

Faculté des bioingénieurs

Assessing the contribution of carbon export from Seychelles' seagrass meadows to blue carbon sequestration

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Abstract

Marine ecosystems are increasingly recognised as a complementary lever to emission cuts in climate change mitigation. Seagrass meadows sequester carbon in two ways: locally through burial beneath the canopy, and remotely through the lateral export of detached fragments to the deep ocean and sedimentation. National inventories credit only the buried share, leaving this lateral pathway unaccounted. This study aims to quantify the missing pathway for *Thalassodendron ciliatum* meadows in the Seychelles. The multiscale hydrodynamic model SLIM¹ drives an offline Lagrangian particle tracking module developed for this thesis. The module releases 253 million *T. ciliatum* propagules across the archipelago between 2016 and 2020 and follows each to its terminal fate. After that, a Monte Carlo run propagates uncertainty on meadow-production parameters.

Long-term sequestration captures 71 % of the exported carbon under the model's assumptions, equivalent to 12 % of meadow net primary production. Within this sequestered pool, 36.0 ± 2.2 % settles inside the Exclusive Economic Zone (EEZ), 35.0 ± 6.8 % settles beyond national borders with comparable residence times, and 29.0 ± 5.2 % is lost to coastal stranding and in-transit decomposition. Meadow location shapes the within-EEZ share more than the monsoon does, with values ranging from near zero to 44.4 % across the archipelago (area-weighted mean 34 %) and inner-plateau Mahé meadows reaching only 21 %.

The Monte Carlo median within-EEZ flux reaches 53 kt C per year (193 kt CO₂e, full range 59 to 447 across meadow-production parameters, with hydrodynamic and interannual uncertainty not propagated). This flux is not yet recognised in carbon accounting frameworks. If it were, it would raise the buried-only national baseline of 510 kt CO₂e by 38 %, approach the national transport sector, and reach 40 % of the energy sector. Under emerging voluntary blue-carbon markets, such a flux could also be valorised. More broadly, this thesis underlines why carbon export from seagrass meadows warrants closer investigation as a component of blue-carbon accounting.

¹Second-generation Louvain-la-Neuve Ice-ocean Model.

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Contribution Statement

	Personal contribution	Other contributors	Artificial intelligence
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Formal analysis	X^1	X^2, X^4, X^9	X^6
Validation	X^1	X^2, X^4, X^9	X^6
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List of acronyms

ΔGMST Change in Global Mean Surface Temperature relative to the pre-industrial period.

ADCP Acoustic Doppler Current Profiler.

Argo Array for Real-time Geostrophic Oceanography.

BCE Blue Carbon Ecosystems.

CTD Conductivity Temperature Depth.

DIC Dissolved Inorganic Carbon.

DOC Dissolved Organic Carbon.

EACC East African Coastal Current.

EEZ Exclusive Economic Zone.

ERA5 ECMWF Reanalysis v5.

GEBCO General Bathymetric Chart of the Oceans.

GHG Greenhouse Gases.

GO-SHIP Global Ocean Ship-based Hydrographic Investigations Program.

LPT Lagrangian Particle Tracking.

MRV Monitoring, Reporting and Verification.

NEMC Northeast Madagascar Current.

NOAA National Oceanic and Atmospheric Administration.

NPP Net Primary Production.

POC Particulate Organic Carbon.

RAMA Research Moored Array for African-Asian-Australian Monsoon Analysis and Prediction.

RBF Radial Basis Function.

RMSE Root-Mean-Square Error.

SC Somali Current.

SEC South Equatorial Current.

SECC South Equatorial Counter Current.

SEMC Southeast Madagascar Current.

SLIM3D Second-generation Louvain-la-Neuve Ice-ocean Model.

TKE Turbulent Kinetic Energy.

UHSLC University of Hawai'i Sea Level Center.

UNFCCC United Nations Framework Convention on Climate Change.

UTide Unified Tidal Analysis Toolbox.

WIO Western Indian Ocean.

1 Introduction

1.1 Global climate context

Climate change, primarily driven by human activities, is widely regarded as one of the defining challenges of the twenty-first century. The human footprint on the Earth system has become so significant that [Paul Crutzen \(2006\)](#) proposed naming the current geological epoch the Anthropocene (literally “the new age of humans”). The main drivers are the combustion of fossil fuels, intensive agriculture and deforestation. Together, they have caused a sharp rise in atmospheric [Greenhouse Gases \(GHGs\)](#), mainly carbon dioxide (CO₂), methane (CH₄) and nitrous oxide (N₂O) ([IPCC, 2021](#)). These gases absorb and re-emit the infrared radiation leaving the Earth’s surface, trapping heat in the lower atmosphere and creating a radiative imbalance that drives global warming ([NASA, 2014](#)).

Among these consequences, the most prominent one is the rise in global mean surface temperature. Over the last century, the temperature has increased by an average of 1.1°C relative to the pre-industrial period (1850–1900) ([IPCC, 2021](#)). This deviation, referred to as the temperature anomaly, reached its highest ever recorded value of 1.55°C in 2024 ([United Nations, 2025](#)). As shown in [Figure 1](#), the warming has not only persisted but accelerated markedly over the past decades ([IPCC, 2023](#)).

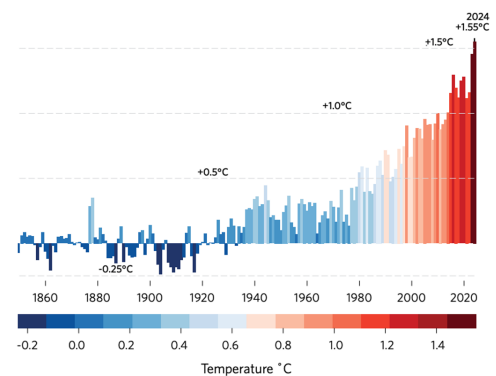


Figure 1: Global annual mean temperature relative to pre-industrial levels (1850–1900 average), showing the accelerating warming trend.

Source: ([United Nations, 2025](#))

This warming triggers profound physical and human consequences. Glaciers are melting at unprecedented rates, threatening freshwater resources and contributing to sea level rise, which increased by 20 cm between 1901 and 2018 ([IPCC, 2023](#)). Extreme weather events are intensifying worldwide, with heat waves, droughts, floods and cyclones becoming more frequent, and associated economic losses now estimated at USD 202 billion per year ([IPCC, 2023](#)). Between 3.3 and 3.6 billion people already live in highly vulnerable regions where most damages cannot be reversed, and land is being lost to sea-level rise, permafrost thaw and ecosystem degradation, threatening agricultural productivity and food security ([IPCC, 2023](#); [United Nations, 2025](#)).

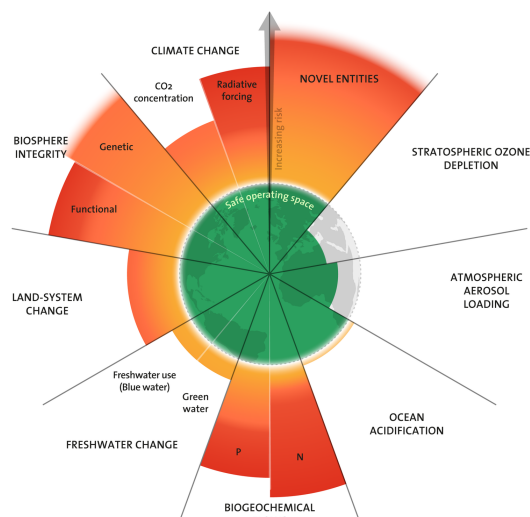


Figure 2: Planetary boundaries framework updated in 2025, showing the seven boundaries already transgressed.

Source: (Stockholm Resilience Centre, 2012)

These impacts show that the Earth system is approaching critical limits. The planetary boundaries framework, introduced in 2009, defines a safe operating space for humanity across nine key Earth system processes, each associated with a quantitative threshold beyond which large-scale, irreversible changes become increasingly likely (IPCC, 2023). The 2025 update reveals that seven of these nine boundaries have already been transgressed, including those linked to climate change (Figure 2). These boundaries are highly interconnected: exceeding one increases the risk of cascading effects on the others.

Faced with these risks, the international community adopted the Paris Agreement in 2015, signed by 195 parties (194 states and the European Union) (IPCC, 2023; UNFCCC, 2015). The agreement aims to keep global warming well below 2°C above pre-industrial levels, with maximum efforts to limit it to 1.5°C (a threshold defined as a 20-year running mean, currently estimated at 1.34–1.41°C (Climate Analytics, 2024)). In 2024, the annual temperature anomaly already reached 1.55°C (United Nations, 2025), and current trajectories suggest the 1.5°C threshold will be definitively exceeded within the next decade.

Limiting this warming requires targeting its main drivers. Figure 3 shows the contribution of each anthropogenic forcing to the observed warming of the 2010–2019 decade relative to 1850–1900 (IPCC, 2021). CO₂ is by far the dominant driver, contributing 0.80°C. Methane adds 0.51°C, nitrous oxide 0.09°C and halogenated gases 0.10°C. These positive forcings are partially offset by aerosols and other human drivers (–0.56°C), giving an observed net warming of approximately 1.07°C.

The dominant role of CO₂ is reinforced by its exceptional persistence. Once emitted, it remains in the atmosphere for centuries to millennia before being removed by natural processes (Archer et al., 2009). Today's emissions will therefore continue to drive warming long after they are released, making it essential not only to cut emissions but also to protect the natural processes that remove CO₂ from the atmosphere.

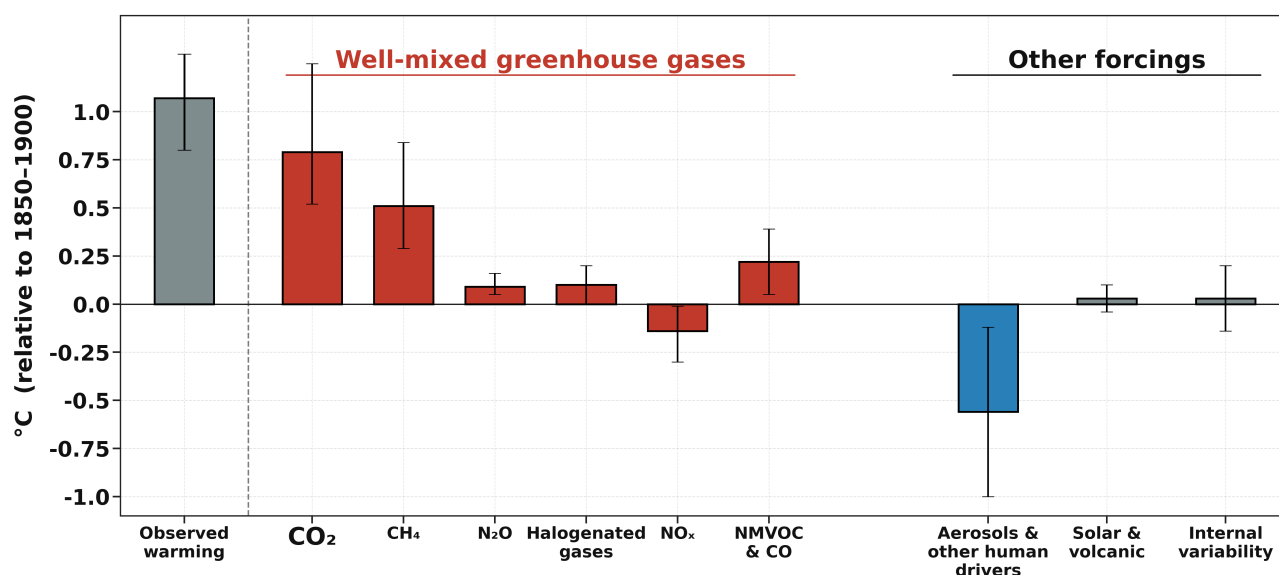


Figure 3: Contributions to observed warming (2010–2019 relative to 1850–1900). CO₂ is by far the dominant driver. Well-mixed GHGs are shown individually; aerosols and other forcings are aggregated. The sum of all contributions equals the observed warming of 1.07°C.

Adapted from IPCC (2021)

1.2 The global carbon cycle

The global carbon cycle describes the continuous exchange of carbon between the main Earth reservoirs: the atmosphere, the terrestrial biosphere and soils, and the oceans. The size of these reservoirs and the rate of exchange between them control the atmospheric concentration of CO₂, and therefore the intensity of the greenhouse effect. Understanding which reservoirs respond to anthropogenic emissions, and over what timescales, is essential to assess how the climate will evolve.

These reservoirs are commonly classified by residence time (the average time a carbon atom remains within them before being exchanged) (Ciais et al., 2013). Slow reservoirs hold carbon for 10,000 years or more in rocks and deep sediments, exchanging it naturally through volcanic emissions, chemical weathering and erosion (NASA Earth Observatory, 2011; Ciais et al., 2013). Fast reservoirs include the atmosphere, the terrestrial biosphere, soils, freshwater, and the surface ocean and its sediments, with residence times from a few years to a few millennia (Falkowski et al., 2000; Ciais et al., 2013). Human activities have now broken the natural separation between the two domains. Fossil fuel combustion mobilises slow-reservoir carbon and transfers it to the fast reservoirs, where it redistributes between the atmosphere, the biosphere and the surface ocean. These fast reservoirs therefore govern present-day climate change (Ciais et al., 2013).

Atmospheric CO₂ concentrations have always varied, but never at the current pace. On geological timescales, they have fluctuated considerably, peaking around 56 million years ago during the Paleocene–Eocene Thermal Maximum, when a sudden release of carbon triggered abrupt warming (Figure 4) (CenCO₂PIP Consortium, 2023). Closer to the present, ice cores show that CO₂ remained within 180–280 ppm for the past 800,000 years, reflecting a near-equilibrium between natural emissions and photosynthetic uptake. Since the Industrial Revolution, however, concentrations have surged from about 280 to over 420 ppm (IPCC, 2023). The current rate of increase is more than 100

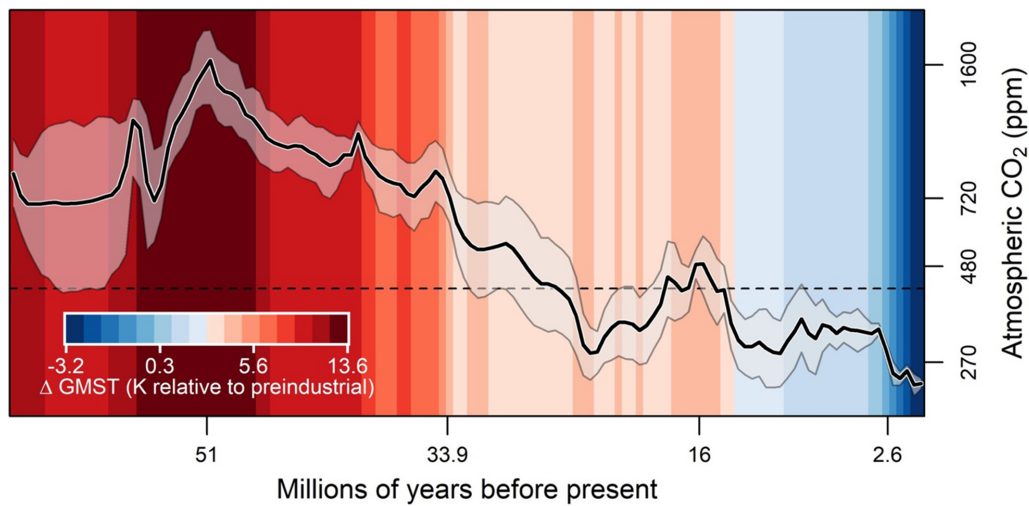


Figure 4: Atmospheric CO₂ concentration and global mean surface temperature anomaly (Δ GMST) over the past 66 million years, showing the direct coupling between CO₂ levels and temperature evolution.

Source: [University of Utah College of Science \(2023\)](#)

times faster than the most rapid natural changes recorded in those cores, which is without parallel in the geological record ([Ciais et al., 2013](#)).

This surge reflects anthropogenic emissions that have grown steadily since 1850. Over the most recent decade (2013–2022), fossil fuel combustion accounted for 88% of total emissions ($9.6 \pm 0.5 \text{ Gt C yr}^{-1}$), and land-use change for the remaining 12% ($1.3 \pm 0.7 \text{ Gt C yr}^{-1}$) ([Friedlingstein et al., 2023](#)). Natural sinks only partly compensate: terrestrial ecosystems absorb about 30% ($3.3 \pm 0.8 \text{ Gt C yr}^{-1}$) and the oceans about 26% ($2.9 \pm 0.4 \text{ Gt C yr}^{-1}$). The remaining 48% ($+5.2 \text{ Gt C yr}^{-1}$) accumulates in the atmosphere, with a small residual imbalance of -4% ($-0.4 \text{ Gt C yr}^{-1}$) attributed to uncertainties and processes not fully accounted for (Figure 5).

The capacity of these natural sinks to buffer anthropogenic emissions ultimately depends on the size of each reservoir. The oceans dominate by a wide margin, holding approximately 39,000 Gt C, of which 37,000 Gt C is dissolved inorganic carbon and 1,750 Gt C is stored in surface sediments. By comparison, soils contain 1,500–2,400 Gt C, permafrost about 1,700 Gt C, terrestrial vegetation 450–650 Gt C, and the atmosphere only about 860 Gt C. The ocean therefore holds roughly 93% of all carbon in the fast reservoirs, making it the dominant sink for anthropogenic CO₂ ([Friedlingstein et al., 2023](#)).

1.3 The ocean stock

The ocean is the largest reservoir in the fast carbon cycle, holding about 45 times more carbon than the atmosphere ([IPCC, 2007](#)). This carbon is stored in three main chemical forms. **Dissolved Inorganic Carbon (DIC)** groups dissolved CO₂, bicarbonate (HCO₃⁻) and carbonate (CO₃²⁻) ions. **Dissolved Organic Carbon (DOC)** covers small organic molecules released by biological activity. **Particulate Organic Carbon (POC)** comprises living plankton and detrital particles. Their relative weights follow a 2000:38:1 ratio between **DIC**, **DOC** and **POC**, so **DIC** holds most of the carbon,

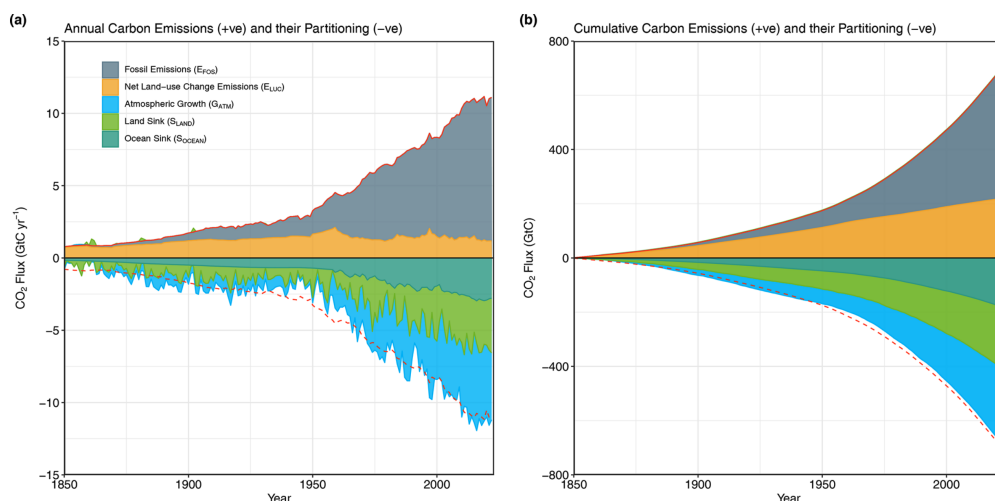


Figure 5: Global carbon budget showing emissions sources and carbon partitioning. (a) Average annual fluxes (2013–2022) dominated by fossil fuels (88%) and land-use change (12%), with partitioning between atmosphere (48%), ocean (26%) and land (30%). (b) Cumulative emissions and their distribution from 1850 to 2023, illustrating the historical evolution of anthropogenic carbon sources and sinks.

Source: [Friedlingstein et al. \(2023\)](#)

while **DOC** and **POC** drive most of the internal cycling ([IPCC, 2007](#)). This storage relies on three main mechanisms collectively known as carbon pumps (Figure 6).

The first is the solubility pump, which is purely abiotic. CO₂ is more soluble in cold water, so high-latitude surface waters absorb atmospheric carbon and sink with dense water masses, sequestering it for centuries to millennia. In the tropics, warm surface waters and upwelling reverse the flux and release CO₂ back to the atmosphere ([IPCC, 2007](#); [Heinze et al., 1991](#))(Figure 6).

The biological pump provides a second pathway. Phytoplankton fix **DIC** into **POC** and **DOC** through photosynthesis in the surface layer. Most of this organic matter is remineralised back to CO₂ within the water column, but a small fraction sinks to the seafloor and enters long-term geological sequestration ([IPCC, 2007](#); [Heinze et al., 1991](#)) (Figure 6).

The carbonate pump involves calcifying organisms that build CaCO₃ shells and skeletons. On short timescales, it acts as a counter-pump: forming each CaCO₃ molecule from two bicarbonate ions liberates one CO₂, which raises surface seawater pCO₂ and reduces the air–sea gradient that drives atmospheric CO₂ uptake. On geological timescales, however, the small fraction of CaCO₃ that escapes dissolution and is permanently buried in deep sediments removes carbon for tens of millions of years ([IPCC, 2007](#); [Heinze et al., 1991](#))(Figure 6).

These three pumps are now being perturbed by ocean acidification. Surface pH has dropped from 8.2 to 8.1 since the pre-industrial era, a shift corresponding to a 30% increase in H⁺ concentration ([IPCC, 2021](#)). Lower pH reduces carbonate ion availability, disrupting calcification in corals, molluscs and planktonic calcifiers and weakening the carbonate pump. Despite these pressures, some coastal ecosystems remain remarkably efficient carbon sinks and are the focus of the next section.

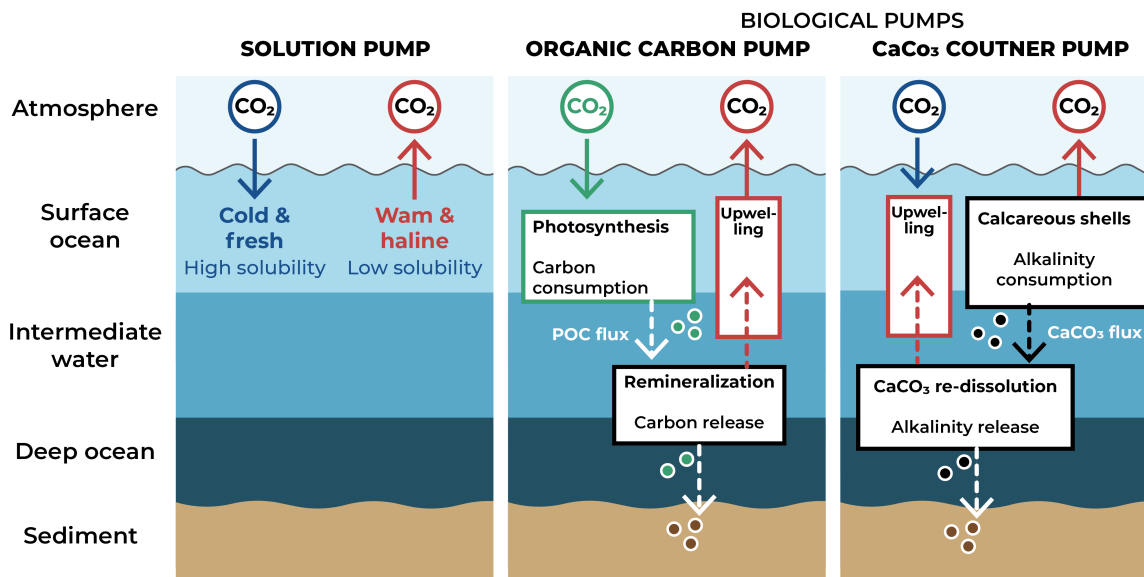


Figure 6: Overview of oceanic carbon pumps and their associated processes. The solubility pump transfers CO_2 to depth through temperature-dependent dissolution. The organic carbon pump sequesters carbon via photosynthesis, with subsequent remineralisation in intermediate waters. The carbonate counter-pump produces calcareous shells, releasing CO_2 at the surface while enabling long-term burial in sediments.

Adapted from: [Heinze et al. \(1991\)](#)

1.4 Blue carbon sink

Among these efficient coastal sinks, a specific category has emerged under the term *blue carbon*. Marine primary producers account for roughly half of global net primary production ([Falkowski et al., 2000](#); [Nellemann et al., 2009](#)), yet most of this carbon is rapidly remineralised in the upper ocean. A small fraction is sequestered for centuries to millennia in the sediments of a few coastal ecosystems. The ecosystems responsible are the **Blue Carbon Ecosystems (BCEs)** ([NOAA Climate.gov, 2022](#)).

