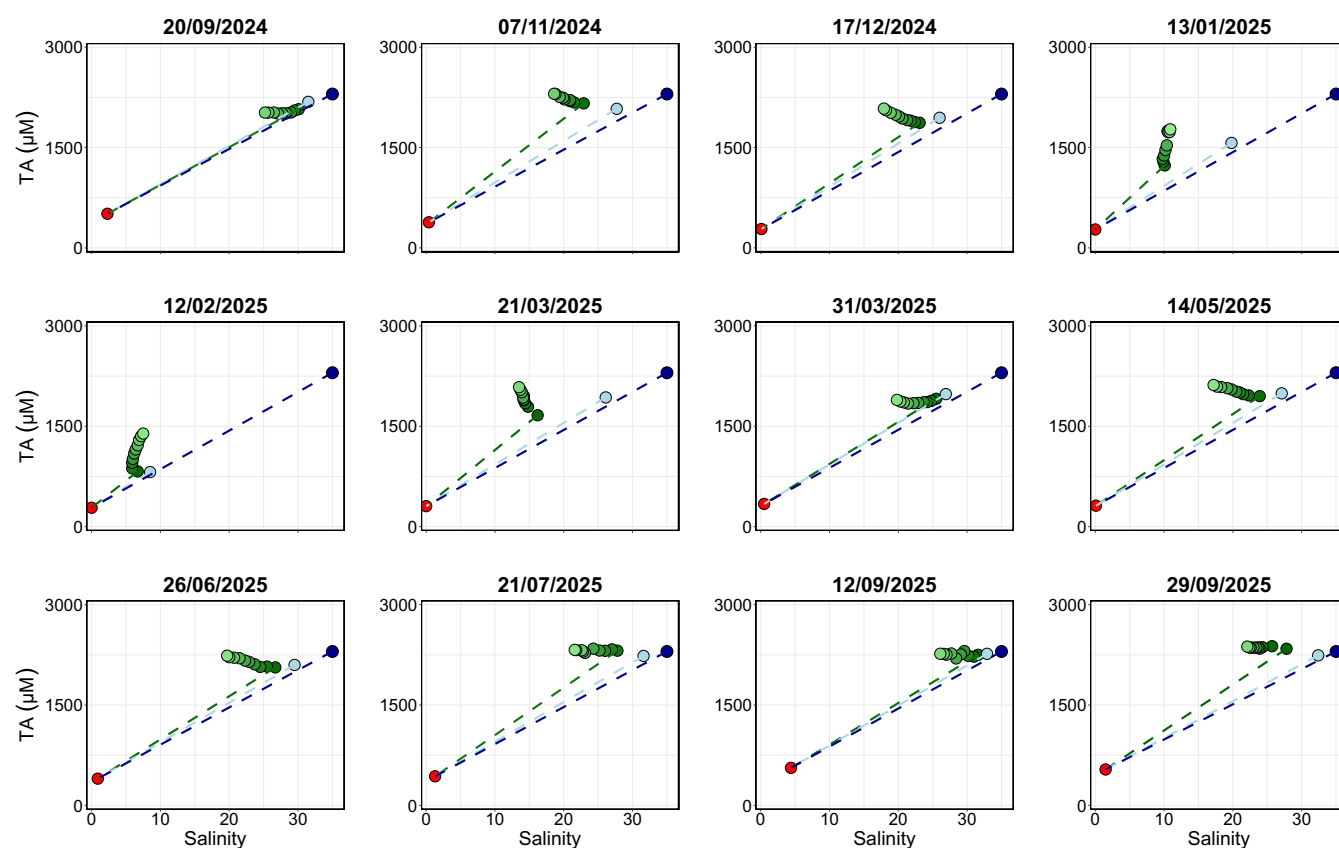
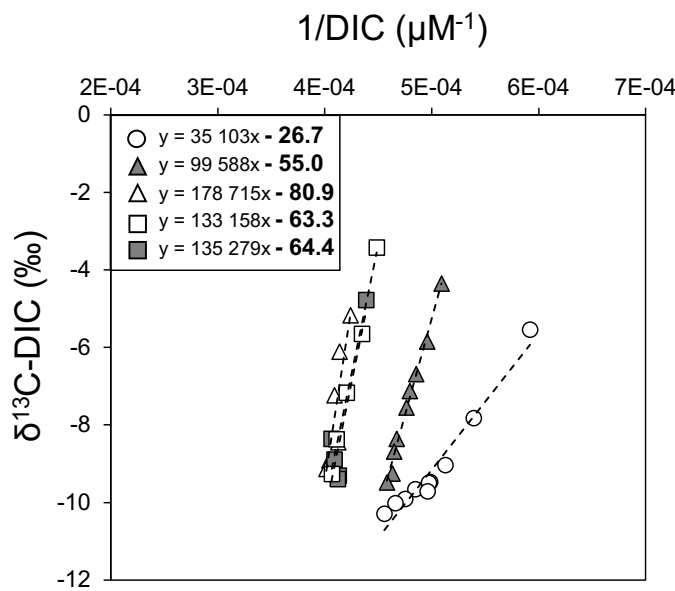


Figure B1: Representation of conservative mixing models for each sampling date. Theoretical mixing lines (dashed lines) connect the following endmembers: offshore theoretical seawater (dark blue), the outer high-tide water sample from the inlet adjacent to the salt marsh (light blue), and the inner high tide point which is the first salt marsh water sample collected at the beginning of the ebb tide (dark green). Individual water samples collected during the subsequent ebb-tide time series within the salt marsh are shown as green circles for whom the shades, from darkest to lightest, represents the time after high tide.



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Appendix C



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Figure C1: Keeling plot of $\delta^{13}\text{C-DIC}$ (‰ vs. VPDB) as a function of the inverse of DIC concentration ($1/\text{DIC}$, μM^{-1}) in tidal creek waters across seasons and tidal conditions. Dashed lines represent seasonal linear regressions, whose y-intercepts reflect the isotopic signature of the sediment-derived DIC endmember. The corresponding regression equations (slope and intercept) are shown for each dataset. Symbols and colour codes are the same as in Figure 7.

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Code and data availability

All numerical data will be accessible in the appendices of Paul Kanfer's thesis manuscript in spring 2027 on the website <https://theses.fr>, by typing Paul Kanfer on the search engine of this site. In the meantime, the data is available upon request to PK and PA.

860 **Supplement link**

Team list

Author contributions

PK conducted the experiment and was responsible for the data collection, analysis, and writing of the manuscript. CC, SS and PA were responsible for the conceptualization, methodology, and supervision of the study. PA was in charge of funding acquisition; PK, CC, PA, JD, GL and MD conducted the field work and collected the samples; SS supervised analysis radionuclides, CC, the analyses of $d^{13}C$ -DIC, and BD the core incubations. PK, AM, EB, MD, DP, and PA realized analysis of the other parameters in the laboratory. PK, PA, SS and BD worked on the initial draft. All authors contributed to the final manuscript.

Competing interests

870 The authors declare that they have no conflict of interest.

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Review statement

The review statement will be added by Copernicus Publications listing the handling editor as well as all contributing referees according to their status anonymous or identified.

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