Three **BCEs** dominate this function: tidal marshes, seagrass meadows and mangroves. Together they cover only 0.25% of the ocean surface ([Schlesinger, 1997](#)), yet their sediments accumulate carbon over thousands of years ([Lo Iacono et al., 2008](#); [McLeod et al., 2011](#)). Burial rates per unit area reach about 35 times those of tropical forests in seagrass meadows and up to 55 times in mangroves and tidal marshes ([McLeod et al., 2011](#)) (Figure 7a). The following numbers carry large uncertainties (global **BCE** area estimates vary by a factor of five, with seagrass cover ranging from 0.17 to 0.6 million km^2 ([Nellemann et al., 2009](#))), but the total **BCE** surface of 0.9 million km^2 is estimated to bury approximately 160 Tg C yr^{-1} , with seagrass beds contributing $\sim 80 \text{ Tg C yr}^{-1}$, salt marshes $\sim 46 \text{ Tg C yr}^{-1}$ and mangroves 32 Tg C yr^{-1} (Figure 7b) ([Chmura et al., 2003](#); [McLeod et al., 2011](#)). For comparison, forests cover approximately 44 million km^2 and bury a comparable 180 Tg C yr^{-1} (Figure 7c) despite a 48-fold larger footprint ([McLeod et al., 2011](#)).

These ecosystems are also rapidly disappearing. Seagrass meadow losses accelerated to about 7% per year after 1990 ([Waycott et al., 2009](#)), mangroves have declined at 0.7–3% and salt marshes at 1–2% ([McLeod et al., 2011](#)), driven by destructive fishing, coastal construction, pollution and

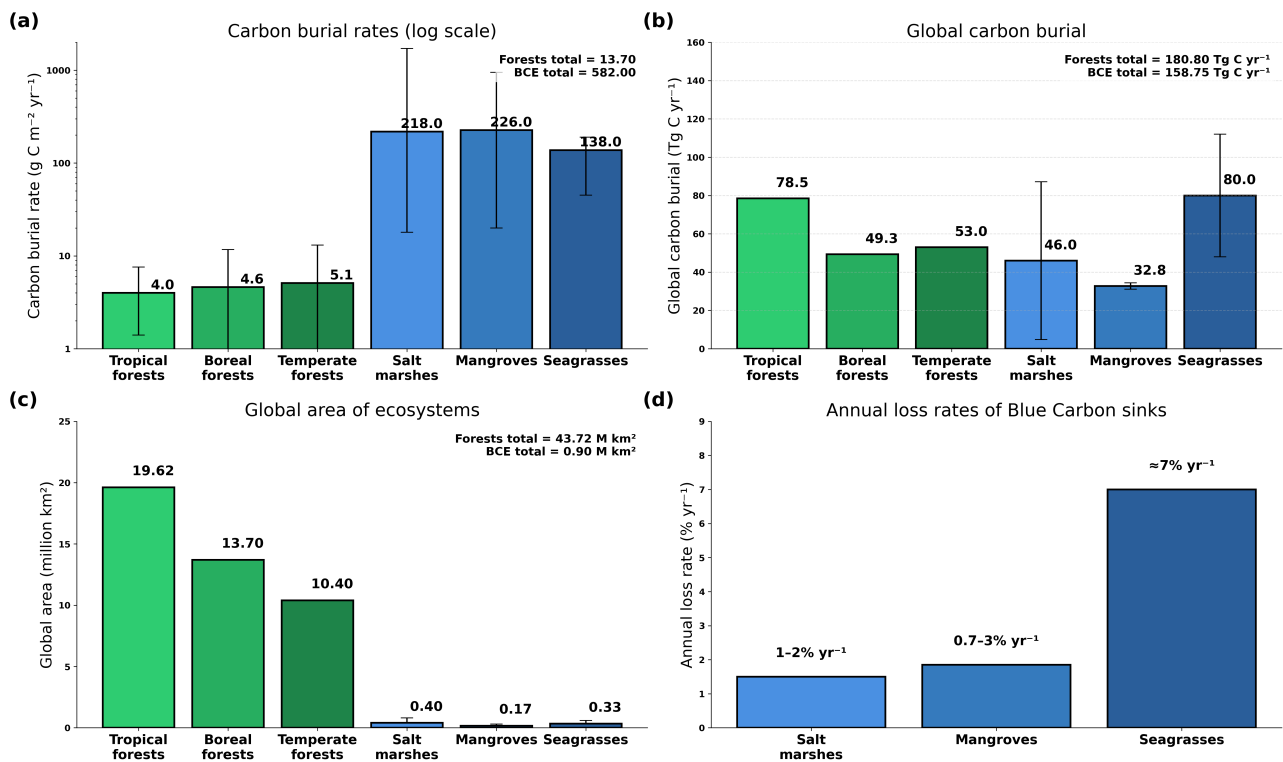


Figure 7: Comparison between forests and BCE. (a) Carbon burial rates: BCE demonstrate approximately 40-fold higher burial efficiency per unit area compared to tropical forests. (b) Global carbon burial: forests contribute 180.80 Tg C yr^{-1} versus 158.75 Tg C yr^{-1} for BCE (14% higher). (c) Global ecosystem coverage: forests occupy 48 times more surface area than BCE, explaining their comparable total sequestration despite lower per-area rates. (d) Annual loss rates: seagrasses experience the highest degradation ($\sim 7\% \text{ yr}^{-1}$), nearly 7-fold greater than salt marshes and mangroves.

adapted from: McLeod et al. (2011)

climate-driven sea level rise (Figure 7d). Because these losses are largely anthropogenic, concrete action is possible: protecting existing meadows, restoring degraded ones, and creating conditions favourable for their growth is therefore a direct lever for climate mitigation (Nellemann et al., 2009).

Among the three, seagrass meadows contribute the largest annual burial flux and are the focus of this thesis.

1.5 Seagrasses

Seagrasses are flowering marine plants that grow in sandy or muddy sediments and form underwater meadows in shallow coastal waters (Coastal Wiki, 2024) (Figure 10 shows the focal species of this thesis, *T. ciliatum*, as an illustration). With 72 species worldwide (Short et al., 2011), they deliver ecosystem services well beyond carbon storage: they act as nurseries for over 20% of the world's 25 largest fisheries by catch volume (Unsworth et al., 2019), reduce pathogenic bacteria by around 50% (Lamb et al., 2017), attenuate wave energy by around 40% (Fonseca and Cahalan, 1992), retain nutrients, sediments and pollutants (Orth et al., 2006), and support tourism and recreational economies (UNEP, 2020).

Their **Net Primary Production (NPP)** shows high interspecific variability, ranging from about 2,000 to over 25,000 kg C ha⁻¹ year⁻¹. This range reflects the functional diversity of species and their contrasting capacities to produce and potentially store blue carbon (Gomis et al., 2025).

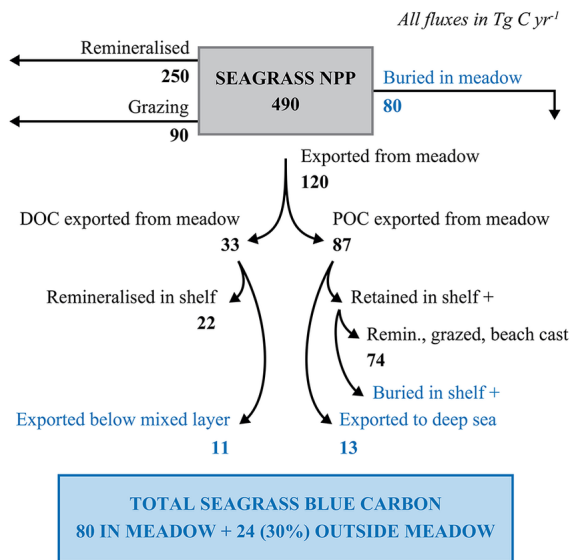


Figure 8: Fate of global seagrass **NPP** (490 Tg C yr⁻¹): in-meadow remineralisation (250), grazing (90), burial within meadows (80), and lateral export (120), the latter splitting into **DOC** (33) and **POC** (87). All values in Tg C yr⁻¹.

Source: (Duarte and Krause-Jensen, 2017)

A part of the exported **POC** is remineralised during transport. Another fraction, however, sinks to deeper layers where biogeochemical conditions strongly disfavour decomposition: oxygen depletion suppresses microbial activity (Devol and Hartnett, 2001), and low temperatures combined with limited resuspension further slow biological recycling.

Once **POC** reaches the deep ocean, conventionally below 1,000 m (Duarte and Krause-Jensen, 2017), carbon is effectively cut off from the surface and can no longer return to the atmosphere on short timescales, even in dissolved form (Sarmiento and Gruber, 2006). Its residence time is then governed by the ventilation age of the surrounding water mass (time elapsed before the ocean's overturning circulation returns it to the surface) (Primeau, 2005). Sequestration times are longest in deep-water formation regions and shortest in upwelling zones (DeVries et al., 2012), reaching several hundred years at high latitudes and approaching a millennium for carbon entrained into the abyssal overturning limb (DeVries et al., 2012; Nowicki et al., 2022). This deep-stabilised pool of exported carbon therefore constitutes an additional form of blue carbon, yet remains absent from current assessments, which account only for carbon buried directly beneath seagrass beds and thus underestimate their sequestration capacity (Duarte and Krause-Jensen, 2017).

The fate of this exported pool depends strongly on local conditions (bathymetry, seagrass distribution, dominant species, current dynamics, temporal variability), and must therefore be assessed at the local scale. This thesis focuses on the share of exported **POC** that is removed from atmospheric exchange for more than 100 years, the timescale above which carbon export

from seagrass meadows contributes to marine sequestration (Duarte and Krause-Jensen, 2017; Krause-Jensen and Duarte, 2016).

1.6 Blue carbon accounting and crediting

Quantifying the fraction of seagrass-exported POC reaching long-term sequestration reduces uncertainty on the blue carbon budget and opens a potential revenue stream. Carbon credits, representing one tonne of CO₂-equivalent removed from the atmosphere, can be counted toward Nationally Determined Contributions, transferred under Article 6 of the Paris Agreement, or traded on voluntary markets at 15–35 USD per tCO₂e (S&P Global Commodity Insights, 2025; Muri et al., 2025). No seagrass project has yet issued credits under an internationally recognised standard, although the VCS Methodology VM0033 for Tidal Wetland and Seagrass Restoration is operational and one project is undergoing verification (Farahmand et al., 2025; Verra, 2024). At the national level, the 2013 IPCC Wetlands Supplement already enables countries to include seagrass sequestration in their GHG inventories (IPCC, 2014).

Under the United Nations Framework Convention on Climate Change (UNFCCC) framework, countries report all anthropogenic removals occurring within their jurisdiction, regardless of carbon origin (IPCC, 2014). Both autochthonous carbon (produced *in situ* by the ecosystem) and allochthonous carbon (transported from external sources) can be counted if sequestered within the national territory. However, carbon exported beyond the EEZ falls outside the country's accounting boundary and cannot be claimed (Muri et al., 2025). This study therefore focuses on autochthonous carbon sequestered within Seychelles' EEZ, as this is the fraction eligible for inclusion in the national GHG inventory. Carbon leaving the EEZ is treated as a loss from the national blue carbon budget.

1.7 Seychelles

This study focuses on the Seychelles archipelago, a country that has become a pioneer in the protection and promotion of blue carbon (Palacios et al., 2022). Through its climate commitments, the nation developed the Seychelles Blue Carbon Roadmap, which aims to map blue carbon ecosystems throughout the EEZ and designate them as marine protected areas (MPAs) (Palacios et al., 2022). The country is already well advanced in its protection efforts, having conserved 99.5% of seagrass habitats within MPAs, including 7% in High Biodiversity Protection Zones (no extraction allowed) and 90% in Medium Biodiversity Protection and Sustainable Use Zones (limited extraction permitted under management) (Rowlands et al., 2024). The government has committed to reducing national GHG emissions by 26.4% by 2030 and to achieving carbon neutrality by 2050 (UNDP, 2021). As part of these efforts, seagrass carbon sequestration is being formally incorporated into the national emissions accounting under the country's updated Nationally Determined Contribution (NDC) (Palacios et al., 2022).

Since 2021, an ongoing mapping effort has been carried out under the Seagrass Mapping and Carbon Assessment project, part of the SeyCCAT (Seychelles Conservation and Climate Adaptation Trust) Coastal Wetlands and Climate Change programme (SeyCCAT, 2024; Rowlands et al., 2024). It

combined remote sensing and extensive field campaigns (Rowlands et al., 2024). SeyCCAT itself is funded by two ocean-conservation instruments: the Seychelles debt-for-nature swap (2015), which restructured roughly US\$21.6 million of sovereign debt arranged by The Nature Conservancy in exchange for marine conservation commitments (Silver and Campbell, 2018), and the world's first sovereign blue bond (2018), through which Seychelles raised US\$15 million from international investors with proceeds dedicated to ocean conservation (World Bank, 2018).

This mapping effort estimated the total area of seagrass beds at 1,599 km² and quantified their carbon stocks (Rowlands et al., 2024). The results reveal that seagrass beds dominate the country's blue carbon inventory: they contain more than 18.87 MtC, representing 96% of Seychelles' blue carbon stock, far exceeding the 0.69 MtC stored in mangroves (Rowlands et al., 2024). At the ecosystem level, the average carbon stock reaches 188 tC/ha, almost entirely in sediments beneath the meadows. The associated annual sequestration rate amounts to approximately 510,000 tCO₂e, enough to offset the emissions of the entire Seychelles energy sector and nearly three times those of the transport sector (Rowlands et al., 2024).

Beyond its ecological value, this 510,000 tCO₂e per year flux is also a tangible financial asset. Through the accounting and crediting pathways outlined in Section 1.6, it could contribute directly to the country's 2030 reduction target (−26.4% vs. business-as-usual) and 2050 net-zero commitment (Palacios et al., 2022; UNDP, 2021), and translate into additional conservation funding building on the 2015 debt-for-nature swap and the 2018 sovereign blue bond.

Yet these figures only capture the carbon buried directly beneath the meadows. They do not account for the fraction exported beyond the canopy as detached leaves and detritus, a component still missing from current blue carbon monitoring frameworks (Muri et al., 2025).

The bathymetry of the Seychelles is particularly distinctive and provides an exceptional natural laboratory for studying the dynamics of carbon export. The oceanic plateau supporting the Inner Islands is unusually wide: around Mahé it extends for nearly 200 km while remaining relatively shallow, with an average depth of about 50 m (Castillo-Trujillo et al., 2021). This configuration is favourable for seagrass development, as shallow waters (typically between 2 and 50 m for most species) provide sufficient light for photosynthesis, and the considerable width of the plateau offers a continuous space for the expansion of vast underwater meadows (Duarte, 1991). The plateau ends abruptly at its edge, where the seafloor plunges to nearly 2,000 m in less than 20 km (Castillo-Trujillo et al., 2021) (Figure 9). This geometry creates an unusually close proximity between shallow seagrass meadows and deep ocean basins (>1,000 m), and the steep continental slope facilitates the rapid transfer of exported POC towards deep waters, minimising the exposure time during which remineralisation can occur.

Given this atypical bathymetry and the extensive seagrass coverage, the exported flux could represent a significant additional contribution to blue carbon sequestration. If a larger-than-typical share of the exported POC reaches depths where sequestration becomes long-term, the climate mitigation potential of Seychelles' seagrass meadows could be substantially greater than current estimates suggest (McMahon et al., 2014; Krumhansl et al., 2025). This raises the central question of this thesis: **what fraction of the particulate organic carbon exported from *Thalassodendron ciliatum* meadows of the Seychelles reaches long-term sequestration?** Understanding this

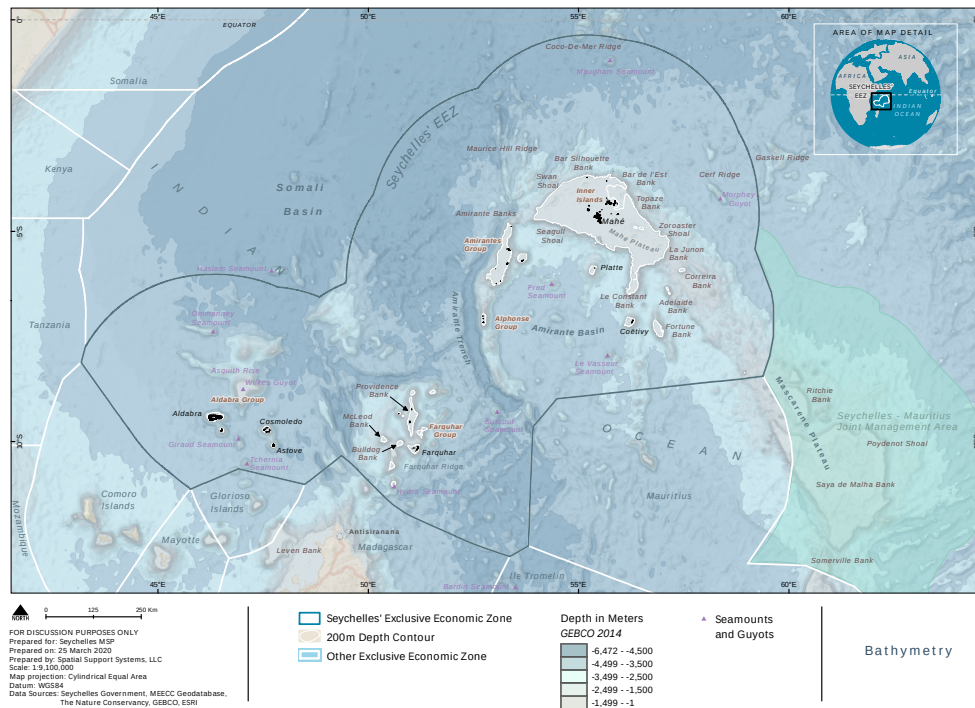


Figure 9: Bathymetric map of the Seychelles showing the extensive shallow plateau (~50 m depth) surrounding the Inner Islands and the steep continental slope transitioning to depths exceeding 4,000 m. Source: [Seychelles Marine Spatial Plan Initiative \(2020\)](#)

export pathway is essential for accurately quantifying the archipelago's contribution to climate change mitigation and may reveal that existing assessments, which account only for sediment burial ($\approx 16\%$ of **NPP**), significantly underestimate the total blue carbon sequestration provided by these ecosystems. Among the species present, *Thalassodendron ciliatum* dominates the Seychelles archipelago, forming extensive monospecific subtidal meadows ([Rowlands et al., 2024](#); [Gullström et al., 2002](#)). It is the only species considered in this thesis and provides the only source of **POC** traced by the modelling. The species is characterised by a creeping rhizome anchored in the substrate, from which upright vertical stems carry apical clusters of strap-like leaves (Figure 10a). On the Seychelles shallow shelves, these shoots aggregate into dense, canopy-forming meadows (Figure 10b).

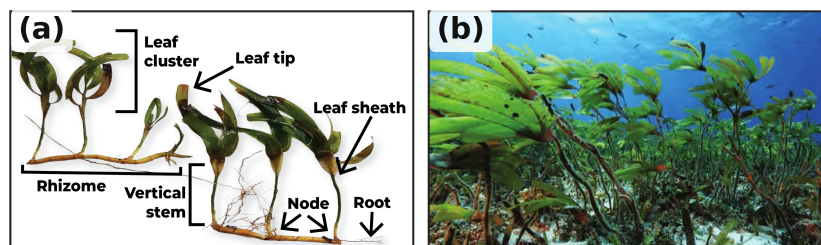


Figure 10: *Thalassodendron ciliatum*. (a) Vegetative shoot anatomy, from [Seagrass-Watch \(2024\)](#). (b) Subtidal meadow at Aldabra Atoll, Seychelles, from [Friedlander et al. \(2014\)](#).

1.8 Objectives

Existing blue carbon estimates for Seychelles seagrass meadows count only the in-meadow burial pool, leaving the exported **POC** fraction outside the national inventory. The aim of this thesis is to estimate that exported contribution, whether it ends up on the shelf seafloor or in the deep ocean, for the *T. ciliatum* meadows of the archipelago.

The driving question is therefore:

What fraction of the particulate organic carbon exported from *T. ciliatum* meadows of the Seychelles reaches long-term sequestration?

Beyond this main question, the work also aims to study how the local currents and the monsoon-reversing circulation over the Seychelles plateaus influence the trajectories of exported particles and their final fate, and whether the position of the meadows relative to the shelf break also plays a role.

2 Material

2.1 Study area

The study area is defined as the Seychelles EEZ (1,342,000 km²) extended by a 200 km smoothed buffer, for a total of 2,520,000 km² (Figure 11). This study focuses on autochthonous POC sequestered within the EEZ. The buffer ensures that moving particles exported from the meadows and crossing the EEZ boundary can be tracked further and can potentially re-enter the domain, rather than being lost at the edge.

The bathymetry of the Seychelles is particularly relevant for this work, especially around the Mahé Plateau (Figure 11b), the largest shelf of the archipelago. With a mean depth of ~50 m and a maximum extent exceeding 250 km along its longest dimension, the plateau offers the wide and shallow conditions required for extensive seagrass cover. At its edge, the seabed drops to more than 2,000 m in less than 20 km (Figure 11d) (Castillo-Trujillo et al., 2021), a sharp shelf break typical of the Seychelles archipelago.

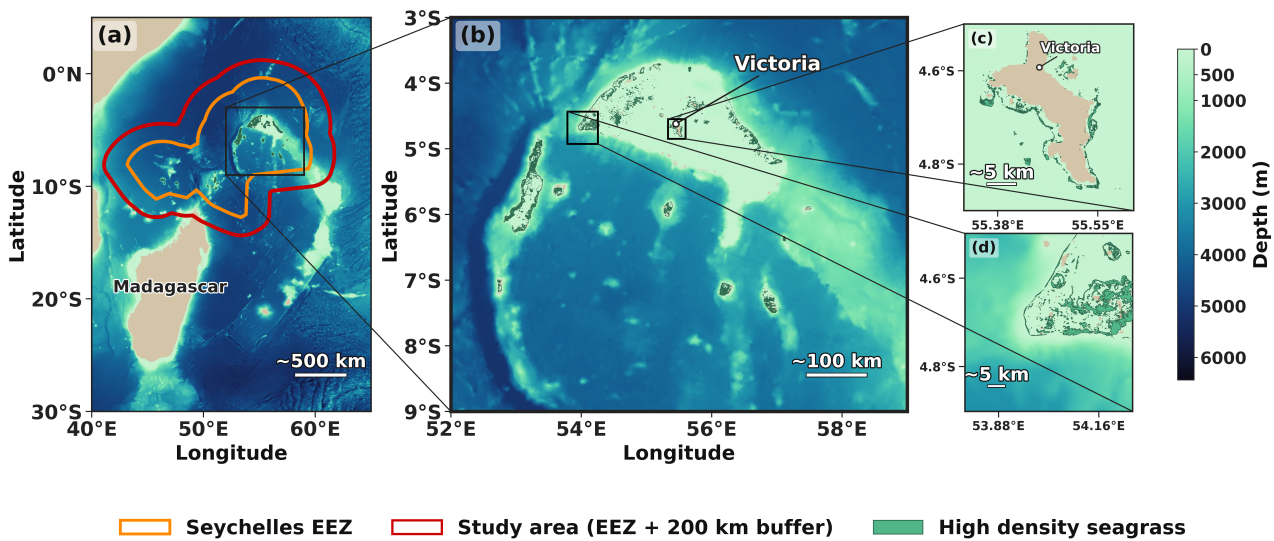


Figure 11: Bathymetry map of the Seychelles. (a) Regional context showing the Seychelles EEZ (orange outline), the study area corresponding to the EEZ extended by a 200 km buffer (red outline) and high-density seagrass distribution (green). (b) Close-up of the Mahé Plateau with Victoria (Mahé) as a reference point, showing the shallow shelf bathymetry. (c) close-up view on Mahé island, with high-density seagrass meadows along the coastline. (d) close-up view on the western edge of the Mahé Plateau, showing offshore seagrass meadows.

Bathymetry: [GEBCO Compilation Group \(2025\)](#). Seagrass: [Rowlands et al. \(2024\)](#).

The seagrass distribution used in this study comes from the countrywide benthic habitat map of [Rowlands et al. \(2024\)](#), provided by Dr Jérôme Harlay (University of Seychelles). The map was derived from Sentinel-2 and NICFI PlanetScope imagery acquired between 2015 and 2020. Classification was performed in Google Earth Engine with a Random Forest algorithm. The algorithm was trained on ~40 000 georeferenced photo quadrats collected at 16 island sites between October 2021 and April 2022 ([Rowlands et al., 2024](#)).

The benthic cover is split into three density classes (low, medium, high). Only the high-density class is retained, as its spectral signature can be reliably attributed to perennial seagrass and therefore most likely corresponds to *T. ciliatum* meadows, whereas the lower density classes are far more uncertain and tend to mix with macroalgae and other benthic organisms of similar spectral signature. All retained meadows are assumed to be monospecific *T. ciliatum* cover. The retained polygons cover 1,765 km² across the Seychelles EEZ.

Once detached from the meadows, particle trajectories depend not only on the local shelf geometry but also on the large-scale circulation of the Western Indian Ocean (WIO), where the seasonal reversal of the monsoon winds drives major changes in the surface currents (Figure 12) (Schott and McCreary, 2001). At the basin scale, the westward-flowing South Equatorial Current (SEC) splits at approximately 17°S into the northward Northeast Madagascar Current (NEMC) and the southward Southeast Madagascar Current (SEMC). The NEMC feeds the permanent East African Coastal Current (EACC), which flows northward along the African coast year-round. The Somali Current (SC), in contrast, reverses seasonally, flowing southward during the NW monsoon (Figure 12a) and northeastward during the SE monsoon (Figure 12b) (Schott and McCreary, 2001).

At the scale of the archipelago, the Mahé Plateau is embedded in a cyclonic gyre bounded by two opposing zonal currents: the eastward South Equatorial Counter Current (SECC) to the north and the westward SEC to the south (Figure 12) (L'Hégaret et al., 2018). During the SE monsoon, the surface signature of the gyre vanishes and the near-surface flow over the plateau turns predominantly westward (Schott and McCreary, 2001; Castillo-Trujillo et al., 2021). This biannual variability in

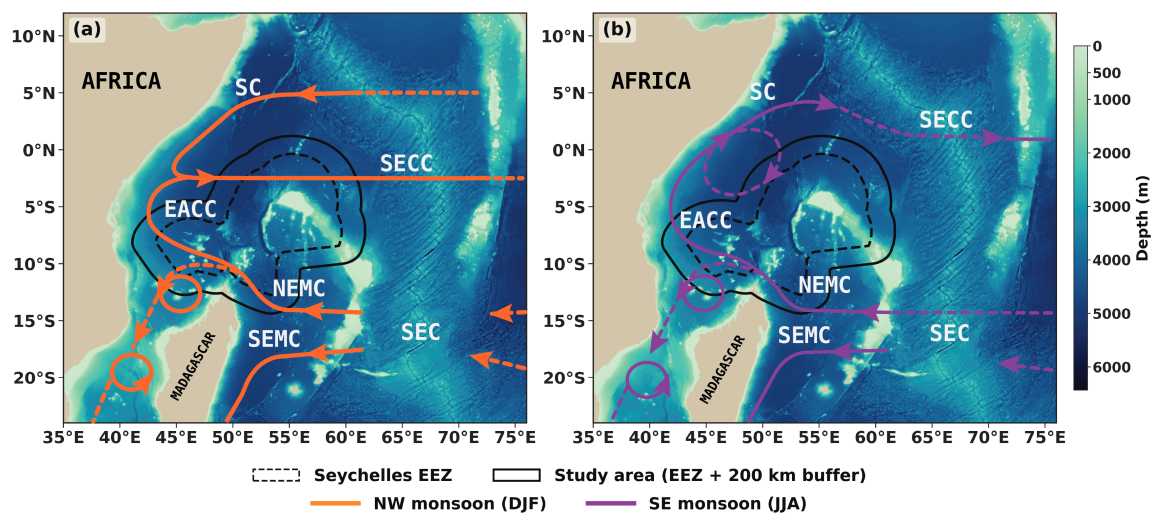


Figure 12: Regional setting and seasonal surface circulation of the WIO. GEBCO bathymetry is shown in the background. The solid black outline delimits the study area and the dashed outline the Seychelles EEZ. Main surface currents during (a) the NW monsoon and (b) the SE monsoon. The EACC flows northward year-round. The SC reverses seasonally: southward during the NW monsoon and northeastward during the SE monsoon. The Mahé Plateau lies between the SECC to its north and the SEC to its south. During the SE monsoon, the surface signature of the cyclonic gyre vanishes and the near-surface flow over the plateau turns predominantly westward (Schott and McCreary, 2001).

Adapted from Schott and McCreary (2001); L'Hégaret et al. (2018); Castillo-Trujillo et al. (2021). Bathymetry: GEBCO Compilation Group (2025).

circulation raises the question of whether monsoon-driven dynamics modulate the sequestration efficiency of carbon exported from the plateau’s meadows.

2.2 Bathymetric baseline data and correction

Using an accurate bathymetry is critical for accurately simulating the ocean circulation and the resulting dispersal patterns. The bathymetric baseline is provided by the GEBCO 2025 bathymetric grid ([GEBCO Compilation Group, 2025](#)), which covers the entire domain at a resolution of 15 arc-seconds (~ 450 m at the equator). However, the dataset contains a known artefact in the northern part of the Mahé Plateau, where it erroneously records depths above 700 m over a shelf with a mean depth of ~ 50 m ([Castillo-Trujillo et al., 2021](#)) (Figure 13b). To address this, soundings digitised from nautical charts by [Marie Jacquemin \(2024\)](#) in a previous master’s thesis on the same area (Figure 13c) were interpolated onto the mesh through a **Radial Basis Function (RBF)** scheme with a radius of influence $r = 5,000$ m and smoothly blended with the raw grid at the edges of the coverage (Appendix A). The corrected bathymetry (Figure 13d) recovers realistic plateau depths, as illustrated by its absolute deviation from the raw GEBCO grid (Figure 13e).

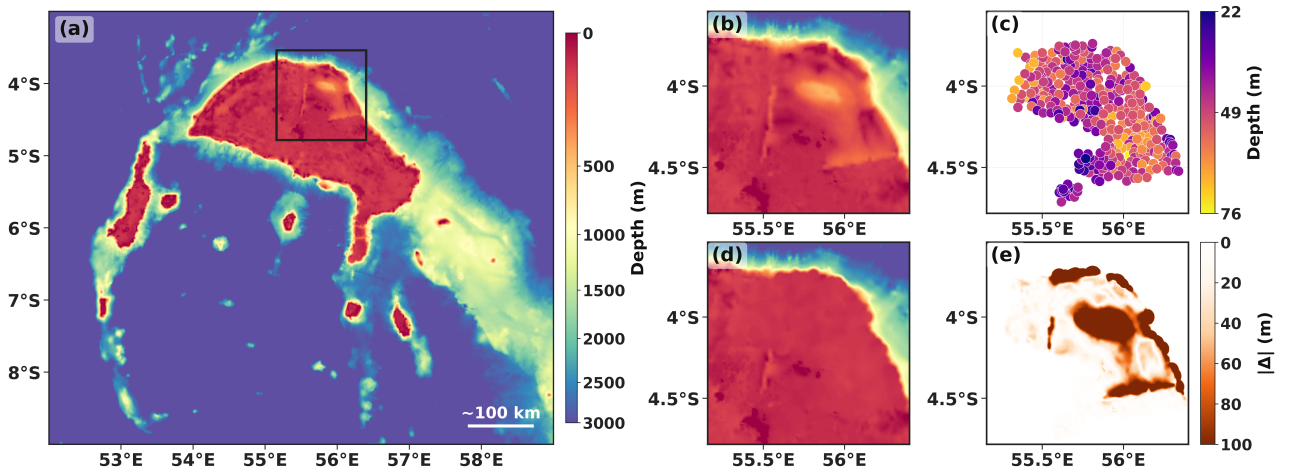


Figure 13: GEBCO bathymetry of the Mahé Plateau and **RBF** correction. (a) Full GEBCO bathymetric map, with the black rectangle outlining the anomalous sector. (b) Close-up of the anomaly. (c) Nautical chart soundings retrieved by [Marie Jacquemin \(2024\)](#) and used for the correction. (d) Corrected bathymetry after **RBF** interpolation. (e) Absolute difference between corrected and raw GEBCO bathymetry.

3 Methods

3.1 Hydrodynamic model

3.1.1 General approach

Ocean circulation in the study area is simulated using the three-dimensional baroclinic version of **SLIM3D** (SLIM team, 2026), a finite-element ocean model designed for coastal and shelf-sea applications. The mathematical formulation adopted in this study follows the model documentation of the SLIM team (2026) and Delandmeter (2017), and is presented in the sections below.

3.1.2 Domain and simulated variables

To simulate the ocean circulation, the domain must first be defined geometrically, as sketched in Figure 14. The model domain Ω is defined in three dimensions, where x and y denote the horizontal coordinates of a map projection of the Earth's surface (e.g. UTM) and z the vertical coordinate. The domain is bounded by three surfaces: fixed lateral boundaries Γ_l , a fixed bottom boundary Γ_b , and a moving free surface Γ_s . The lateral boundary Γ_l is further partitioned into closed boundaries $\Gamma_{l,c}$, corresponding to coastlines, and open boundaries $\Gamma_{l,o}$, adjacent to the open ocean (Delandmeter, 2017). In this study, both types coexist: the coastlines of Mahé and neighbouring islands form $\Gamma_{l,c}$, while the outer edges of the domain facing the open ocean form $\Gamma_{l,o}$. The bottom boundary corresponds to the seafloor, located at $z = -h(x, y)$, where $h(x, y)$ is the local bathymetry. The free surface is time-dependent and located at $z = \eta(t, x, y)$, defined relative to the reference surface at rest $\Gamma_{s,0}$ (where $\eta = 0$). The total water column depth is therefore

$$H(t, x, y) = h(x, y) + \eta(t, x, y). \quad (1)$$

Within this domain, **SLIM3D** solves for five prognostic variables (time-stepped) and two diagnostic variables (w and ρ' , derived at each time step from the prognostic fields), summarised in Table 1. In total, five governing equations (momentum, continuity, pressure and tracers) are described in the following sections, along with a supplementary equation for the computation of the density deviation (not detailed here) (SLIM team, 2026).

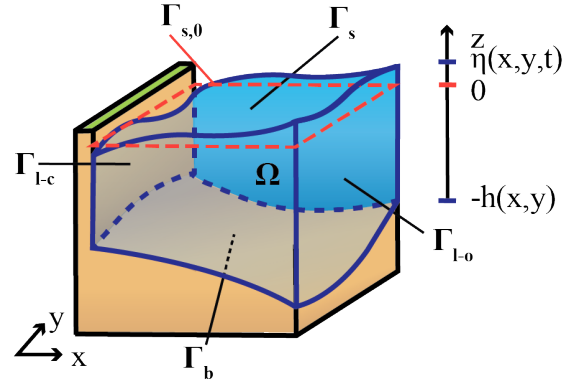


Figure 14: Sketch of the 3D model domain Ω and its boundaries: fixed bottom boundary Γ_b (seafloor), moving free surface Γ_s , and lateral boundary Γ_l partitioned into closed boundaries $\Gamma_{l,c}$ (coastlines) and open boundaries $\Gamma_{l,o}$ (open ocean).

Adapted from (Delandmeter, 2017).

Table 1: Description of the computed variables (SLIM team, 2026).

Variable	Description	Units
$\mathbf{u}(t, \mathbf{x})$	Horizontal velocity field	m s^{-1}
$w(t, \mathbf{x})$	Vertical velocity component	m s^{-1}
$\eta(t, x, y)$	Sea surface elevation	m
$S(t, \mathbf{x})$	Salinity	g kg^{-1}
$T(t, \mathbf{x})$	Temperature	$^{\circ}\text{C}$
$p(t, \mathbf{x})$	Hydrostatic pressure	N m^{-2}
$\rho'(t, \mathbf{x})$	Density deviation to reference	kg m^{-3}

3.1.3 Underlying assumptions

The governing equations of **SLIM3D** rely on two approximations: the hydrostatic hypothesis and the Boussinesq approximation (Cushman-Roisin and Beckers, 2011). The hydrostatic hypothesis assumes that the vertical pressure gradient is in balance with gravity, reducing the full Navier-Stokes vertical momentum equation to

$$\frac{\partial p}{\partial z} = -\rho g, \quad (2)$$

where g is the gravitational acceleration. The latter further assumes that density variations are small relative to a constant reference value $\rho_0 = 1000 \text{ kg m}^{-3}$, so that

$$\rho(t, \mathbf{x}) = \rho_0 + \rho', \quad (3)$$

where ρ' is a small deviation such that $\rho' \ll \rho_0$. Under this approximation, density variations are neglected everywhere in the momentum equation except in the gravity term, where they give rise to the baroclinic pressure gradients driven by horizontal density contrasts (Delandmeter, 2017).

3.1.4 Governing equations

Building on these two approximations (Eqs. (2) and (3)), we obtain the governing equations describing the ocean circulation and the evolution of the prognostic variables listed in Table 1. The horizontal momentum equation reads:

$$\underbrace{\frac{\partial \mathbf{u}}{\partial t}}_{\text{acceleration}} + \underbrace{\nabla_h \cdot (\mathbf{u}\mathbf{u}) + \frac{\partial(w\mathbf{u})}{\partial z}}_{\text{advection}} = \underbrace{\nabla_h \cdot (\nu_h \nabla_h \mathbf{u}) + \frac{\partial}{\partial z} \left(\nu \frac{\partial \mathbf{u}}{\partial z} \right)}_{\text{diffusion}} - \underbrace{f \mathbf{e}_z \times \mathbf{u}}_{\text{Coriolis}} - \underbrace{\frac{1}{\rho_0} \nabla_h p}_{\text{pressure gradient}}, \quad (4)$$

where $\nabla_h = \left(\frac{\partial}{\partial x}, \frac{\partial}{\partial y} \right)$ is the horizontal gradient operator, f the Coriolis factor, $\mathbf{e}_z$ the upward vertical unit vector, and ν_h and ν the horizontal and vertical eddy viscosities, respectively. Mass conservation under the Boussinesq approximation (Eq. (3)) reduces to the incompressibility condition:

$$\frac{\partial w}{\partial z} = -\nabla_h \cdot \mathbf{u}. \quad (5)$$

Impermeability is imposed on Γ_b and $\Gamma_{l,c}$, while the surface layer is influenced by currents and freshwater fluxes. The kinematic conditions read:

$$w - \mathbf{u} \cdot \nabla_h b = 0, \quad \mathbf{x} \in \Gamma_b, \quad (6)$$

$$\mathbf{u} \cdot \mathbf{n}_h = 0, \quad \mathbf{x} \in \Gamma_{l,c}, \quad (7)$$

$$w - \frac{\partial \eta}{\partial t} - \mathbf{u} \cdot \nabla_h \eta + s = 0, \quad \mathbf{x} \in \Gamma_s, \quad (8)$$

where $\mathbf{n}_h = (n_x, n_y, 0)$ is the outward horizontal unit normal, $s = P - E$ the net surface freshwater flux, and $b = -h(x, y)$ the bathymetry. Integrating Eq. (5) over the water column and applying the kinematic boundary conditions yields the free surface equation:

$$\frac{\partial \eta}{\partial t} = -\nabla_h \cdot \int_{-h}^{\eta} \mathbf{u} dz. \quad (9)$$

Under the hydrostatic assumption (Eq. (2)), the pressure gradient decomposes into three contributions:

$$\frac{1}{\rho_0} \nabla_h p = \underbrace{\frac{1}{\rho_0} \nabla_h p_{\text{atm}}}_{\text{atmospheric}} + \underbrace{g \nabla_h \eta}_{\text{barotropic}} + \underbrace{\frac{g}{\rho_0} \nabla_h \int_z^{\eta} \rho' dz'}_{\text{baroclinic}}, \quad (10)$$

where p_{atm} is the atmospheric surface pressure, $g \nabla_h \eta$ is the barotropic term driven by sea surface slopes, and the integral term is the baroclinic contribution arising from horizontal density gradients within the water column. The density deviation ρ' is computed from the equation of state (Jackett et al., 2006) as a function of T and S , whose transport is governed by the advection–diffusion equation, written for a generic tracer C (either T or S):

$$\underbrace{\frac{\partial C}{\partial t}}_{\text{rate of change}} + \underbrace{\nabla_h \cdot (\mathbf{u}C) + \frac{\partial(wC)}{\partial z}}_{\text{advection}} = \underbrace{\nabla_h \cdot (\kappa_h \nabla_h C) + \frac{\partial}{\partial z} \left(\kappa \frac{\partial C}{\partial z} \right)}_{\text{diffusion}}, \quad (11)$$

where κ_h and κ are the horizontal and vertical tracer diffusivities, respectively. Turbulent subscale processes are parameterised through closure schemes: ν_h follows (Smagorinsky, 1963), κ_h follows (Okubo, 1971), while ν and κ are derived from the k - ϵ turbulence closure model (Rodi, 1987). The horizontal diffusivity is computed as a function of the local mesh size Δ according to the Okubo (1971) parameterization:

$$\kappa_h = c_{\text{Okubo}} \Delta^{1.15}, \quad (12)$$

where $c_{\text{Okubo}} = 2 \times 10^{-4} \text{ m}^{0.3} \text{ s}^{-1}$ is the Okubo coefficient, which relates the diffusivity to the spatial scale of turbulent mixing.

To improve computational efficiency, the depth-integrated water transport $\mathbf{Q} = H\mathbf{u}$ is solved separately as a fast barotropic mode, governed by:

$$\frac{\partial \mathbf{Q}}{\partial t} = -gH \nabla_h \eta - \frac{H}{\rho_0} \nabla_h p_{\text{atm}} + S_{3D}, \quad (13)$$

$$S_{3D} = \nabla_h \cdot \int_b^\eta (-\mathbf{u}\mathbf{u}) dz + \int_b^\eta \left[\nabla_h \cdot (v_h \nabla_h \mathbf{u}) + \frac{\partial}{\partial z} \left(v_v \frac{\partial \mathbf{u}}{\partial z} \right) - f \mathbf{e}_z \times \mathbf{u} - \frac{g}{\rho_0} \nabla_h \int_z^\eta \rho' d\zeta \right] dz, \quad (14)$$

where S_{3D} aggregates all three-dimensional contributions from the internal baroclinic mode.

3.1.5 Boundary conditions

The governing equations are closed by boundary conditions at the three domain boundaries shown in Figure 14: the sea surface Γ_s , the seafloor Γ_b , and the lateral boundaries Γ_l . Heat and freshwater exchanges with the atmosphere are imposed as Neumann conditions at Γ_s . The temperature and salinity fluxes read:

$$\kappa_v \frac{\partial T}{\partial z} = -\frac{Q_{\text{sens}} + Q_{\text{lat}} + Q_{\text{LW}} + Q_{\text{SW}}}{\rho C_p} - Q_{\text{nudg}}, \quad (15)$$

$$\kappa_v \frac{\partial S}{\partial z} = -S(P - E) - Q_{\text{nudg}}. \quad (16)$$

where κ_v is the vertical diffusivity, C_p the specific heat capacity of seawater, Q_{sens} , Q_{lat} , Q_{LW} , Q_{SW} the sensible, latent, longwave, and shortwave heat fluxes, respectively, and P the precipitation rate prescribed at the surface. The nudging term $Q_{\text{nudg}} = \Delta z_s (C_0 - C)/\tau$ relaxes the surface tracer toward a reference value C_0 over a timescale τ . Evaporation is defined as $E = -Q_{\text{lat}}/(\rho L_e(T))$, where $L_e(T)$ is the temperature-dependent latent heat of vaporization (Henderson-Sellers, 1984). Bulk formula corrections further account for air-sea flux feedbacks (Barnier et al., 1995). Free surface deformation is tracked through an Arbitrary Lagrangian-Eulerian formulation (Donea et al., 2004). A full-slip condition is applied on $\Gamma_{l,c}$. On the open lateral boundaries $\Gamma_{l,o}$, velocity, temperature, salinity and sea surface elevation are prescribed from the CMEMS GLORYS12 (Lellouche et al., 2021) reanalysis (Section 3.3). Bottom and surface stresses enter the momentum equation as (SLIM team, 2026; Delandmeter, 2017):

$$v_v \frac{\partial \mathbf{u}}{\partial z} = \frac{\boldsymbol{\tau}_b}{\rho_0}, \quad \mathbf{x} \in \Gamma_b, \quad (17)$$

$$v_v \frac{\partial \mathbf{u}}{\partial z} = \frac{\boldsymbol{\tau}_s}{\rho_0}, \quad \mathbf{x} \in \Gamma_s. \quad (18)$$

The bottom stress follows a logarithmic drag law:

$$\frac{\boldsymbol{\tau}_b}{\rho_0} = C_d \|\mathbf{u}_{bf}\| \mathbf{u}_{bf}, \quad C_d = \left(\frac{\kappa}{\ln((z_{bf} - b)/z_0)} \right)^2, \quad (19)$$

where $\kappa \simeq 0.41$ is the von Kármán constant, z_{bf} the vertical coordinate of the bottom layer midpoint, $\mathbf{u}_{bf}$ the velocity at z_{bf} , and z_0 the bottom roughness length. The surface wind stress reads:

$$\boldsymbol{\tau}_s = \rho_{\text{air}} C_w \|\mathbf{u}_{10}\| \mathbf{u}_{10}, \quad (20)$$

where ρ_{air} is the air density, $\mathbf{u}_{10}$ the wind velocity at 10 m, and $C_w = 10^{-3}(0.63 + 0.066 \|\mathbf{u}_{10}\|)$ the drag coefficient (Smith and Banke, 1975; Cushman-Roisin and Beckers, 2011).

3.2 Spatial and temporal discretization

3.2.1 Temporal discretization

Time integration relies on a split-IMEX Runge-Kutta scheme (Ishimwe et al., 2025) that separates the dynamics into a fast external (barotropic) mode and a slower internal (baroclinic) mode. The barotropic mode is a 2D depth-averaged transport equation that resolves the fast surface gravity waves at a small time step, while the full 3D baroclinic motions are advanced with a larger time step and their contribution is passed to the 2D equation as a source term. Several short barotropic sub-steps are performed within each longer baroclinic step. This mode-splitting greatly reduces the computational cost compared with applying the small barotropic time step to the full 3D system. For the present simulation, this yields a baroclinic time step $\Delta t_{bc} \approx 8.3$ s and a barotropic sub-step $\Delta t_{baro} \approx 1.0$ s, that is, about eight barotropic sub-steps per baroclinic step.

3.2.2 Spatial discretization and mesh generation

SLIM3D solves the governing equations with a Discontinuous Galerkin Finite-Element (DG-FE) scheme and a $P_1^{DG} - P_1^{DG}$ discretization on triangular prisms (SLIM team, 2026; Delandmeter, 2017). Each prism carries six independent nodal values (three at the bottom, three at the top, Figure 15), giving a linear variation in all three directions. At shared corners, neighbouring prisms therefore hold two distinct values (Figure 15, points A and A'), and the resulting jump is resolved by numerical fluxes at element interfaces, ensuring element-wise conservation of mass and tracers, a property well suited to advection-dominated ocean dynamics.

Running the simulation requires discretizing the domain into a 3D mesh, built from a horizontal surface triangulation and a stack of vertical layers in depth. The horizontal mesh (Figure 16) contains 297,238 triangles and is produced with GMSH (Geuzaine and Remacle, 2009) via the seamsh library (Lambrechts, 2020). At each point of the domain, the local element size is set as the minimum of three size functions (Table 2),

$$\text{size}(\mathbf{x}) = \min(\text{size}_{\text{coast}}(\mathbf{x}), \text{size}_{\text{seagrass}}(\mathbf{x}), \text{size}_{\text{grad}}(\mathbf{x})), \quad (21)$$

two of which grow linearly with the distance to the coastline and to the seagrass meadow boundaries (Appendix B), while the third refines the mesh wherever the bathymetric gradient is steep (Appendix C) to resolve the abrupt shelf break of the Mahé Plateau. Together, these three criteria yield a mesh that refines down to roughly 150 m along the Mahé coastline and the seagrass meadows (Figure 16c), stays around 1,000 m near the shelf break (Figure 16d), and coarsens to about 10 km in the open ocean (Figure 16a,b).

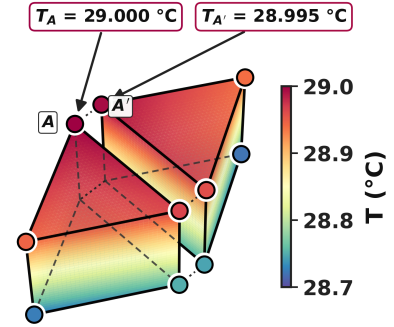


Figure 15: Example illustration of the temperature field on two adjacent P_1^{DG} prisms with six nodal values each. Points A and A' share a location but belong to different elements, materialising the DG jump.

Table 2: Mesh size function parameters.

Size function	Threshold	Scale	size _{min} (m)	size _{max} (m)
Coastline distance	400–15 000 m	linear	1 000	20 000
Seagrass distance	100–5 000 m	linear	400	7 000
Bathymetric gradient	$ \nabla h > 10^{-5}$	$0.8 h/ \nabla h $	1 600	15 000

Once the surface mesh is fixed, it is vertically extruded into prismatic columns to form the 3D domain (SLIM team, 2026). Three vertical coordinate systems are available in SLIM3D: z -layers at fixed depths (Figure 17a), σ -layers following the bathymetry (Figure 17b), and a hybrid of both (Figure 17c). The z -system ensures numerical stability but represents the seafloor as a staircase of discrete steps, degrading near-bottom dynamics. The σ -system conforms smoothly to the seafloor but becomes strongly tilted over steep gradients, introducing pressure gradient errors mitigated in SLIM3D by a preliminary bathymetry smoothing step (Haney, 1991).

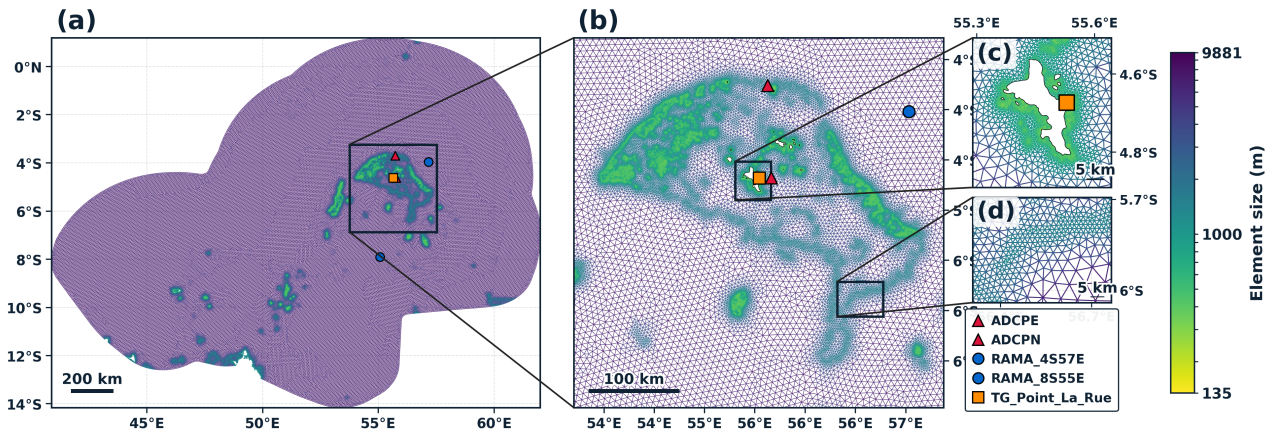


Figure 16: Horizontal mesh over the Mahé Plateau, coloured by element size. (a) Full domain with validation stations (ADCP, RAMA moorings, tide gauge). (b) Close-up of north-east Mahé showing coastal and seagrass refinement. (c) Southern shelf break with gradient-based refinement.

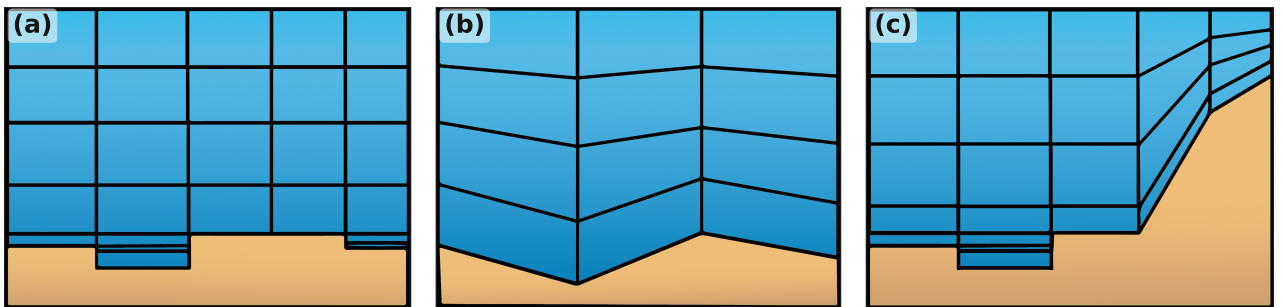


Figure 17: Vertical layer systems in SLIM3D: (a) z -layers, (b) σ -layers, (c) hybrid σ/z adopted here. Adapted from SLIM team (2026).

The hybrid configuration is adopted to resolve the Mahé Plateau accurately while keeping numerical stability at depth. The upper 125 m are covered by 10 σ -layers following the bathymetry, and 12 z-layers of exponentially increasing thickness extend below down to the seafloor. The resulting 3D mesh is shown in Figure 18, where the σ/z interface (red line) closely tracks the shelf break of the Mahé Plateau.

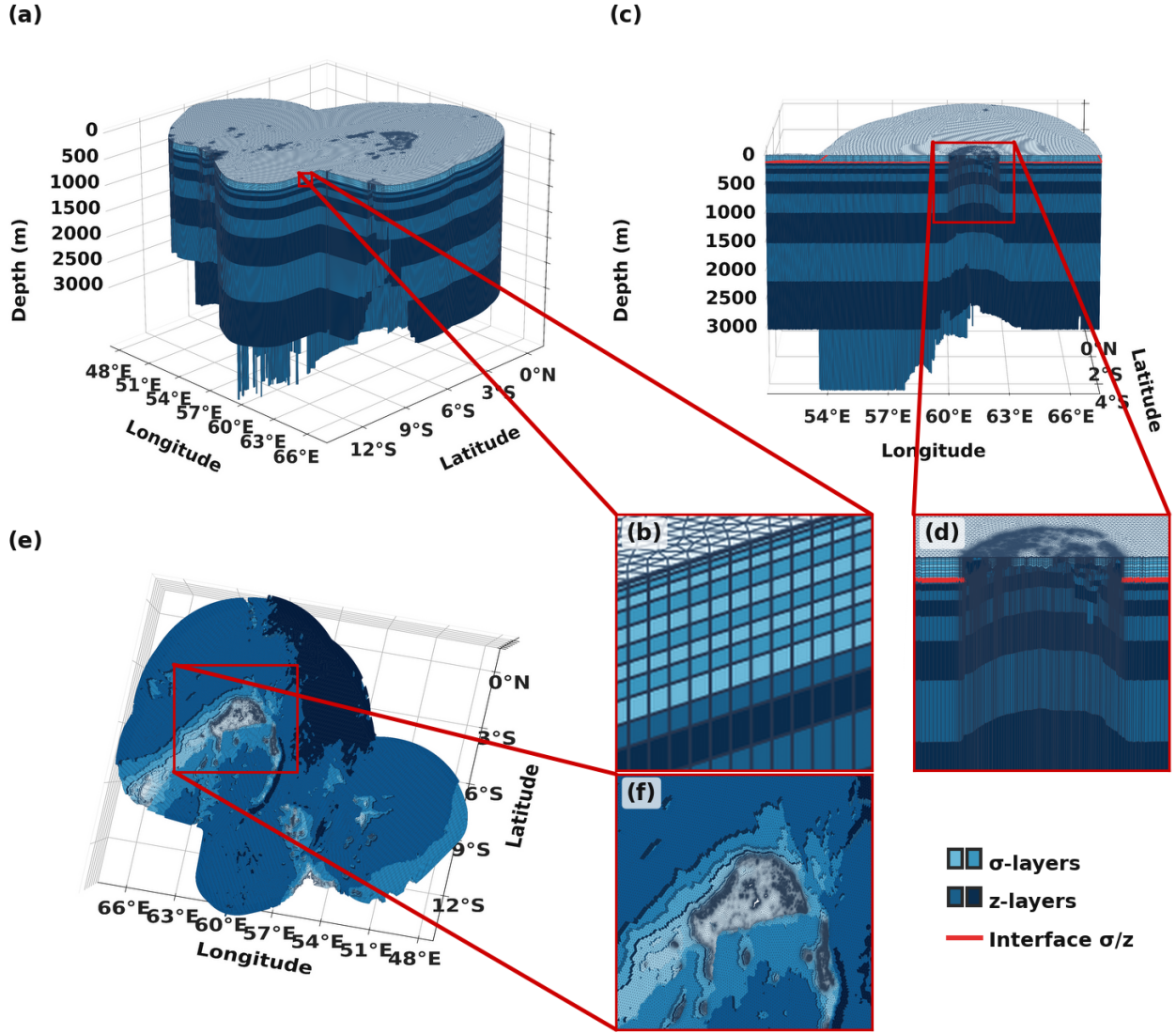


Figure 18: Hybrid σ/z mesh of the study domain. (a) Full 3D view. (b) close-up view on the σ -layers (light blue) and z-layers (dark blue). (c) Cross-section through the 3D mesh. (d) close-up view on the Mahé Plateau in cross-section. (e) View from below. (f) close-up view on the Mahé Plateau seen from below.

3.3 Model configuration and forcing

The 3D hydrodynamic simulation runs from 13 January 2016 to 16 July 2020, with hourly outputs. To ensure a numerically stable startup, external forcings are progressively ramped up over the first day. The subsequent three months serve as a spin-up period, during which the model fields adjust from uniform initial conditions to a dynamically consistent state. Particle seeding therefore begins on 13 April 2016, three months after the hydrodynamic start. A four-month buffer is similarly reserved at the end of the period, allowing particles released near its close to be tracked to their final destinations by the Lagrangian model (Section 3.4). The simulation is driven by three categories of external forcing, summarised in Table 3.

Table 3: External forcing datasets used to drive the **SLIM3D** simulation.

Forcing	Dataset	Type	Resolution	Variables
Ocean boundaries	GLORYS12	Reanalysis	1/12° (~8 km)	u, T, S, η
Atmosphere	ERA5	Reanalysis	~31 km, hourly	τ_s , heat fluxes, P
Tides	TPXO10	Inverse model	1/30°	Tidal elevation and velocity

GLORYS12: (Lellouche et al., 2021). ERA5: (Hersbach et al., 2020). TPXO10: (Egbert and Erofeeva, 2024)

Ocean lateral boundary conditions for velocity, temperature, salinity, and sea-surface elevation are prescribed from the CMEMS GLORYS12 reanalysis (Lellouche et al., 2021). At 1/12° (~8 km), GLORYS12 does not resolve shelf-scale circulation around the Mahé Plateau, but the **SLIM3D** mesh (Section 3.2) refines these fields locally down to about 150 m. A surface relaxation of temperature and salinity over the upper 10 m further keeps the model consistent with observed large-scale conditions throughout the simulation. Atmospheric forcing comes from the ERA5 reanalysis (Hersbach et al., 2020) at hourly frequency, supplying wind stress, surface heat fluxes, and precipitation. Finally, since GLORYS12 does not resolve tidal dynamics, barotropic tides are introduced separately by imposing harmonic constituents from the TPXO10 atlas (Egbert and Erofeeva, 2024) at the open boundaries.

3.4 Lagrangian particle tracking

To simulate the transport of seagrass detritus from the Mahé Plateau, a **LPT** model is driven by the three-dimensional velocity fields from the **SLIM3D** simulation. Virtual particles, each carrying a set of physical and behavioural attributes, are advected through these fields to quantify blue carbon export pathways.

3.4.1 Governing equations

The three-dimensional position of each particle evolves according to a stochastic advection-diffusion equation combining deterministic advection by the local current, a diffusivity-gradient drift, and turbulent mixing at subgrid scales.

In the horizontal plane,

$$d\mathbf{X}(t) = \underbrace{(\mathbf{u} + \alpha \mathbf{U}_{\text{wind}} \delta_{\text{surf}}) dt}_{\text{advection}} + \underbrace{\overbrace{\nabla \kappa_h dt}^{\text{deterministic}} + \overbrace{\sqrt{2 \kappa_h} dt \mathbf{R}(t)}^{\text{stochastic}}}_{\text{diffusion}}, \quad (22)$$

where $\mathbf{u} = (u, v)$ is the horizontal velocity interpolated from the **SLIM3D** output, $\mathbf{U}_{\text{wind}}$ the 10 m wind velocity, α the windage coefficient (Section 3.4.2), and δ_{surf} a surface indicator equal to 1 within 1 m of the free surface η and 0 otherwise. $\mathbf{R}(t) = (r_x, r_y) \sim \mathcal{U}(-1, 1)$ are independent uniform random variables. The deterministic drift $\nabla \kappa_h$ prevents artificial particle accumulation in regions of low diffusivity (Spagnol et al., 2002). The horizontal diffusivity follows $\kappa_h = c \Delta^{1.15}$ (Okubo, 1971), where Δ is the local mesh element size and $c = 2 \times 10^{-4}$.

In the vertical,

$$dz(t) = \underbrace{(w + w_p) dt}_{\text{advection}} + \underbrace{\overbrace{\frac{\partial \kappa_v}{\partial z} dt}^{\text{deterministic}} + \overbrace{\sqrt{2 \kappa_v(z)} dt r_z}_{\text{stochastic}}}_{\text{diffusion}}, \quad (23)$$

where w is the vertical velocity from the **SLIM3D** output, w_p the biological particle velocity (positive upward when rising, negative when sinking), $\kappa_v(z)$ the vertical eddy diffusivity computed from the k - ϵ turbulence closure (Section 3.3), and r_z an independent random draw defined as in Eq. (22). The correction $\partial \kappa_v / \partial z$ plays the same anti-diffusion role near the surface and the seabed, where κ_v changes sharply. All biological velocity terms and their parameterisation are detailed in Section 3.4.2. The particle tracking time step is $\Delta t = 300$ s.

3.4.2 Particle lifecycle and fates

Each particle is released, follows a buoyancy-driven lifecycle through up to four phases (rising, floating, sinking, decomposition), and ends in one of five mutually exclusive fates (Figure 20). The five fates are split between four that constitute the blue carbon pool (deep-ocean sequestration inside the **EEZ**, deep-ocean sequestration beyond the **EEZ**, lateral export, seafloor sedimentation) and one that does not (coastal stranding). The rest of this section details the lifecycle parameters and the criteria triggering each fate.

Each numerical particle represents one day of detached *T. ciliatum* fragments released from a small portion of seagrass meadow. Particles are seeded within the mesh triangles that intersect the high-density seagrass layer (mapped in Section 2.1). Each triangle receives 100 particles per km^2 (one per hectare), weighted by the seagrass fraction ξ_e of the triangle, with a minimum of one particle per triangle. With this ξ -weighted seeding, each particle represents an effective area of about $11,253 \text{ m}^2$ (weighted mean), and the seeded domain covers about $1,765 \text{ km}^2$ of actual seagrass across the Seychelles archipelago, totalling roughly 173,000 particles per day (172,787 reported by the seeder), about 253 million over the 3.5-year seeding window (July 2016 to January 2020), released at a random time. Each particle is released at a random horizontal position within its triangle and at a random height within the local canopy h_c (Rowlands et al., 2024) (Figure 20, step 1). Each particle

represents a fraction of the meadow area within its triangle and carries the daily carbon export from that fraction (Section 3.4.3).

Once released, each particle follows a buoyancy-driven trajectory. Freshly detached fragments contain air-filled cavities (lacunae) in both their leaves and stems, which confer positive buoyancy (McMahon et al., 2014). This is represented by setting $w_p > 0$ in Eq. (23), with w_p between 1×10^{-2} and $3 \times 10^{-2} \text{ m s}^{-1}$ estimated from a buoyancy–drag balance (Appendix D). The particle ascends until it reaches the surface layer (Figure 20, step 2). During the subsequent floating phase (Figure 20, step 3), w_p is assumed to remain at its initial value (between 1×10^{-2} and $3 \times 10^{-2} \text{ m s}^{-1}$) to prevent vertical turbulent mixing from artificially dragging the particle back below the surface. At the surface, wind also acts on the emerged portion of each fragment, adding a wind-driven drift to the surface current (windage). In the absence of species-specific data for *T. ciliatum*, this windage is inferred from the six-group leaf-shape classification of Hanuise et al. (2025). Based on its leaf dimensions (7–15 cm long, 6–13 mm wide (den Hartog, 1970)), *T. ciliatum* is assigned to group G6, corresponding to short and thick ribbon-like leaves. Each floating fragment is therefore advected by the surface current plus a wind-driven velocity equal to a fraction α of the 10 m wind speed, with α drawn from a uniform distribution $\mathcal{U}(0.00881, 0.01313)$ reflecting the G6 range reported by (Lai et al., 2020; Hanuise et al., 2025). As gases gradually escape from the lacunae, the fragment loses buoyancy and eventually starts to sink, switching from positive to negative buoyancy. This transition follows the daily rates of group G6: 0.02 day^{-1} during the first week and 0.035 day^{-1} thereafter (Figure 19, (Hanuise et al., 2025)). The original G6 fragment-survival decay is not retained here, since this study tracks carbon mass rather than fragment integrity. Mass loss is instead handled by the depth-dependent decomposition model (Eq. (24)).

Once negatively buoyant, the particle sinks through the water column (Figure 20, step 4) at a velocity $w_s \sim \mathcal{N}(7.5 \times 10^{-3}, 2.0 \times 10^{-3} \text{ m s}^{-1})$, taken from *Zostera muelleri*, the only seagrass with published fragment settling data (Weatherall et al., 2016). Sensitivity experiments span 6×10^{-3} to $6 \times 10^{-2} \text{ m s}^{-1}$ to bracket cross-species variability (Oldham et al., 2014). During descent, the particle loses mass to decomposition. The only species-specific measurement for *T. ciliatum* comes from a Kenyan

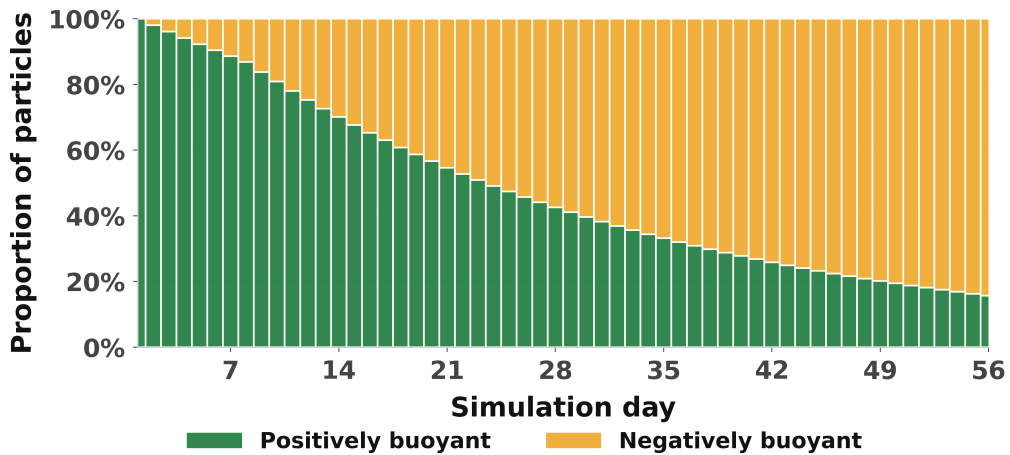


Figure 19: Buoyancy state evolution over 56 days using the G6 transition rates of Hanuise et al. (2025). Only the buoyancy transitions are retained from the original G6 framework. The viability decay is replaced by a continuous, depth-dependent decomposition model (Eq. (24)).

litterbag experiment, where stranded leaves decayed at $k = 0.017 \text{ d}^{-1}$ (half-life ≈ 41 days) (Ochieng and Erftemeijer, 1999). Because those leaves decayed in air rather than seawater, this rate likely overestimates the in-water value and is taken as an upper bound. The cross-species seagrass average compiled by Duarte and Krause-Jensen (2017) ($\approx 0.006 \text{ d}^{-1}$) provides a lower bound. Each particle is therefore assigned an initial decay rate $k_0 \sim \mathcal{U}(0.006 \text{ d}^{-1}, 0.017 \text{ d}^{-1})$, spanning these two bounds.

The depth dependence of k has never been measured for seagrass POC, but for sinking marine particles it decreases with depth as temperature drops (Laufkötter et al., 2017), a pattern captured by Martin-curve parameterisations. The same depth attenuation is hypothesised here for seagrass POC. The first-order exponential decay conventionally applied to seagrass litter (Vonk and Stapel, 2008) is retained, but k decreases with depth rather than staying constant,

$$\frac{dm}{dt} = -k(z) m, \quad k(z) = k_0 \exp(-\lambda |z|), \quad (24)$$

where λ controls how k declines with depth. Its value is calibrated against Martin curve fits to sinking POC observations in the same thermal regime, which give about two orders of magnitude of flux attenuation between the surface and 1,000 m (Marsay et al., 2015; Cram et al., 2018). Assuming k inherits the same fractional decline, $k(1,000 \text{ m}) = 0.01 k_0$ gives the central value $\lambda_0 = \ln(100)/1,000 \text{ m} \approx 4.6 \times 10^{-3} \text{ m}^{-1}$, a decay length of 217 m. To reflect literature variability in Martin curve exponents, a normal distribution $\lambda \sim \mathcal{N}(\lambda_0, 0.4 \lambda_0)$ ($\pm 40\%$ standard deviation) is assigned to λ , and a uniform distribution $k_0 \sim \mathcal{U}(0.006 \text{ d}^{-1}, 0.017 \text{ d}^{-1})$ to k_0 . The simulation itself runs with the central values $k_0 = 0.0115 \text{ d}^{-1}$ and $\lambda = \lambda_0$ applied uniformly to every particle, and the parameter uncertainty is propagated analytically through Eq. (24) a posteriori.

Below 1,000 m, decomposition is no longer tracked. Water mass ventilation ages in the tropical Indian Ocean exceed a century at this depth (Matsumoto, 2007; Primeau, 2005; DeVries et al., 2012), so any carbon crossing the threshold counts as sequestered on climatically relevant timescales, matching the operational blue carbon definition of Carlos Duarte and Krause-Jensen (2017).

The residual mass m_{final}/m_0 of each particle is attributed to one of five mutually exclusive fates (Figure 20) (Nellemann et al., 2009). Four of them are assumed to constitute the blue carbon pool. Deep-ocean sequestration (*i*, *ii*) is triggered when a particle crosses 1,000 m depth (Duarte and Krause-Jensen, 2017), with the in-EEZ share counted as (*i*) and the beyond-EEZ share as (*ii*). Lateral export (*iii*) is triggered when a particle crosses an open lateral boundary. Seafloor sedimentation (*v*) is triggered when a sinking particle lies within 1 m of the seabed and the local current remains below $U_c = 4.5 \times 10^{-2} \text{ m s}^{-1}$ (van der Mheen et al., 2024) for seven consecutive days (Krumhansl et al., 2025). The remaining fate is assumed not to contribute to sequestration. Coastal stranding (*iv*) occurs when a particle contacts a closed lateral boundary, after which the carbon is assumed to decompose at the surface. Non-blue carbon also includes the mass lost to decomposition along each particle's trajectory.

Among the four blue carbon fates, lateral export (*iii*) is the only one not credited to the Seychelles' stock, since the carbon leaves the EEZ. Its destination nonetheless remains scientifically relevant, so its vertical lifecycle is completed offline to estimate the contribution to the global blue carbon budget. Each exported particle is propagated using the floating duration t_{buoy} and sinking velocity

w_s drawn at seeding. A particle still positively buoyant at exit first finishes its remaining floating time at the surface and then sinks to $z_{\text{seq}} = 1,000$ m. A negatively buoyant particle sinks directly to z_{seq} . Decomposition (Eq. (24)) continues throughout both phases, and the residual mass at z_{seq} is attributed to (iii). The choice of $z_{\text{seq}} = 1,000$ m as the sequestration horizon is defensible because no coastline lies within 1,000 km of the study area and 96 % of the neighbouring seafloor exceeds this depth (mean 4,099 m, Figure 21) (GEBCO Compilation Group, 2025).

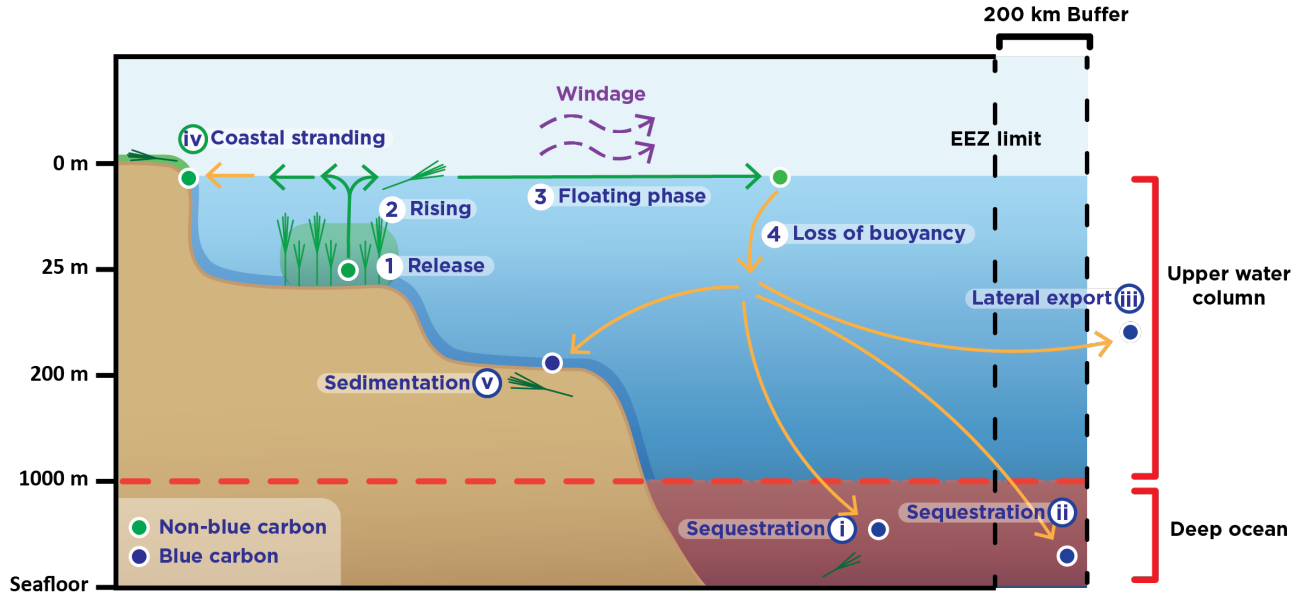


Figure 20: Conceptual diagram of the LPT model showing the four phases of the particle lifecycle: release (1), rising (2), floating (3), and loss of buoyancy (4). Terminal fates are: (i) deep-ocean sequestration below 1,000 m inside the EEZ, (ii) deep-ocean sequestration beyond the EEZ, (iii) lateral export across the open boundary, (iv) coastal stranding, and (v) seafloor sedimentation on the shelf. The red dashed line marks the 1,000 m sequestration threshold. Blue dots indicate blue carbon fates (i, ii, iii, v). Green dots indicate non-blue carbon fate (iv).

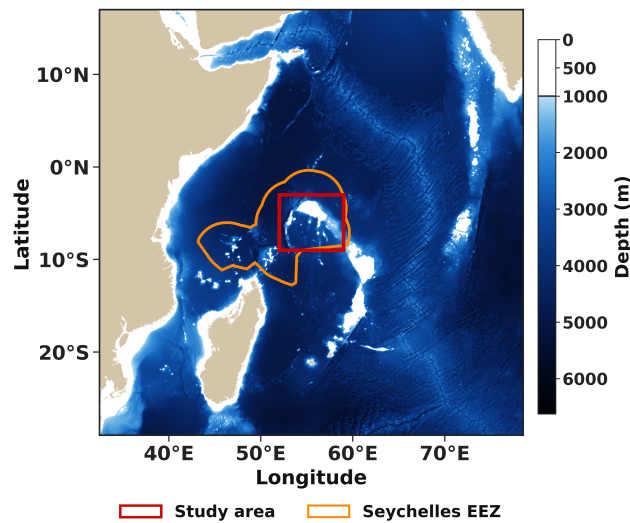


Figure 21: Bathymetry of the western Indian Ocean GEBCO Compilation Group (2025). Areas deeper than 1,000 m are highlighted in blue and shallower areas in white. Around the study area (red rectangle), most of the surrounding ocean is deeper than 1,000 m and very few coastlines are present within reach. The orange line marks the Seychelles EEZ.

All model parameters are summarised in Table 4. When a published standard deviation or observational range is available (w_s , windage, λ , U_c), it is used directly. When the parameter is extrapolated from a morphological proxy or derived from a physical balance (w_{rise} , G6 transition rates, k_0), a $\pm 50\%$ envelope is applied. For k_0 and λ , the `slim4` solver takes a single scalar per decomposition parameter, so their central values are applied uniformly during the run and the distributions are propagated analytically a posteriori (Section 3.4.2).

Table 4: Model parameters. PB: positive buoyancy. NB: negative buoyancy.

Parameter	Value	Source
Time step (Δt)	300 s	(Krumhansl et al., 2025)
Release height (h_c)	17.4 cm (3.4 cm to 54.4 cm)	(Rowlands et al., 2024)
Rising velocity (w_{rise})	$2 \times 10^{-2} \pm 1 \times 10^{-2} \text{ m s}^{-1}$	(Lai et al., 2020)
PB→NB rate, week 1	$0.02 \pm 0.01 \text{ d}^{-1}$	(Hanuise et al., 2025)
PB→NB rate, after	$0.035 \pm 0.017, 5 \text{ d}^{-1}$	(Hanuise et al., 2025)
Windage (α)	$\mathcal{U}(0.00881, 0.01313)$	(Hanuise et al., 2025; Lai et al., 2020)
Sinking velocity (w_s)	$7.5 \times 10^{-3} \pm 2.0 \times 10^{-3} \text{ m s}^{-1}$	(Weatherall et al., 2016)
Sinking range (sensitivity)	$6 \times 10^{-3} \text{ m s}^{-1}$ to $6 \times 10^{-2} \text{ m s}^{-1}$	(Oldham et al., 2014)
Decomp. rate (k_0)	$\mathcal{U}(0.006, 0.017 \text{ d}^{-1})$, central $0.011, 5 \text{ d}^{-1}$	(Ochieng and Erftemeijer, 1999; Duarte and Krause-Jensen, 2017)
Depth attenuation (λ)	$\mathcal{N}(4.6 \times 10^{-3}, 1.8 \times 10^{-3} \text{ m}^{-1})$, central $4.6 \times 10^{-3} \text{ m}^{-1}$	(Marsay et al., 2015; Cram et al., 2018)
Deposition threshold (U_c)	$0.045 \pm 0.011 \text{ m s}^{-1}$	(van der Mheen et al., 2024; Krumhansl et al., 2025)
Deposition duration	7 days	(Krumhansl et al., 2025)
Sequestration depth	$\geq 1,000 \text{ m}$ (convention)	(Duarte and Krause-Jensen, 2017)

3.4.3 From trajectories to carbon flux

The `LPT` simulation produces trajectories that carry a residual mass fraction $m_{\text{final},i}/m_0$, where m_0 is the unit mass at seeding and $m_{\text{final},i}$ the mass remaining when particle i reaches its final state (Eq. (24)). Translating this dimensionless mass fraction into an absolute carbon flux requires mapping each particle to a seagrass area and a daily biomass production. The parameters introduced below are summarised in Table 5. With N_e the daily count of particles seeded in mesh element e of horizontal area A_e (Section 3.4.2), each particle represents a seagrass area $a_e = A_e/N_e$ (m^2) and one day of leaf production over that area.

The daily carbon flux carried by particle i , F_i (g C d⁻¹), is then

$$F_i = E_{\text{ref}}(s_e) \frac{m_{\text{final},i}}{m_0} a_e, \quad (25)$$

where $E_{\text{ref}}(s_e)$ is the reference daily carbon export per unit seagrass area (g C m⁻² d⁻¹) for substrate $s_e \in \{\text{rock, sand}\}$, defined as

$$E_{\text{ref}}(s) = 0.178 f_C P(s). \quad (26)$$

Here $f_C = 0.292 \pm 0.058$ g C g⁻¹ DW is the carbon content of dry biomass measured at Seychelles sites (Rowlands et al., 2024), and $P(s)$ is the substrate-specific leaf production. The coefficient 0.178 is the POC fraction of seagrass NPP exported beyond the meadow (Figure 8), restricted to the particulate share that the Lagrangian particles can track (the remaining 6.7 % of NPP exported as DOC is excluded). The only direct *T. ciliatum* production measurement comes from Inhaca Island, Mozambique (Bandeira, 2002), yielding $P_{\text{rock}} = 25.5$ g DW m⁻² d⁻¹ (4 561 shoots m⁻²) on rocky reef and $P_{\text{sand}} = 9.5$ g DW m⁻² d⁻¹ (888 shoots m⁻²) on sand. These rates are applied under two simplifications. First, production is assumed constant in time, consistent with the weak seasonality of photoperiod and sea surface temperature at the near-equatorial Mahé Plateau (~ 4° S) and the low seasonal variability of tropical seagrasses (Lee et al., 2007). Second, production is assumed constant with depth. The Inhaca measurements were conducted in shallow water (exact depth not reported), whereas *T. ciliatum* on the Mahé Plateau spans 9 m to 25 m (Rowlands et al., 2024), with shoot density declining with depth as light availability drops (Serrano et al., 2019; Collier et al., 2007). Applying shallow-water rates uniformly therefore yields an upper-bound estimate of carbon export.

To select between P_{rock} and P_{sand} , the substrate type s_e is assigned from the Allen Coral Atlas benthic habitat map (Allen Coral Atlas, 2022; Lyons et al., 2020), with missing classifications in the deeper offshore stands defaulting to rock. This default reflects expert confirmation (Dr. Jérôme Harlay, University of Seychelles, pers. comm.) and the ecology of *T. ciliatum*, which in the western Indian Ocean grows almost exclusively on rocky or reef substrates (Gullström et al., 2002; Green and Short, 2003). Since P_{rock} is 2.7 times P_{sand} , this choice biases the carbon flux upward for elements where the true substrate is uncertain.

Finally, the annualised carbon export from element e to fate f , $C_{e \rightarrow f}$ (g C yr⁻¹), is obtained by summing F_i over all particles from e that end in fate f and rescaling to a full year,

$$C_{e \rightarrow f} = \frac{365}{N_{\text{seed}}} \sum_{i \in (e,f)} F_i, \quad (27)$$

where N_{seed} is the total number of release days over the simulation. Results are aggregated into two fate matrices of size $n_{\text{seagrass}} \times 5$, recording for each source element and each of the five fates the total particle count and the cumulative carbon mass, respectively. Sedimented carbon is counted as sequestered because burial isolates it from remineralisation on centennial timescales (Duarte and Krause-Jensen, 2017).

Table 5: Post-processing parameters for the trajectory-to-carbon-flux conversion (Section 3.4.3).

Parameter	Value	Source
Seeding density (ρ_p)	100 particles km^{-2} on the raw polygon coverage of $1,765 \text{ km}^2$	Section 3.4.2
Number of release days (N_{seed})	SLIM3D run duration, excluding the three-month spin-up and leaving a 40 d margin at the end for the released particles to complete their trajectories	—
Carbon content (f_C)	$0.292 \pm 0.058 \text{ g C g}^{-1} \text{ DW}$	(Rowlands et al., 2024)
POC exported fraction	17.8 % of NPP	(Duarte and Krause-Jensen, 2017)
Leaf production, rock (P_{rock})	$25.5 \text{ g DW m}^{-2} \text{ d}^{-1}$ (4 561 shoots m^{-2})	(Bandeira, 2002)
Leaf production, sand (P_{sand})	$9.5 \text{ g DW m}^{-2} \text{ d}^{-1}$ (888 shoots m^{-2})	(Bandeira, 2002)
Sequestration depth	$\geq 1,000 \text{ m}$ (convention)	(Duarte and Krause-Jensen, 2017)

3.5 Model validation

The hydrodynamic model is validated against four independent observational datasets covering the simulation period (13 January 2016 to 16 July 2020), after discarding the three-month spin-up consistent with Section 3.3, the period during which the model relaxes from its idealised initial conditions toward dynamical balance. The position, period and variables of each station are summarised in Table 6.

Sea-surface elevation is compared against the Pt La Rue tide gauge on Mahé, part of the University of Hawai'i Sea Level Center network (Caldwell et al., 2015) (Figure 23). The record extends from before 1993 and fully covers the simulation period.

Currents are validated against two open-ocean moorings and two plateau moorings (Figure 22). The two RAMA buoys (McPhaden et al., 2009), each anchored to the seabed and carrying a string of instruments through the water column (Figure 23), sample the open ocean. RAMA 4S57E (57°E , 4°S) returns daily currents at 10 m depth and RAMA 8S55E (55°E , 8°S) returns monthly currents. The two bottom-mounted ADCPs on the Mahé Plateau complement these records by listening upward through the water column and reconstructing velocity profiles from the Doppler shift of acoustic pulses scattered back by suspended particles (Teledyne RD Instruments, 2011). ADCP-N (55.60°E , 3.75°S) delivers hourly profiles from 15 January to 5 May 2017, and ADCP-E (55.65°E , 4.67°S) delivers 55 min profiles from 16 December 2014 to 12 March 2018.

Temperature and salinity are validated against three complementary datasets covering the full water column. The two RAMA moorings deliver daily T and S profiles at their respective locations (McPhaden et al., 2009). The GO-SHIP I07N 2018 cruise (GO-SHIP, 2018), sampled by the National

Oceanic and Atmospheric Administration (NOAA) research vessel *Ronald H. Brown* between 7 and 23 May 2018 (Figure 23), extends validation far below the surface through 43 Conductivity Temperature Depth (CTD) casts that reach 5,300 m along a south-to-north transect at $\sim 55^\circ\text{E}$. Each CTD is lowered from the ship as a rosette of sampling bottles fitted with sensors that log conductivity, temperature and pressure during the descent. 1,263 Array for Real-time Geostrophic Oceanography (Argo) profiles complement the GO-SHIP transect by covering the full simulation period and the entire study area. The Argo programme (Argo, 2024) maintains a global fleet of profiling floats that drift at a parking depth of $\sim 1,000$ m and surface every ten days to record a temperature and salinity profile from $\sim 2,000$ m to the surface.

For each station, SLIM3D fields are extracted at the nearest model node and interpolated onto the observation timestamps. At the RAMA moorings, modelled velocities are taken at the 10 m measurement depth. At the ADCP stations, both observed and modelled velocities are depth-averaged over the water column. At the tide gauges, modelled sea-surface elevation is extracted at the nearest coastal node. At each GO-SHIP CTD cast and Argo profile, modelled temperature and salinity profiles are interpolated onto the observed pressure levels (converted to depth using the Gibbs SeaWater toolbox `gsw_z_from_p`). Model performance is quantified by RMSE and mean bias for current speed, velocity components (u , v), temperature, and salinity, and by a directional RMSE for current direction.

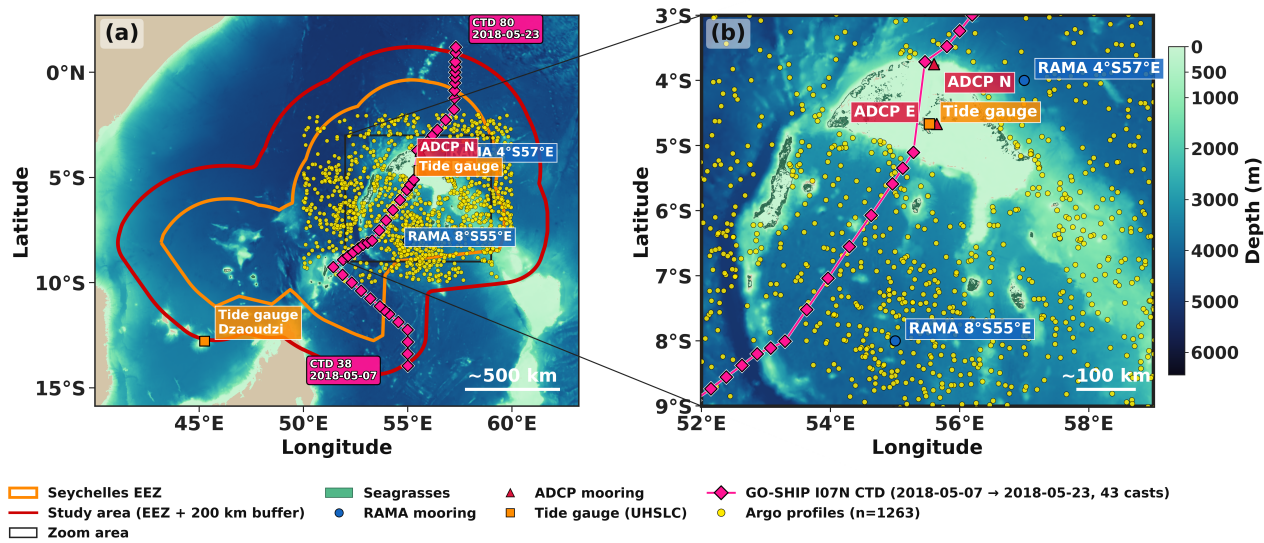


Figure 22: Observational datasets used for model validation. Two RAMA buoys (blue circles), two ADCP moorings (red triangles) on the Mahé Plateau, one University of Hawai'i Sea Level Center (UHSLC) tide gauge (orange square) at Point La Rue, the 43 CTD stations (magenta diamonds) of the GO-SHIP I07N 2018 transect, and 1,263 Argo profiles (yellow dots) across the study area. Seagrass meadows in green.

Bathymetry from General Bathymetric Chart of the Oceans (GEBCO).

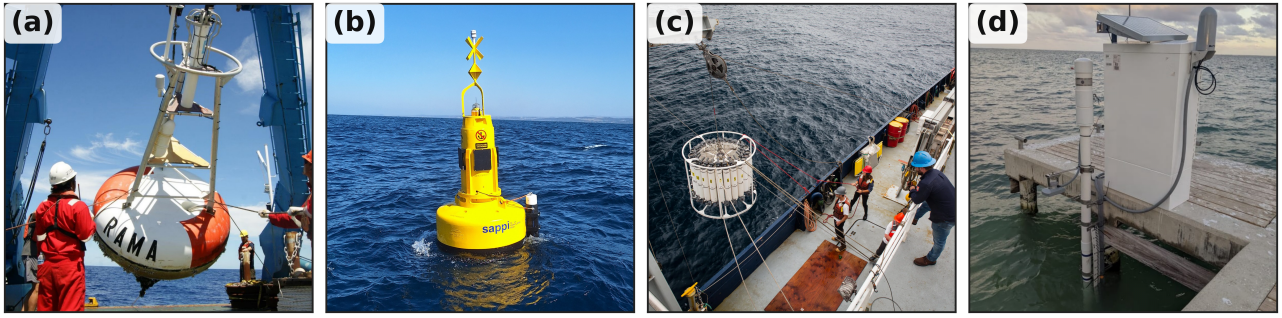


Figure 23: In-situ instruments used for model validation. (a) RAMA buoy NOAA PMEL (2019). (b) ADCP mooring Obscape (2023). (c) Shipboard CTD rosette deployment during a GO-SHIP cruise Rocha (2019) (d) Coastal tide gauge station EuroSea Consortium (2023).

Table 6: Validation datasets: position, sampling period within the simulation window (13 January 2016 to 16 July 2020), and observed variables.

Station	Position	Coverage	Variables
RAMA 4S57E	4°S, 57°E	full simulation period	u , v at 10 m, T , S (daily)
RAMA 8S55E	8°S, 55°E	full simulation period	u , v at 10 m (monthly), T , S (daily)
ADCP-N	3.75°S, 55.60°E	15 Jan – 5 May 2017 (110 days)	u , v profiles (hourly)
ADCP-E	4.67°S, 55.65°E	1 Jan 2016 – 12 Mar 2018 (802 days)	u , v profiles (55-min)
Tide gauge Pt La Rue	Mahé (4.67°S, 55.53°E)	full simulation period	sea-surface η (hourly)
GO-SHIP I07N CTD	43 stations ~55°E	7 – 23 May 2018 (16 days)	T , S profiles to 5,300 m
Argo floats	1,263 profiles across study area	full simulation period	T , S profiles to ~2,000 m

The Lagrangian particle tracker cannot be validated quantitatively, as the only available field evidence consists of opportunistic photographs of seagrass propagules on the seabed, with no constraints on release date, source location or transit pathway. Validation is therefore restricted to the hydrodynamic model output, which is treated as a necessary condition for the realism of the simulated dispersal.

3.6 Uncertainty quantification

Tracing carbon from meadow to sequestration relies on the modelling steps just presented, each adding its own uncertainty. The aim here is an approximate uncertainty budget that strikes the best compromise between coverage and implementation cost, rather than an exhaustive propagation. The SLIM3D solver that simulates the currents is deterministic, but the simulated field still depends

on empirical sub-grid parameters whose own uncertainty would be too costly to propagate explicitly. These contributions are left outside the budget for simplicity. The model performance against in-situ records is reported in Section 4.1 as a qualitative check, since the sparse observational network does not support a quantified domain-wide error. The Lagrangian step introduces a stochastic component through the diffusion parameterisation (Section 3.4.1), assumed to be absorbed by the 253 million particles aggregated in Section 4.3. What remains is the conversion of particle trajectories into carbon mass, where the four scaling factors are varied stochastically within plausible ranges drawn from the literature and are assumed to dominate the budget. They are propagated through a Monte Carlo simulation following Approach 2 of the IPCC 2006 uncertainty guidance (Frey et al., 2006) (Figure 24). The four parameters fall into two distribution families (Figure 24a-d). A Gaussian is used when a central value comes with a small symmetric spread: carbon content f_C (inter-site spread of the Seychelles inventory (Rowlands et al., 2024)), the POC export ratio r_{exp} (inter-species spread around the 17.8 % global mean reported by Duarte and Krause-Jensen (2017)), and meadow area A (30 % envelope reported in the source map of (Rowlands et al., 2024)). A triangle is used when only lower and upper bounds and a preferred value are available: leaf production P brackets its substrate dependence between the sand and rock values measured by Bandeira (2002), with the mode at the rock value. Each iteration draws one value independently from each distribution (Figure 24, step 1) and feeds Eq. (25) with the deterministic LPT output (step 2) to give the creditable flux F_k , stored at step 3. The draw is repeated (step 4) until the mean and the empirical distribution stabilise, reached at $N = 200,000$ here. The result is then summarised by the median and 95 % confidence interval of $\{F_1, \dots, F_N\}$ (step 5).

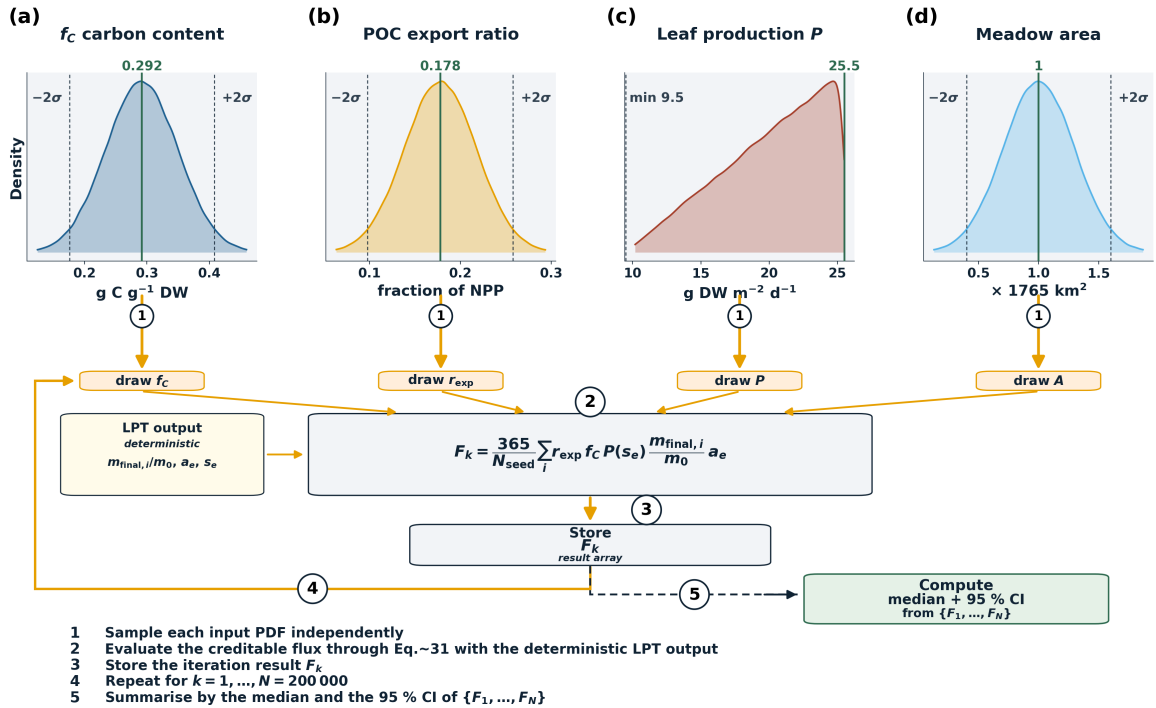


Figure 24: Monte Carlo propagation of the four input uncertainties through Eq. (25). Panels (a)–(d) show the input distributions $f_C \sim \mathcal{N}(0.292, 0.058)$, $r_{\text{exp}} \sim \mathcal{N}(0.178, 0.040)$, $P \sim \text{Tri}(9.5, 25.5, 25.5)\ g\ m^{-2}\ d^{-1}$ and $A \sim \mathcal{N}(1.0, 0.30) \times 1,765\ km^2$. Green lines mark the central value, grey dashed lines the $\pm 2\sigma$ bounds (or min/max for the triangular). Numbered arrows trace one Monte Carlo iteration: prior draw (1), evaluation of F_k with the deterministic LPT output (2), storage (3), loop (4), and median + 95 % CI of $\{F_1, \dots, F_N\}$ (5).

4 Results

The results first validate the simulated hydrodynamics against in-situ observations (Section 4.1). The validated velocity field then drives the particles, from the seasonal circulation over the Mahé Plateau (Section 4.2) to their domain-wide fates and carbon budget (Section 4.3). The robustness of these estimates is finally tested through a sensitivity analysis on the lifecycle parameters and a Monte Carlo propagation of input uncertainty on the carbon-mass scaling (Section 4.3.4).

4.1 Model validation

The validation starts with the simplest signal an ocean model is asked to reproduce, the free-surface elevation η . The only publicly available record on the Mahé Plateau is the **UHSLC** tide gauge at Pointe La Rue, located in the centre of the plateau (Figure 25b). Over the full overlap between the tide gauge record and the SLIM simulation (Figure 25a), the simulation matches the observations well, with a mean bias of -3.5 cm, an **Root-Mean-Square Error (RMSE)** of 8.9 cm (less than 9 % of the mean tidal range of about 1 m) and a correlation of $R = +0.97$. A three-month close-up over February to April 2017 (Figure 25c) resolves the spring–neap modulation of the tidal range, with a period of about 15 days, and the mean diurnal cycle (Figure 25g) is semi-diurnal, with two high and two low

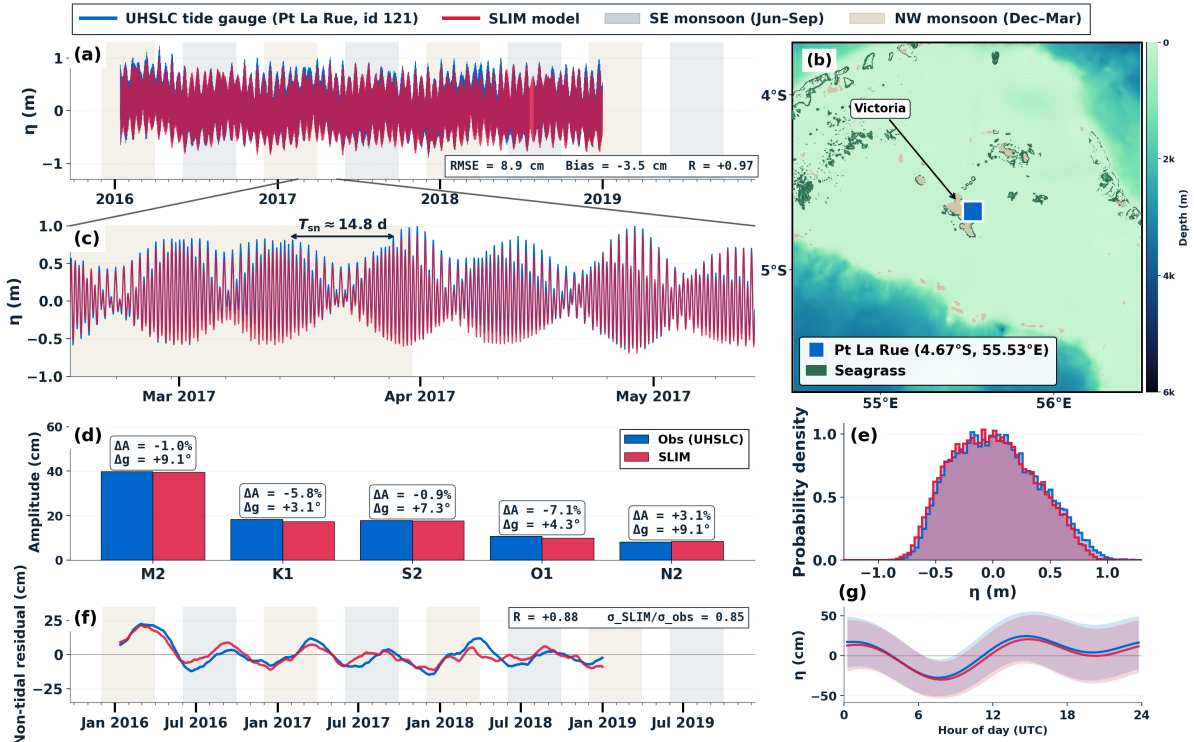


Figure 25: Validation of the simulated free-surface elevation η against the **UHSLC** tide gauge at Pointe La Rue (Mahé), 2016–2018. (a) Hourly observed (blue) vs modelled (red) η , monsoon shading. (b) Locator map: Pointe La Rue (orange square), GEBCO bathymetry, seagrass meadows (darkgreen). (c) February–April 2017 close-up, 15 days tidal modulation. (d) Five dominant tidal constituents from **Unified Tidal Analysis Toolbox (UTide) Codiga (2011)**, annotated with amplitude error ΔA and phase lag Δg . (e) PDFs of observed and modelled η . (f) Non-tidal residual: 30 days rolling mean of η after harmonic removal. (g) Mean diurnal cycle, semi-diurnal regime swinging ± 50 cm.

tides every 24 hours and the second peak slightly weaker than the first. The harmonic structure is reproduced with comparable fidelity. Both series are decomposed into their dominant frequencies with the **UTide** toolbox (Codiga, 2011), retaining the five leading tidal constituents (Figure 25d), with a mean amplitude error $|\Delta A|/A_{\text{obs}}$ of 3.6 % and a mean phase lag $|\Delta g|$ of 6.6° . Beyond the harmonic prediction, the non-tidal residual (Figure 25f, the 30-day rolling mean of η after removing the harmonic prediction) traces a clear seasonal swing of about 25 cm peak-to-peak, higher in the NW monsoon and lower in the SE, which the simulation tracks at $R = +0.88$.

Particle advection relies on the simulated three-dimensional velocity field, which is validated against two bottom-mounted **ADCPs** on the Mahé Plateau (Figure 26j). ADCP-E sits on the inner plateau (27 m depth, 802 days of hourly record), and ADCP-N near the northern shelf break (29 m depth, 110 days of record). The two sites therefore sample contrasting current regimes. Instantaneous vectors agree weakly at both moorings (Figure 26a–d), with daily **RMSEs** on u and v ranging from 11.8 cm s^{-1} to 21.4 cm s^{-1} (Figure 26e–h). Correlations on the four velocity components stay low: only the zonal component at ADCP-E reaches $R = +0.20$, the three others remain at $|R| \leq 0.07$. Beyond this common weak point-by-point agreement, the two sites diverge. Vector alignment, quantified by $\arg(\rho_c)$ (Kundu, 1976) (Appendix E), is $+22^\circ$ at ADCP-E (Figure 26k) but reaches -57° at ADCP-N (Figure 26l), a much larger misalignment on flow direction. The mean depth-averaged speed shows the same pattern: overestimated by 6 % at ADCP-E (Figure 26o) and by 24 % at ADCP-N (Figure 26p). The vertical structure shows the same contrast more sharply: at ADCP-E, the simulated profile matches the surface-intensified observed structure ($\text{RMSE} = 3.9 \text{ cm s}^{-1}$, $R = +0.99$, Figure 26m), while at ADCP-N it inverts the observed depth gradient, increasing toward the surface instead of toward the bed ($\text{RMSE} = 20.1 \text{ cm s}^{-1}$, $R = -0.55$, Figure 26n). Consistent with this inversion, the mean simulated speed at ADCP-N reaches 33.7 cm s^{-1} , against 14.1 cm s^{-1} observed. Finally, the simulated **Turbulent Kinetic Energy (TKE)** exceeds observations 1.4-fold at ADCP-E and 5-fold at ADCP-N (Figure 26i). Overall, the model captures the inner-plateau regime well but overshoots the shelf-break dynamics.

The **ADCP** coverage stops at the plateau edge, but particles released from the meadows are expected to spend most of their post-detachment drift further offshore, so the validation is extended into the open ocean. Two **RAMA** buoys at $4^\circ\text{S } 57^\circ\text{E}$ and $8^\circ\text{S } 55^\circ\text{E}$ provide daily currents at 10 m depth (Figure 27f). Before turning to the comparison, the observations themselves require a check: at $4^\circ\text{S } 57^\circ\text{E}$, observed u and v collapse from typical daily amplitudes of 30 cm s^{-1} to 50 cm s^{-1} over 2016–2018 to near-zero in 2019 (Figure 27a,b, blue). The January–March 2019 mean speed reaches 5.8 cm s^{-1} , well below the 30 cm s^{-1} to 50 cm s^{-1} recorded at the same buoy before 2019 (Figure 27e). This abrupt change is presumed to reflect an instrument issue at the buoy, and post-2018 data are therefore excluded. Over the retained 2016–2018 window, the simulation reproduces the daily u, v variability with $R = +0.63$ and $R = +0.62$ (**RMSEs** of 28.0 and 23.8 cm s^{-1} , Figure 27a,b). The direction rose aligns with observations ($\arg(\rho_c) = -5^\circ$, Figure 27g), and the simulated speed distribution overlaps the observed one with a slight excess at high speeds and a 5.1 cm s^{-1} bias (Figure 27i). At $8^\circ\text{S } 55^\circ\text{E}$, the paired record reduces to roughly 53 days, and the component correlations are $R = +0.32$ and $R = +0.09$. The rose (Figure 27h) shows a rotation of $\arg(\rho_c) = -59^\circ$, and the speed distribution (Figure 27j) peaks near the observed values with a 2.5 cm s^{-1} bias.

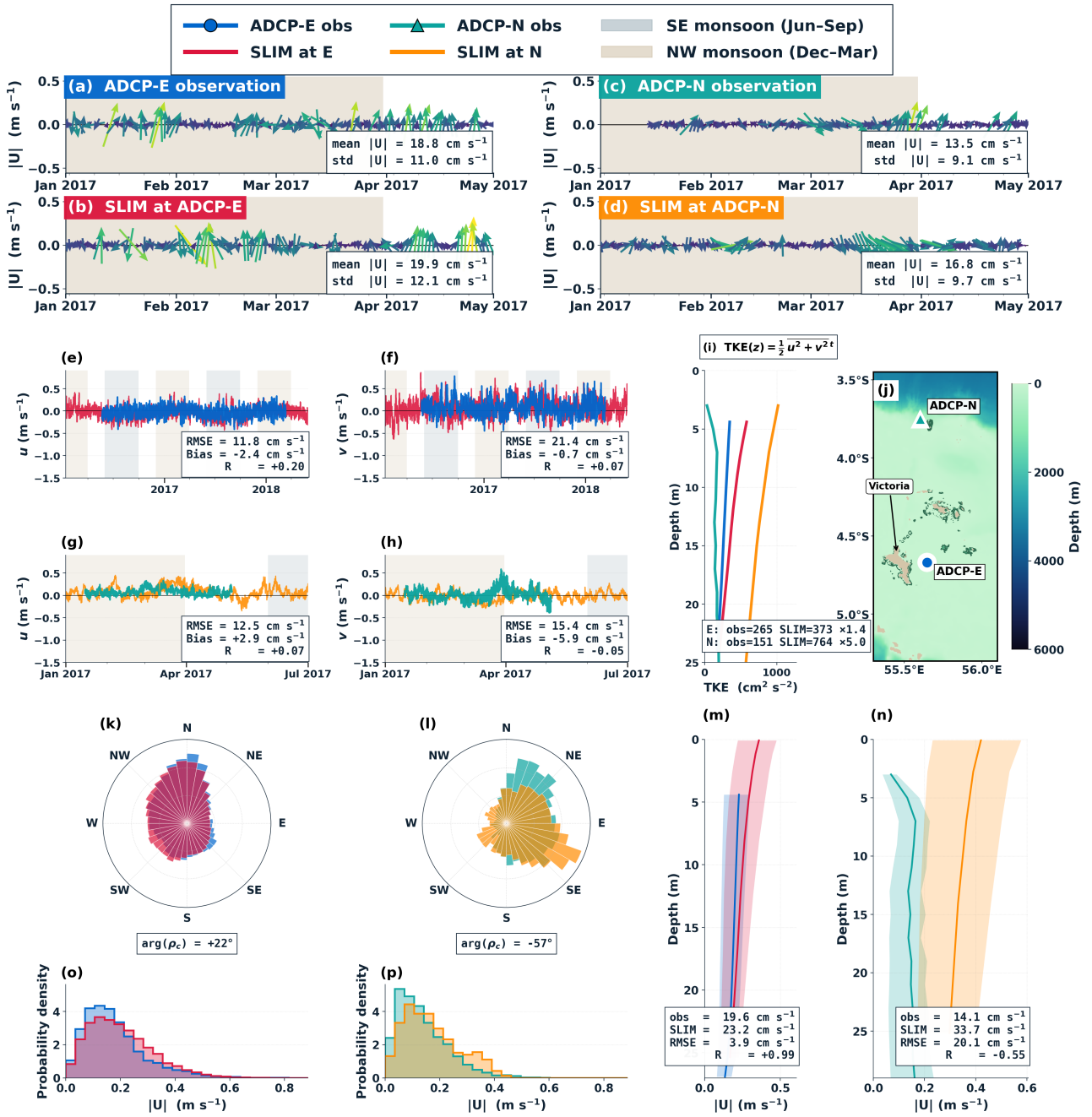


Figure 26: Validation of simulated currents against the two Mahé Plateau ADCPs (2016–2018). (a–d) Feather plots of hourly depth-averaged horizontal velocity over January–May 2017 with monsoon shading: (a) ADCP-E observed, (b) **SLIM3D** at ADCP-E, (c) ADCP-N observed, (d) **SLIM3D** at ADCP-N. (e–h) Time series of the depth-averaged velocity components: (e, f) zonal u and meridional v at ADCP-E over the full record, (g, h) zonal u and meridional v at ADCP-N over its observation window. Inset RMSE, bias and R are computed on daily means of the corresponding component. (i) Time-mean vertical profiles of kinetic energy $\text{TKE}(z) = \frac{1}{2} \overline{u^2 + v^2}$, at both sites over the depth range common to observations and **SLIM3D**. (j) Map showing **GEBCO** bathymetry, seagrass meadows, and ADCP locations. (k, l) Depth-averaged current-direction roses at ADCP-E and ADCP-N on a common frequency scale. (m, n) Time-mean vertical profiles of $|U|$ with p25–p75 interquartile envelopes. These panels use $\overline{|U|}$ (time-mean of the instantaneous speed) instead of $|\bar{U}|$ (speed of the mean vector) used elsewhere, which inflates the apparent **SLIM3D**–observation gap because vector averaging cancels opposing currents. (o, p) Probability density functions of $|\bar{U}|$. Detailed analysis of each panel is provided in the main text.

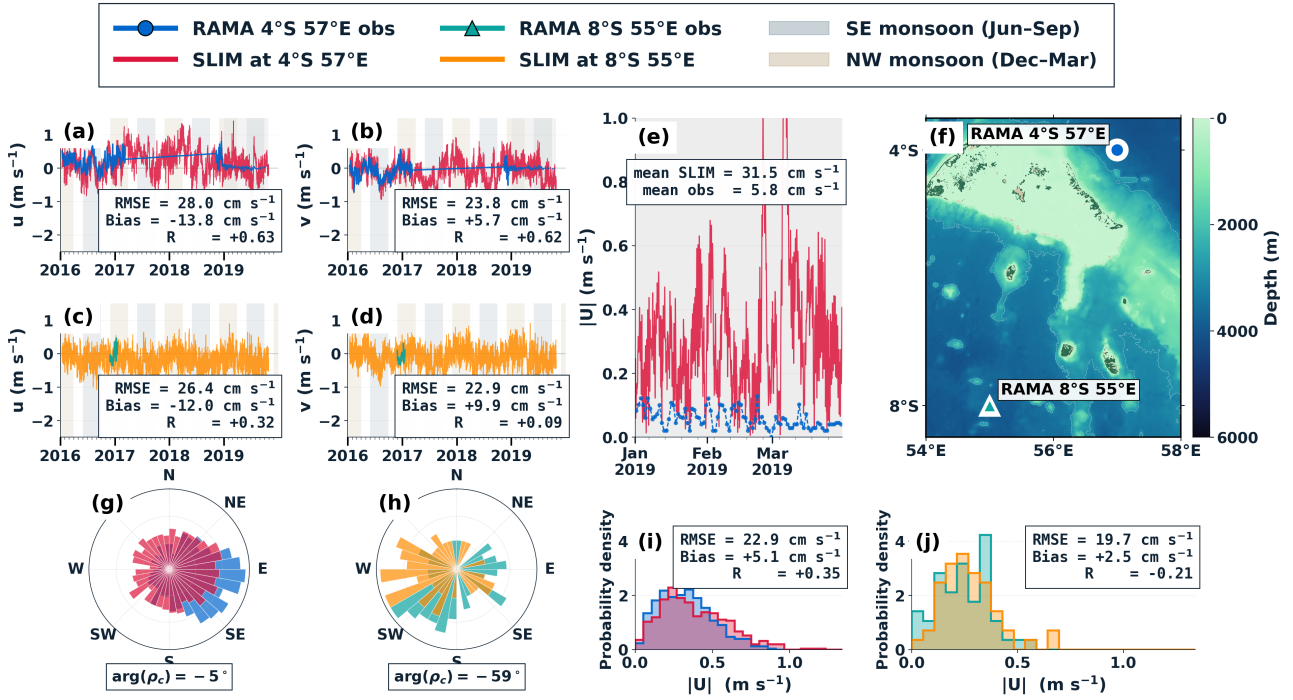


Figure 27: Validation of simulated currents against the two RAMA buoys nearest to the Mahé Plateau (4°S 57°E and 8°S 55°E), at the nominal observation depth of 10 m. The 4°S 57°E currentmeter failed after 2018, so the comparison is restricted to 2016–2018 at that station and to the same window at 8°S 55°E for consistency. (a–d) Daily time series of the horizontal velocity components with monsoon shading: (a) zonal u at 4°S 57°E (SLIM3D red, obs blue), (b) meridional v at 4°S 57°E, (c) zonal u at 8°S 55°E (SLIM3D orange, obs teal), (d) meridional v at 8°S 55°E. Inset RMSE, bias and R are computed on the daily values of the corresponding component. (e) Two-month close-up view of $|U|$ at 4°S 57°E over January–March 2019 (sensor-failure period), with mean SLIM3D and mean obs over the window. (f) Map showing GEBCO bathymetry and the two RAMA buoy locations. (g, h) Depth-averaged current-direction roses at 4°S 57°E and 8°S 55°E on a common frequency scale, with the complex-correlation rotation angle $\arg(\rho_c)$ annotated (Appendix E). (i, j) Probability density functions of $|U|$ at 4°S 57°E and 8°S 55°E, with inset RMSE, bias and R on the SLIM3D versus obs speeds.

Beyond currents, T and S are also compared to in-situ observations, even though their role in particle advection is more indirect (through density and stratification). The thermohaline structure is evaluated against three independent datasets across the domain (Figure 28a): 1,263 Argo profiles, 41 Global Ocean Ship-based Hydrographic Investigations Program (GO-SHIP) I07N CTD casts along a meridional transect, and two RAMA moorings (4°S 57°E and 8°S 55°E). SLIM3D is sampled at the same location and time as each observation. The three observed $T(z)$ profiles are nearly indistinguishable across networks, and the model matches them on average (Figure 28b). SLIM3D fits the observations closely down to 250 m, overestimates T between 250 m to 1,000 m and underestimates it slightly below. RMSE(T) reaches up to 4 °C, with a broad maximum around 500 m to 1,000 m and a secondary peak near 1,200 m, with a depth- and profile-averaged value of 2.0 °C (Figure 28c). The observed $S(z)$ profiles are likewise indistinguishable across networks, and SLIM3D follows them closely (Figure 28d). The upper 100 m are well reproduced, SLIM3D underestimates S between 100 m to 200 m and overestimates it below. RMSE(S) peaks near 10 m and 500 m before decreasing at depth, remaining below 0.4 psu with a mean of 0.12 psu (Figure 28e).

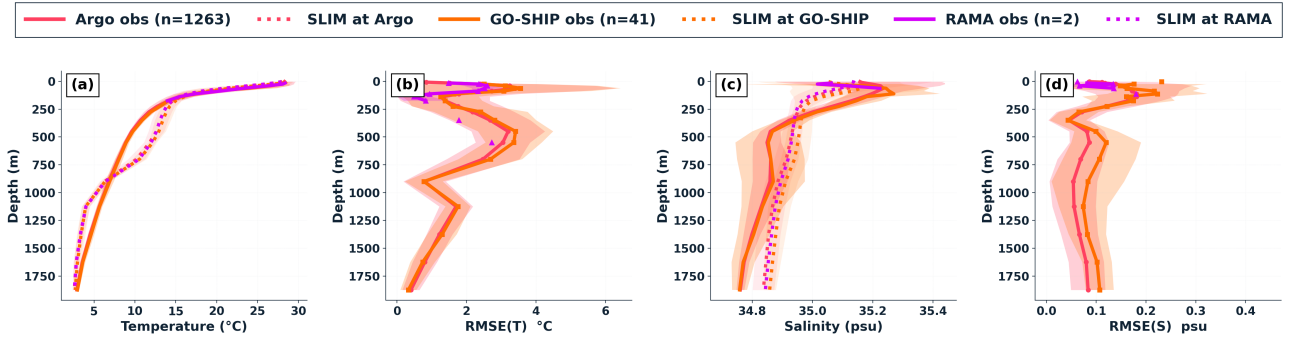


Figure 28: Validation of **SLIM3D** temperature and salinity against in situ observations over the simulation period. (a) Location of Argo floats ($n=1263$), GO-SHIP CTD casts ($n=41$) and RAMA moorings ($n=2$) within the Seychelles **EEZ** (orange) and the study area (red). (b,d) Mean vertical profiles of temperature and salinity from observations (solid) and **SLIM3D** collocated in space and time (dotted), with shaded bands indicating $\pm 1\sigma$. (c,e) Depth-resolved **RMSE** between **SLIM3D** and each observational dataset.

4.2 Local-scale dynamics on the Mahé Plateau

Circulation over the Seychelles **EEZ** is shaped by the seasonal reversal between the NW and SE monsoons (Section 2.1), and the seasonal-mean simulated surface currents (vertical average over 0–10 m) and the 10 m ERA5 forcing wind (Hersbach et al., 2020) resolve this contrast (Figure 29). During the NW monsoon (Figure 29a,b), the simulated flow forms a broad clockwise gyre around the Mahé Plateau, with a dominant west-to-east transport over the plateau itself and streamlines partly confined to the shelf. The associated wind blows in a dipole pattern across the **EEZ**, southwestward in the northern half and westward in the southern half, which combines into a steady northwest-to-southeast flow over the plateau (Figure 29c,d). The SE monsoon (Figure 29e,f) reverses this organisation: the simulated flow turns predominantly westward across the full **EEZ** with nearly parallel streamlines, the clockwise gyre is no longer visible in the seasonal-mean field over the plateau, and the mean arrows point uniformly westward with no return flow toward Mahé island. The wind blows steadily from the southeast toward the northwest at both scales (Figure 29g,h), replacing the north-south dipole, and SE winds and currents are markedly stronger than their NW counterparts (mean wind speed 7.32 m s^{-1} against 2.55 m s^{-1} , mean current speed 41.2 cm s^{-1} against 17.1 cm s^{-1}). The two inter-monsoon transitions fall between these extremes, on average aligned with the SE direction and with intermediate wind intensities of 4.14 m s^{-1} (Inter-monsoon 1, April–May) and 4.77 m s^{-1} (Inter-monsoon 2, October–November).

These contrasting wind and current regimes shape how long particles released from Mahé Plateau meadows (the largest shelf in the **EEZ**, hosting 720.9 km^2 of seagrass, 40.7 % of the **EEZ** total) remain on the plateau before being advected away (Figure 30). The plateau exit time differs strongly across regimes (Figure 30a): the SE monsoon drives the fastest evacuation, with a median exit time of 11 days and 71.5 % of particles having left by 50 days, whereas the NW monsoon is the slowest, with a median of 94 days and only 44.7 % by 50 days. The two inter-monsoon transitions fall between these extremes, with median exit times of 22 days (IM1) and 83 days (IM2). The coastal stranding rate (Figure 30b) contrasts most sharply between the NW (12.6 % at 180 days) and the SE (7.1 %). The net-return rate (Figure 30c), defined as the fraction of re-entered particles that subsequently leave the plateau again (second departures over re-entries), reaches 11.7 % in SE against 2.6 % in NW. Both

values remain low, indicating that particles that leave the plateau rarely come back, regardless of the regime, although the strong SE currents recirculate a slightly larger fraction onto the shelf before leaving again.

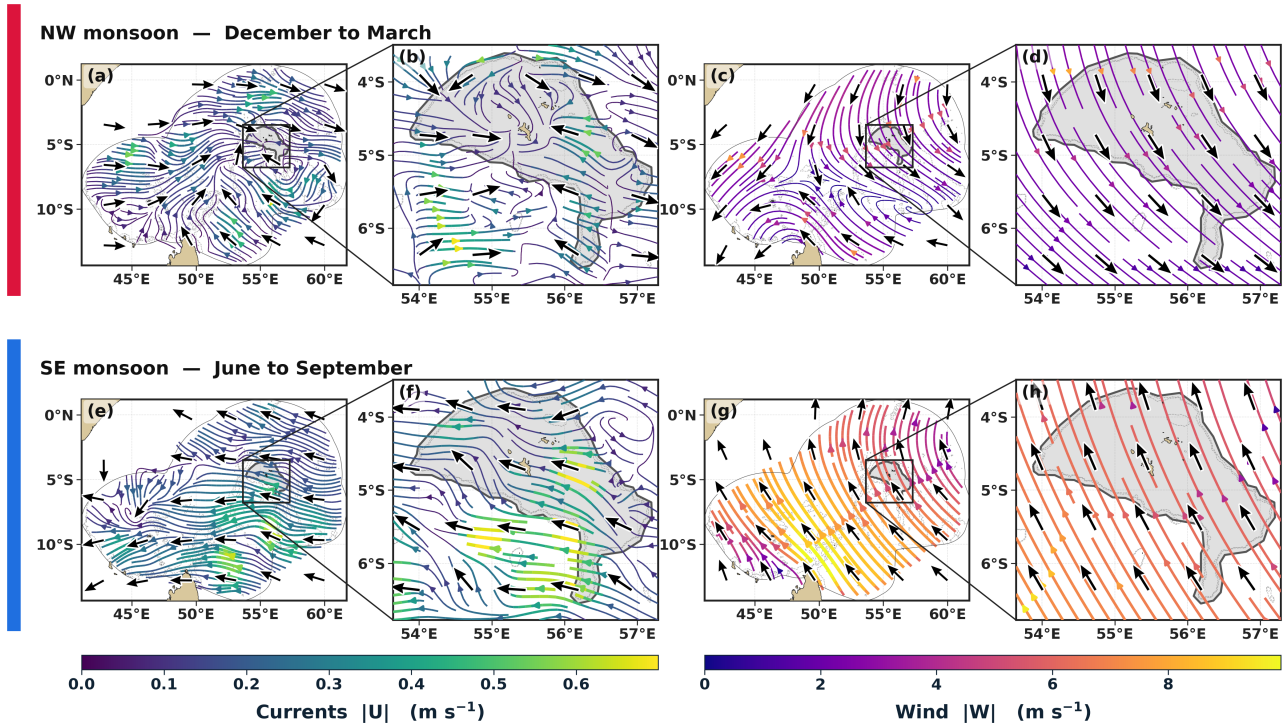


Figure 29: Seasonal-mean near-surface currents (0–10 m depth) and 10 m wind from [ECMWF Reanalysis v5 \(ERA5\) Hersbach et al. \(2020\)](#), averaged over 534 daily snapshots (February 2016–July 2020). Left panels show the [EEZ-scale](#) view, right panels the Mahé Plateau zoom. (a,b) NW monsoon currents. (c,d) NW monsoon wind. (e,f) SE monsoon currents. (g,h) SE monsoon wind. Streamlines are coloured by speed, with black arrows showing the local vector-mean flow direction, and the grey contour marks the –200 m isobath.

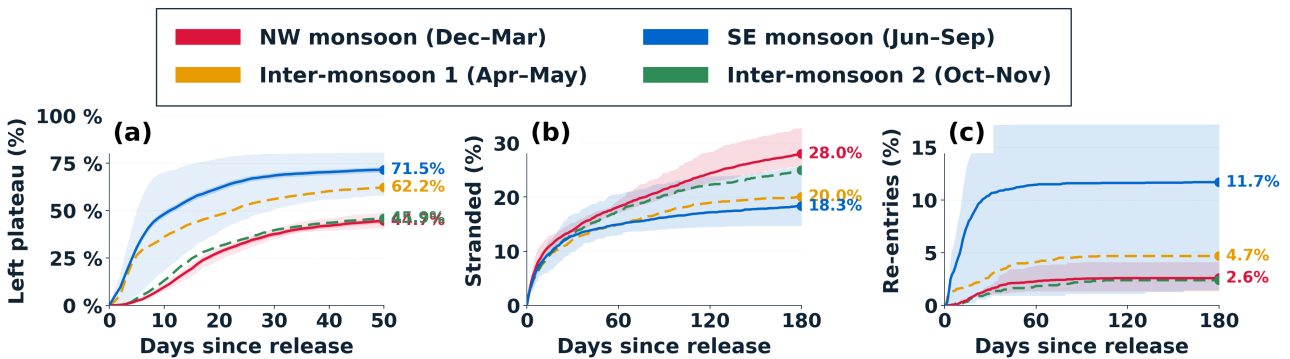


Figure 30: Seasonal Lagrangian dynamics of particles released from Mahé Plateau seagrass meadows. (a) Exit-time curves of released particles by season (bootstrap 95 % CI). (b) Cumulative coastal stranding fraction. (c) Net-return fraction. Season codes: NW Dec–Mar, IM1 Apr–May, SE Jun–Sep, IM2 Oct–Nov.

4.3 Domain-wide outcomes

This section presents the central result of the thesis. The **LPT** model runs across the full study area on the simulated currents validated in Section 4.1, from July 2016 to January 2020, with no particles released over the last four months of the simulation to ensure that all included particles have reached their terminal fate. Two analyses are then carried out on the same set of simulated particles, from their terminal fates to their carbon contribution across the reservoirs defined in Section 3.4.2. The first follows each particle along its trajectory and counts it as a single unit at its terminal fate, giving the distribution of destinations across reservoirs. The second uses the same trajectories but tracks how the carbon mass each particle carries varies along the way: each particle is weighted by an exported **POC** mass set in proportion to the meadow area it represents and to its sandy or rocky substrate, and decomposition erodes that mass along the trajectory as a function of transit time and depth. What remains in each reservoir gives the resulting carbon budget.

4.3.1 Particle fates

The Lagrangian experiment releases 173,000 particles per day, at a density of 100 particles per km² of meadows, from the *T. ciliatum* meadows of the Mahé Plateau, 72 million per year and 253 million in total, each tracked until it reaches a terminal fate. These end states are split into five categories (Figure 31a), and two of them dominate. The first, deep ocean sequestration, comprises particles sinking past 1,000 m depth, reaching 43.0 ± 4.4 % inside the Seychelles **EEZ** (i) and a further 10.0 ± 1.0 % in the surrounding 200 km buffer (ii), 53 % in total. The second, lateral export (iii), gathers particles that cross the open boundary of the domain, beyond which currents are no longer resolved and exported particles are assumed to sink. They account for 35.0 ± 6.1 %, so that the two pathways together carry 88 % of the partition. The remainder splits between coastal stranding (iv), at 9.0 ± 0.7 %, where particles reach the closed coastal boundary and are assumed to strand there, and seafloor sedimentation (v), at 2.0 ± 0.1 %, where particles settle on the shelf seabed under near-bottom currents too weak to resuspend them.

Regrouping the five fates along the blue carbon accounting boundary yields three policy-relevant outcomes (Figure 31b). All shares below count particles reaching each outcome, not the carbon mass they carry. Blue carbon creditable to the Seychelles, gathering particles finishing in deep ocean sequestration inside the **EEZ** (i) and the in-**EEZ** share of seafloor sedimentation (v), reaches 46.0 ± 4.5 % of released particles. Blue carbon sequestered beyond national borders gathers particles finishing in deep ocean sequestration in the buffer (ii), the buffer share of seafloor sedimentation (v), and lateral export (iii), assumed to sink past the open boundary within roughly 1.5 days of crossing, totalling 45.0 ± 5.1 % of released particles. The non-blue residual reduces to coastal stranding (iv) at 9.0 ± 0.7 % of released particles.

Figure 31c stacks the particles released on each day, coloured by the fate they eventually reach weeks to months later². From one date to the next, deep ocean sequestration in the **EEZ** (i) and

²For instance, the column for 1 January 2018 shows about 41 % of that day's release ending in lateral export (iii), against 37 % for deep ocean sequestration inside the **EEZ** (i).

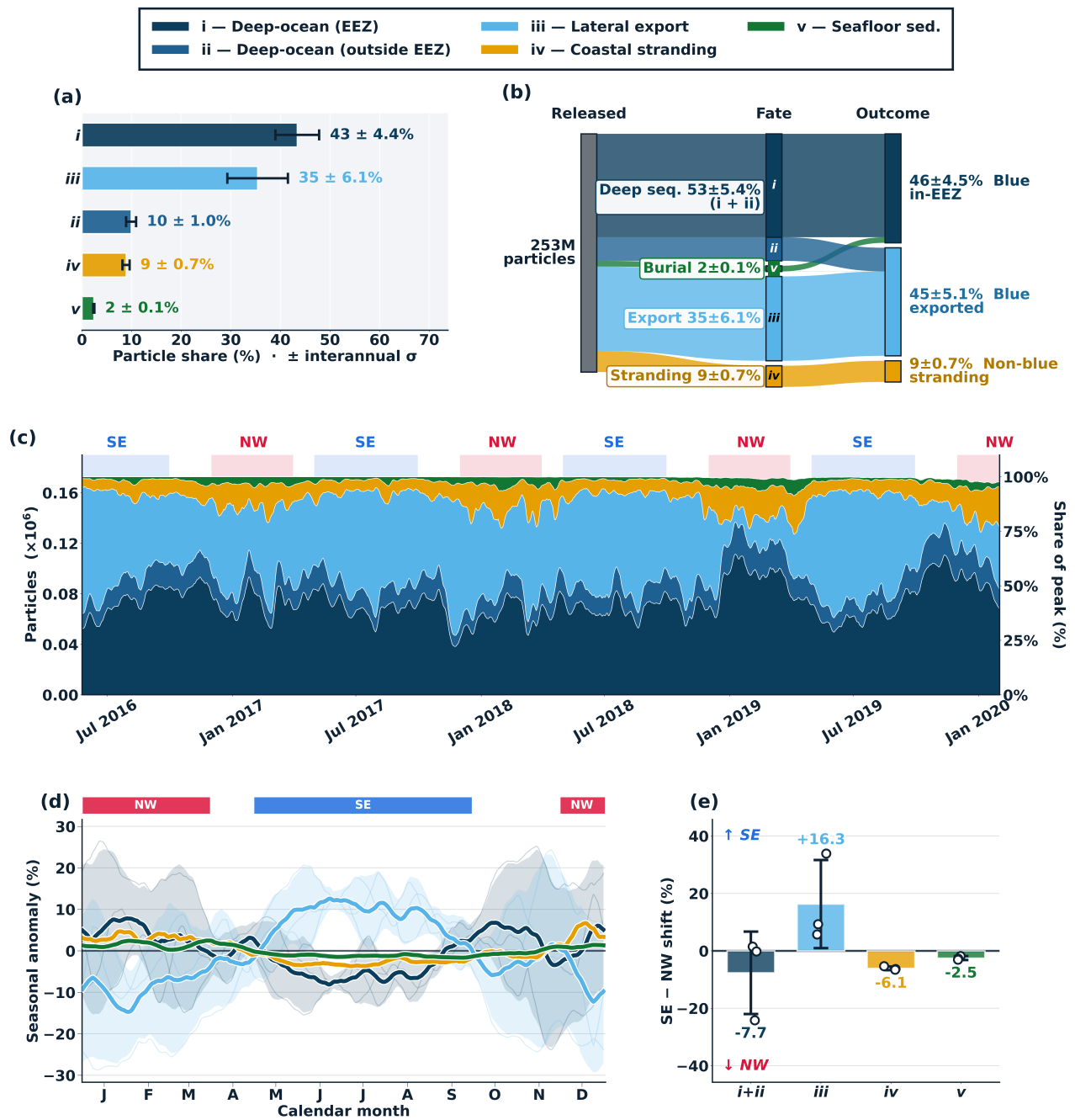


Figure 31: Domain-wide particle fates pooled over all Seychelles meadows, July 2016–January 2020 (particle counts). (a) Mean fate shares in per cent $\pm$ interannual σ for fates i–v. (b) Sankey diagram of the 253 million released particles, from release through fate to outcome. (c) Particles stratified by release day on the x-axis, with each column stacking the cumulative count of particles released that day, coloured by the terminal fate they eventually reach weeks to months later. Left axis: particle count $\times 10^6$, right axis: share in per cent, SE and NW monsoon windows shaded. (d) Monthly departure of each fate's share from its annual mean, averaged across years in bold, with individual years lighter for the two most variable fates (deep ocean and lateral export). (e) SE minus NW shift per fate, in per cent.

lateral export (*iii*) are anticorrelated, so their combined share stays nearly constant, while coastal stranding (*iv*) and seafloor sedimentation (*v*) hold a steady thin share.

Figure 31d gives the monthly climatology, with each fate's share plotted as a departure from its annual mean in bold and individual years drawn lighter around (*i*) and (*iii*). Lateral export (*iii*) peaks during SE and drops during NW, in mirror to deep ocean sequestration in the EEZ (*i*), confirming the anticorrelation seen at the daily scale. The envelope on these two curves is wide, often comparable to the seasonal swing itself, while coastal stranding (*iv*) and seafloor sedimentation (*v*) stay close to zero throughout the year, dipping slightly in SE. Figure 31e condenses this signal into a single seasonal shift per fate. Lateral export (*iii*) carries the largest swing at 16.3 % in SE, mirrored by combined deep ocean sequestration (*i+ii*) at -7.7 %, while coastal stranding (*iv*) and seafloor sedimentation (*v*) shift only -6.1 % and -2.5 %. The error bars on (*iii*) and (*i+ii*) span 30 % to 40 %, larger than the seasonal swing itself, so individual years can flip the mean signal.

Figure 32 maps where particles end and how far they travel. At basin scale (Figure 32a), lateral export (*iii*) concentrates along the open boundary north of 13° S, combined deep ocean sequestration (*i+ii*) spreads across the basin with higher density around the Seychelles release sites, and coastal stranding (*iv*) marks the Seychelles coasts and part of Madagascar. The plateau zoom (Figure 32b) resolves seafloor sedimentation (*v*) as dense patches at the shelf break. On Mahé, stranding lines the coast (Figure 32c) and extends northward along the eastern shore (Figure 32d). Endpoint density follows the monsoon regime: under NW conditions (Figure 32e) particles drift east and south of the plateau, while under SE conditions (Figure 32f) they shift north and west. Monthly mean distances (Figure 32g) show combined deep ocean sequestration (*i+ii*), coastal stranding (*iv*) and seafloor sedimentation (*v*) holding near-constant distances throughout the year, while lateral export (*iii*) carries nearly all the seasonal variation, spanning 400 km to 600 km with peaks at the monsoon shoulders (April–May and September–October). Averaged across the simulation, lateral export (*iii*) travels 454 km from release, against 185 km for combined deep ocean sequestration (*i+ii*), 77 km for coastal stranding (*iv*) and 63 km for seafloor sedimentation (*v*) (Figure 32h). Because the daily seeding rate is constant, the aggregated daily transport (Figure 32i) directly visualises the monsoon imprint: lateral export (*iii*) peaks during the SE monsoon while combined deep ocean sequestration (*i+ii*) peaks during the NW monsoon. The shift between the SE and NW monsoons (Figure 32j) condenses this contrast: combined deep ocean sequestration (*i+ii*) is larger in NW, lateral export (*iii*) in SE, while coastal stranding (*iv*) and seafloor sedimentation (*v*) shift only modestly.

4.3.2 Carbon mass budget

Trajectories are now weighted by the carbon mass each particle carries, shifting from a count view to a mass view. Only the remaining fraction reaches a sink and feeds the blue carbon budget, so this view is the one that answers the research question on seagrass carbon export. In this simulation, Seychelles meadows release 737 kt C in total (Figure 33), 200 kt C per year and 0.55 kt C per day. That total combines the meadow surface, a substrate-specific (sandy or rocky) net primary production rate, and the 17.8 % POC fraction exported from the meadow. Each particle starts with a mass proportional to the meadow area it represents, and decomposition erodes that mass along the trajectory.

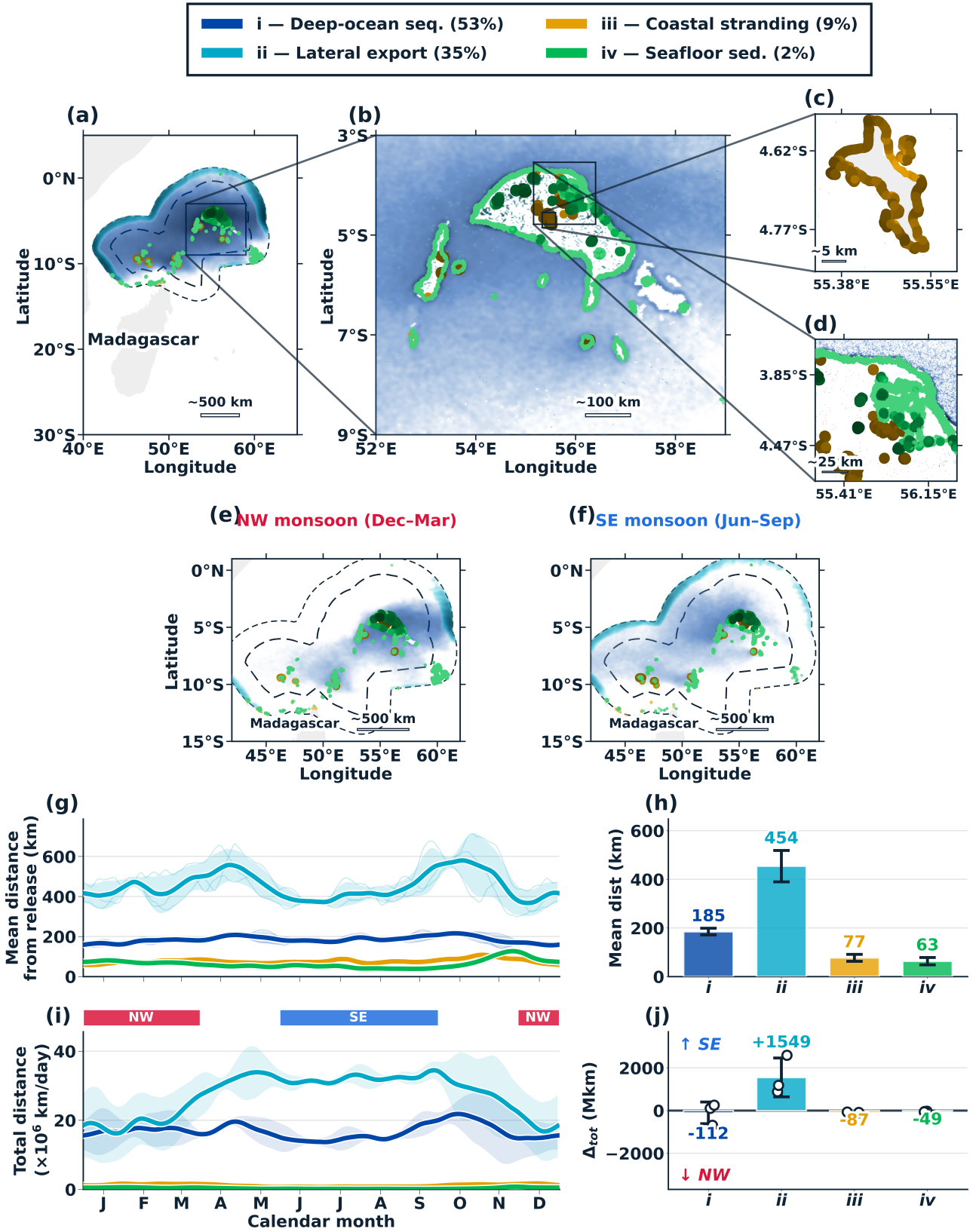


Figure 32: Spatial distribution of particle terminal fates across the Seychelles EEZ and surrounding buffer. (a–f) Seeding sites coloured by terminal fate, with progressive zooms on the Mahé Plateau (a–d) and the Aldabra/Farquhar/Providence outer islands (e–f). (g) SE monsoon (May–Sep) and (h) NW monsoon (Dec–Mar) endpoint density of released particles, shading from low to high.

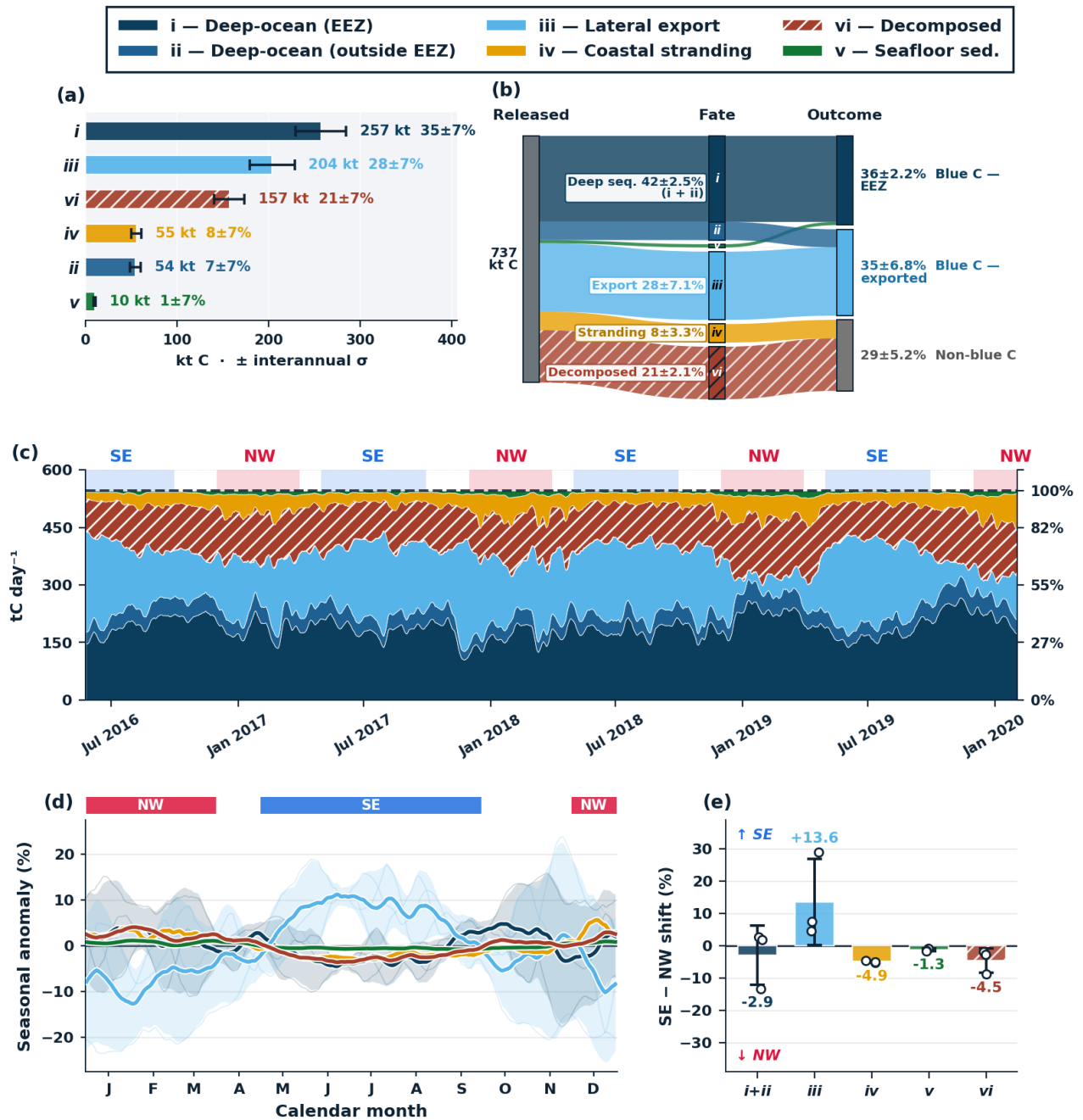


Figure 33: Domain-wide carbon mass budget over all Seychelles meadows, July 2016 to January 2020. (a) Mean share of the released carbon reaching each terminal fate, with interannual spread. (b) Sankey diagram grouping fates into three outcome categories. (c) Daily stacked composition of the fate budget. (d) Monthly departure from the annual mean per fate, bold curve for the mean across years and lighter curves for individual years on the high-variance fates. (e) SE minus NW shift per fate.

The carbon view adds a sixth fate, decomposition in transit (*vi*), which removes $21 \pm 7\%$ of the released carbon (157 kt C) before each particle reaches one of the five other sinks (Figure 33a). This loss scales with depth and transit time, and reshapes every other share. This does not account for the decomposition of stranded particles on beaches, which falls under coastal stranding (*iv*). With this in mind, deep ocean sequestration leads at $42.0 \pm 2.5\%$ of the released carbon (311 kt C combined), split into $35 \pm 7\%$ inside the EEZ (*i*, 257 kt C) and $7 \pm 7\%$ in the surrounding buffer (*ii*, 54 kt C). Lateral export (*iii*) follows at $28 \pm 7\%$ (204 kt C), coastal stranding (*iv*) holds at $8 \pm 7\%$ (55 kt C) and seafloor

sedimentation (*v*) drops to $1 \pm 7\%$ (10 kt C). Regrouped along the blue carbon accounting boundary (Figure 33b), the released carbon splits almost evenly into three outcomes. The pool potentially creditable to the Seychelles national stock gathers deep ocean sequestration inside the EEZ (*i*) and the in-EEZ portion of seafloor sedimentation (*v*), reaching $36.0 \pm 2.2\%$. The pool exported beyond national borders gathers deep ocean sequestration in the buffer (*ii*), lateral export (*iii*) and the buffer portion of seafloor sedimentation (*v*), reaching $35.0 \pm 6.8\%$. The non-blue residual, made of coastal stranding (*iv*) and decomposition (*vi*), closes the budget at $29.0 \pm 5.2\%$.

Figure 33c stacks the carbon released on each day, coloured by the fate it eventually reaches. The anti-correlation between deep ocean sequestration in the EEZ (*i*) and lateral export (*iii*) seen in the particle view carries over here. The new feature is decomposition (*vi*), which peaks around January 2019 when lateral export (*iii*) reaches its lowest. Figure 33d gives the monthly climatology with the same pattern as before, the two most variable fates again carrying the seasonal cycle. Figure 33e condenses this into a single seasonal shift per fate. Lateral export (*iii*) reaches 13.6 % in SE, combined deep ocean sequestration (*i+ii*) -2.9% , decomposition (*vi*) -4.5% , coastal stranding (*iv*) -4.9% and seafloor sedimentation (*v*) -1.3% . All these swings are smaller than in the particle view because decomposition smooths the seasonal contrast, eroding the long-transit particles that carry the extremes.

How long particles take to reach each fate varies considerably across the five pathways (Figure 34). Coastal stranding (*iv*) is the fastest at 14 days on average, deep ocean sequestration (*i*, *ii*) and lateral export (*iii*) follow at 23 to 24 days, and seafloor sedimentation (*v*) is the slowest at 115 days. This last value is counter-intuitive but reflects the trigger criterion: a particle within 1 m of the seabed must wait for the near-bed current to stay below the threshold U_c for seven consecutive days before settling, which lengthens the effective time to fate. The fraction lost to decomposition along the way scales accordingly. Particles bound for stranding lose only 11 %, though they eventually decompose on the coast. Those bound for deep ocean and lateral export lose 20 % to 23 %, and those bound for seafloor sedimentation 54 %, the slow shelf drift leaving decomposition the longest window to act.

The carbon partition is now resolved at domain scale, but it gathers all meadows into a single average. Figure 35 shows how the position of each meadow across the Seychelles EEZ shapes its contribution, mapping the percentage of the carbon emitted per m^2 of seagrass that ends up credited

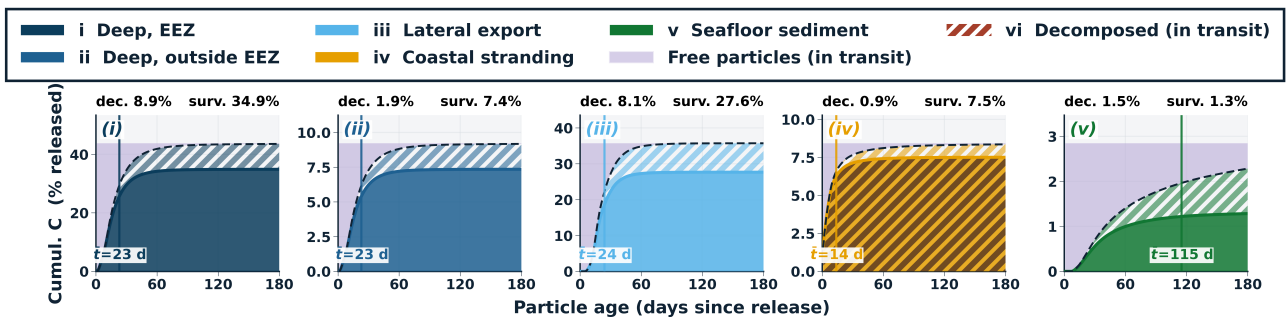


Figure 34: Time-to-fate per terminal fate (*i*)–(*v*). Cumulative share of released carbon (%) against particle age (days). Solid colour shows remaining carbon at the fate, hatched overlay the carbon decomposed in transit before arrival, and purple shading the particles still in transit. The mean transit time $\bar{t}$ is marked as a vertical line.

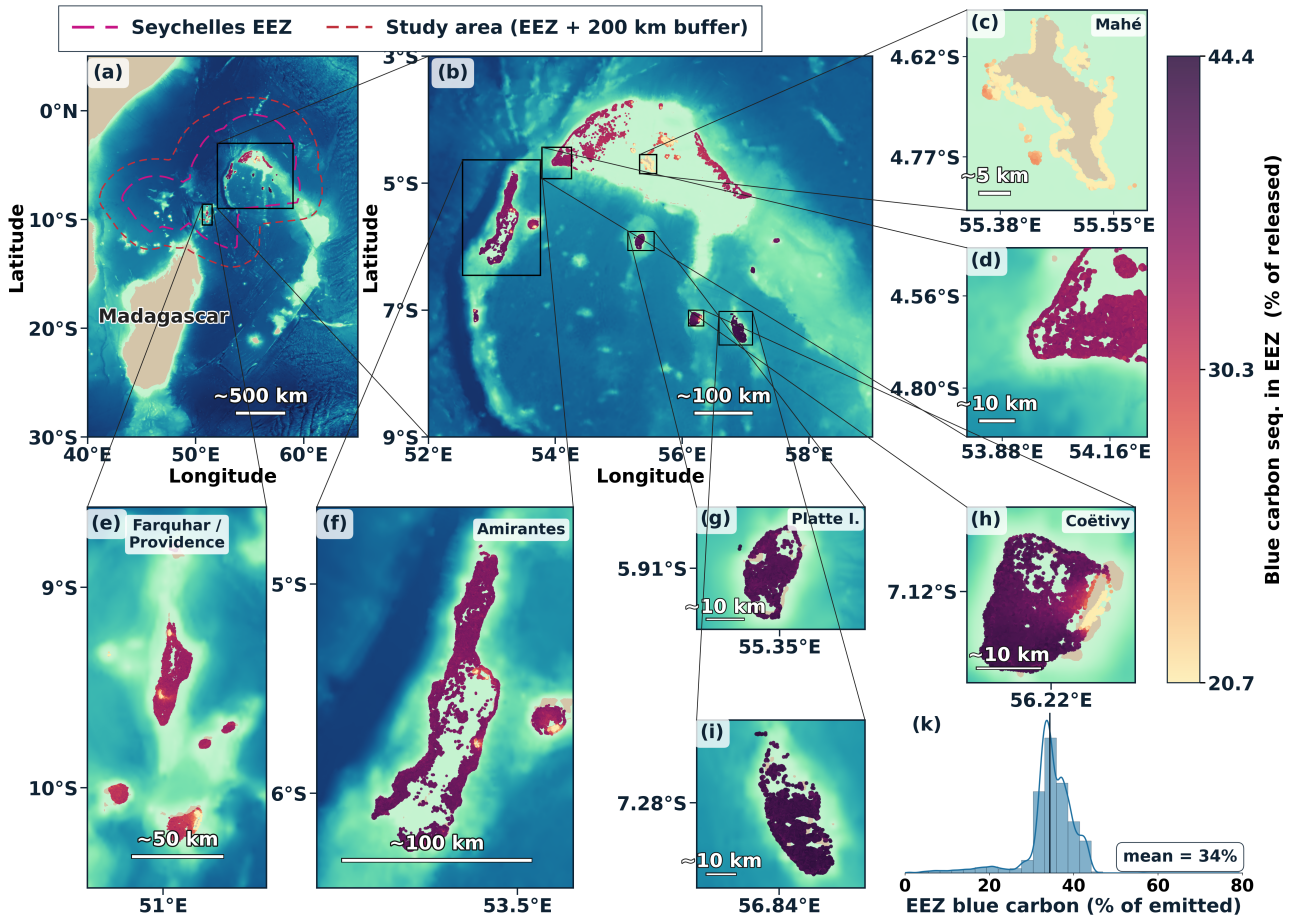


Figure 35: Per-meadow fraction of released particulate carbon credited to in-EEZ Blue Carbon sequestration, expressed as a percentage of the carbon emitted per m² of seagrass. Colour scale (logarithmic) spans near zero to 44.4 %, bounded by the 5th and 99.8th percentiles of the per-element distribution. (a) Regional view with Madagascar, the Seychelles EEZ (magenta dashed line) and the study area (EEZ + 200 km buffer, red dashed line). (b) Mahé Plateau overview. (c) Mahé (lowest in-EEZ share). (d) Western shelf-break meadows (highest share). (e) Farquhar and Providence atolls. (f) Amirantes. (g) Platte I. (h) Coëtivy. (i) Small isolated bank east of Coëtivy. (k) Distribution of the per-element in-EEZ share weighted by seagrass area, with area-weighted mean 34 % (vertical line).

to in-EEZ blue carbon sequestration. The area-weighted mean reaches 34 % (Figure 35k), with most meadows in the upper range of the colour scale. Distance to the nearest shelf break drives the contrast. Mahé (Figure 35c), set in the inner plateau, contributes only 21 %. Its distance from the shelf traps particles in shallow water, where coastal stranding and in-transit decomposition consume the remaining 79 %. Meadows at the shelf edge (Figure 35d) and isolated reefs (Platte I., Figure 35g, and Coëtivy, Figure 35h) export most efficiently. This is probably due to two combined effects: particles reach sequestration depths before decomposing, and the scarcity of downstream coastline keeps coastal stranding (*iv*) low. The Amirantes (Figure 35f) and Farquhar/Providence (Figure 35e) match this high-export regime across most of their extent. The main Mahé Plateau (Figure 35b) spans the full range, with Mahé itself low and the outer-plateau reefs already approaching the outer-island values.

Figure 36 summarises this section and brings the answer to the central question of this thesis. The annual carbon partition (Figure 36c) gives 35.9 % of the exported POC reaching long-term sequestration inside the EEZ as blue carbon creditable to the Seychelles, 35.3 % sequestering beyond

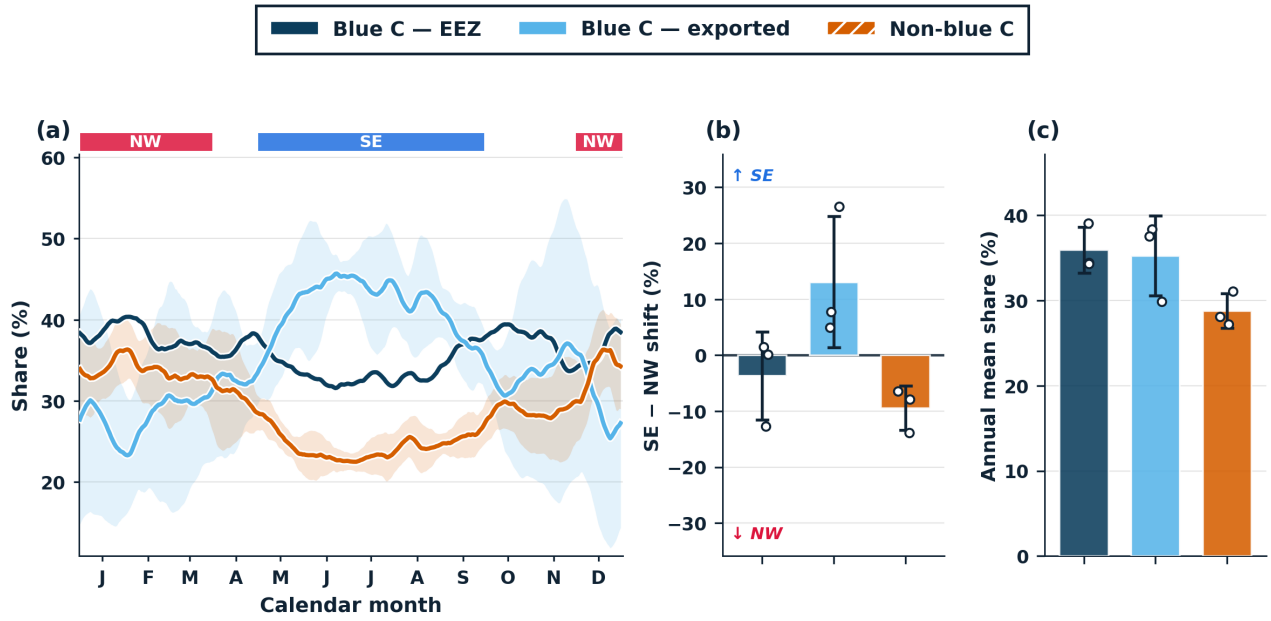


Figure 36: Seasonal climatology of the three grouped carbon outcomes (Blue C in-EEZ, Blue C exported, Non-blue). (a) Monthly share of released carbon averaged across full years, with monsoon windows shaded (NW in red, SE in blue). (b) SE minus NW shift per outcome ($\pm$ interannual standard deviation). (c) Annual mean per outcome with interannual spread.

the national borders and falling outside the national accounting, and 28.8 % lost to coastal stranding or in-transit decomposition. The creditable pool stays flat across the year, oscillating between 32 % and 40 % in the monthly climatology (Figure 36a) and shifting by only -3.7 pp between the two monsoons (Figure 36b). The monsoon signal lives entirely in the other two outcomes, which swing in opposite phase: lateral export climbs from 25 % in February to 45 % in June–August (13.1 pp in SE), while the non-blue residual peaks near 35 % in February and loses 9.4 pp in SE. The Seychelles-creditable estimate of 36 % is therefore not a seasonal average masking large excursions but a structurally stable figure, robust across the monsoon cycle. In absolute terms, this share sets the order of magnitude of the creditable flux at 72 kt C yr^{-1} . A sensitivity analysis of the four conversion factors (Section 4.3.4) refines this to a median of 53 kt C yr^{-1} (95 % CI 16 to 122 kt C yr^{-1}).

4.3.3 Sensitivity analysis

The results of this thesis depend on the parameters that produced them. Three are examined here, all controlling how much carbon each particle loses along its trajectory and how mass is redistributed between fates: the sinking threshold z'_{seq} , the decomposition rate k_0 , and the depth attenuation λ . Resolving their impact requires the full Lagrangian history of every particle, so the LPT was rerun over the year 2018 (27 % of the full simulation window) with continuous trajectories saved for post-processing (Figure 37). The four remaining flux-scaling parameters (carbon content f_C , POC export ratio r_{exp} , leaf production P , meadow area A) are propagated in the Monte Carlo envelope of Section 4.3.4.

Figure 37a varies the sequestration depth threshold z'_{seq} , set by hypothesis at 1,000 m of the deep export in (Duarte and Krause-Jensen, 2017). Moving it from 250 to 3,000 m mainly redistributes

mass between combined deep ocean sequestration (*i+ii*, ~48 to ~32 %) and lateral export (*iii*, ~22 to ~33 %), as fewer particles count as sequestered and a larger share drifts further to reach the open boundary. In policy terms, the total blue carbon sequestered stays roughly constant, but its geographic attribution shifts toward more export beyond the **EEZ**. Decomposition (*vi*) barely moves because the attenuated rate $k_0 \exp(-\lambda z)$ vanishes below a few hundred metres, so deepening the threshold only reshuffles where the surviving particles end up, not how much organic matter is lost in transit, while coastal stranding (*iv*) and seafloor sedimentation (*v*) shift by less than 3 %.

Figure 37b varies the surface decomposition rate k_0 across the uniform distribution used in the model, $\mathcal{U}(0.006, 0.017 \text{ d}^{-1})$, raising the decomposed share (*vi*) from ~13 to ~28 %, drawn roughly equally from combined deep ocean sequestration (*i+ii*, ~47 to ~39 %) and lateral export (*iii*, ~31 to ~25 %), since faster breakdown consumes particles before they can either settle into deep water or exit the domain.

Figure 37c varies the depth attenuation λ , whose impact stays within a few percentage points of the central value. The decomposition rate $k_0 \exp(-\lambda z)$ is close to k_0 throughout the upper 100 m where most particles spend the bulk of their transit, regardless of λ . By 500 m, the rate has already dropped to roughly 10 % of k_0 for any plausible λ . Changing λ therefore only modulates an intermediate band that contributes little to the total mass loss.

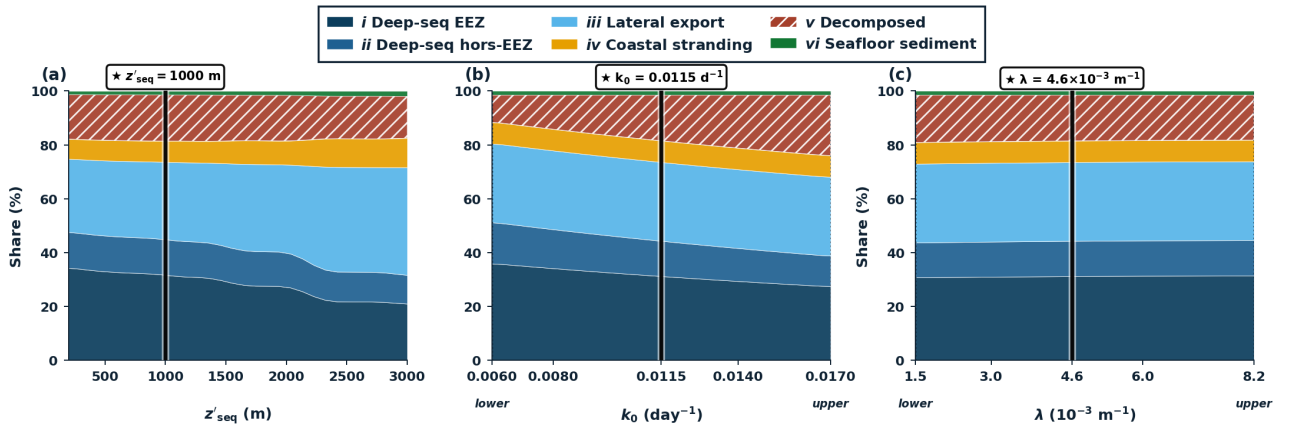


Figure 37: Sensitivity of the partition between fates to the three decomposition parameters. (a) Sequestration depth threshold z'_{seq} , (b) surface decomposition rate k_0 , (c) depth attenuation coefficient λ , swept across their plausible range with the central value marked by a star.

4.3.4 Uncertainty envelope

The 72 kt C yr⁻¹ figure of Section 4.3.2 should be read as one realisation of the postprocessing chain rather than as the potentially creditable flux itself. Section 3.6 already flagged that the scaling factors used to convert the **LPT** output into carbon mass carry sizeable input uncertainty, so any single number from the chain stands for a particular choice of parameter values rather than a final estimate. The 72 kt C yr⁻¹ value corresponds to each scaling factor held at its mode. For example, in the absence of any local substrate survey, every pixel left unclassified by the Allen Coral Atlas inherits $P_{\text{rock}} = 25.5 \text{ g m}^{-2} \text{ d}^{-1}$. Following Section 3.6, the analysis here maps how the spread of the four input distributions *a-d* of Figure 24 propagates into the result rather than refining this single

point estimate. Leaf production P dominates, ahead of the **POC** export ratio r_{exp} , because the rock default channels P to the upper bound of its admissible range over a sizeable fraction of the domain.

Propagating the four input distributions of Section 3.6 through Eq. (25) shifts the central value to a median of 53 kt C yr⁻¹ (193 kt CO₂e yr⁻¹), with a 95 % confidence interval of 16 to 122 kt C yr⁻¹ (59 to 447 kt CO₂e yr⁻¹, Figure 38e). The deterministic value therefore overshoots the median by 36 %, and the envelope spans a factor of 7.6 between its lower and upper bounds. The overshoot is structural rather than physical. The triangular distribution on P has its mode at the upper bound 25.5 g m⁻² d⁻¹, because no higher *T. ciliatum* value is reported and the rock default channels every unclassified pixel to that mode, so any random draw can only revise P downward (Figure 38c). The three other distributions are symmetric (Figures 38a, b, d) and widen the envelope without shifting the median. The 53 kt C yr⁻¹ median is therefore the more defensible central estimate, and the 36 % gap is the price of the rock-default assumption.

The envelope captures only the variability of the four scaling factors. The hydrodynamic and Lagrangian uncertainties left aside in Section 3.6 would widen it further, so the reported interval is a lower bound on the full uncertainty around the creditable flux.

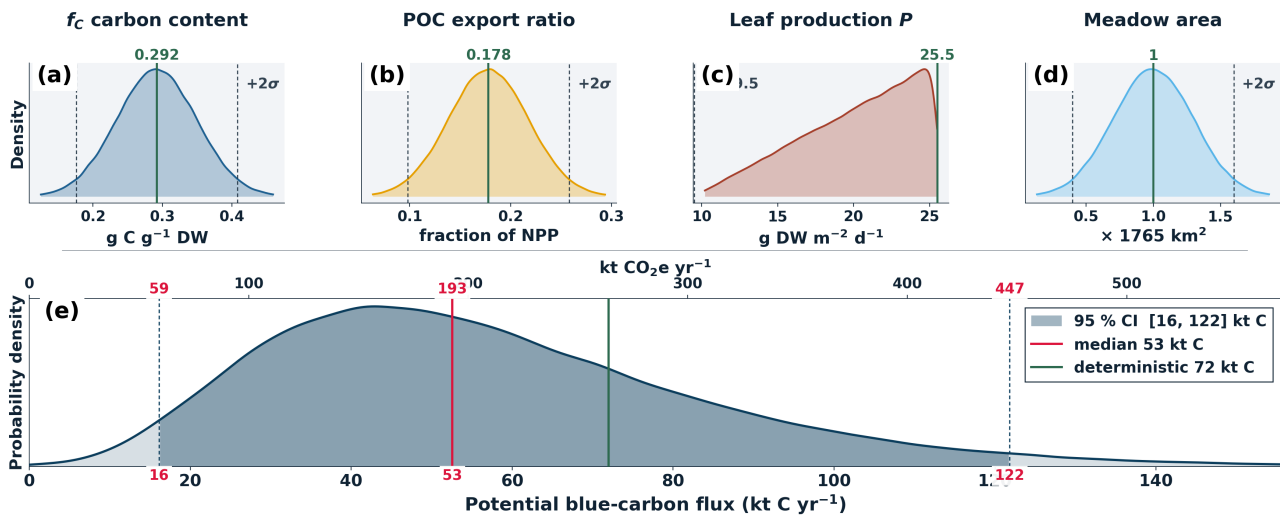


Figure 38: Monte Carlo propagation of input uncertainty on the in-EEZ potential blue-carbon flux (200,000 draws). (a) Carbon content f_C , (b) **POC** export ratio r_{exp} and (d) meadow area follow normal distributions. (c) Leaf production P follows a triangular distribution. Solid line marks the central value, dashed lines the $\pm 2\sigma$ bounds (normal) or the min/max bounds (triangular). (e) Output distribution with the median (red), the 95 % confidence interval (shaded) and the full simulated range. Top axis converts kt C to kt CO₂e.

5 Discussion

At the national scale of the Seychelles, the exported flux is large enough to reorder the carbon budget. Long-term sequestration captures 71 % of the exported pool, split between 36 % inside the **EEZ** and 35 % just as durably beyond national borders, with 29 % lost to coastal stranding and in-transit decomposition. Under the model's assumptions, the Monte Carlo median creditable share reaches 53 kt C per year (193 kt CO₂e, range 59 to 447 kt CO₂e on the four flux parameters of Section 4.3.4, with hydrodynamic uncertainty and interannual scatter not propagated).

Compared to the Seychelles' annual emissions (Rowlands et al., 2024), this median creditable export of 193 kt CO₂e per year would approach the national transport sector and reach 40 % of the energy sector. Priced at the voluntary blue-carbon range of 15 to 35 USD per tonne CO₂e (S&P Global Commodity Insights, 2025), the median could potentially represent, if ever recognised, 3 to 7 million USD per year. For comparison, the 2018 Sovereign Blue Bond raised US\$15 million over 10 years (about US\$1.5 million per year on average) under a partial World Bank guarantee to fund marine conservation in the Seychelles (World Bank, 2018), so the exported pool alone could yield two to five times the annual flow of that instrument. The pathway aligns with the 2021 NDC commitment to integrate seagrass into the National Greenhouse Gas Inventory by 2025 (Rowlands et al., 2024) and could help reach the 2050 net-zero target. Whether this estimate can be trusted depends first on how well the underlying circulation is reproduced.

Hydrodynamic validation is necessary but not sufficient for the realism of the simulated dispersal. The tide propagates correctly to the plateau centre (Pointe La Rue, **RMSE** 8.9 cm, $R = +0.97$, Figure 25). The inner-plateau **ADCP-E** matches the depth-averaged speed within 6 % and reproduces the vertical structure, while the shelf-break **ADCP-N** carries a 24 % speed bias, a -57° rotation and five-fold excess energy (Figure 26), reflecting the difficulty of resolving the shelf-break edge in the model over a 110-day record. Offshore, the northern **RAMA** buoy reaches $R \approx 0.6$ on daily u, v (Figure 27), with weaker agreement at the southern buoy. Hourly circulation errors are assumed to have only a limited impact on the partition: fate proportions aggregate 253 million trajectories over 3.5 years, so pointwise misfits are assumed to average out and only systematic biases distort the partition. At the regional scale, **SLIM3D** reproduces the seasonal switch of Castillo-Trujillo et al. (2021): a clockwise recirculation around the Mahé Plateau during the NW monsoon gives way to an **EEZ**-wide westward jet during the SE monsoon (Figure 29), the very contrast that drives the seasonal partition examined next.

Focusing on the monsoon impact on the simulated particles reveals that they reshuffle non-creditable fates while leaving the potentially creditable in-**EEZ** sequestration share nearly fixed (Figure 36b). Between NW (December–March) and SE (June–September), the in-**EEZ** sequestration share shifts by only -3.7 ± 0.9 % against 13.6 ± 1.3 % for lateral export (mean $\pm$ interannual standard deviation across the 3.5 simulated years). The mechanism is a trade-off between the two non-potentially-creditable allochthonous fates. The intensified SE westward jet carries particles off the Mahé Plateau and out of the **EEZ**, so lateral export rises and the non-blue residual falls. During the NW monsoon, the gyre retains particles longer over the plateau, so coastal stranding and in-transit decomposition rise while lateral export drops. Total deep-ocean sequestration is therefore higher in SE, driven by the

gain in lateral export, even though the creditable in-EEZ share itself barely moves. The monsoon thus sets where non-creditable particles leak, not how much carbon stays potentially creditable. In the results of this study, geographic controls (meadow location relative to the western jet and to the EEZ boundary) appear to have a larger impact than the seasonal cycle, and are examined next.

The simulation suggests, hypothetically, that meadow location shapes the potentially creditable in-EEZ share more than the monsoon does. A meadow's position relative to the shelf break would govern the fate of its detached POC: shelf-edge meadows release POC close to deep water, while inshore meadows lose more to coastal stranding and decomposition. The narrow outer platforms of Platte I. (Figure 35g) and Coëtivy (Figure 35h) deliver simulated particles to the shelf edge within days, with a short downstream coastline. Inner-plateau Mahé, lacking both features, reaches only 21 % of its exported carbon staying in-EEZ blue carbon, against an area-weighted mean of 34 % across the Seychelles (as a reminder, in this paragraph we focus on in-EEZ share only, without the lateral export). The in-EEZ creditable share of meadows located on the Mahé Plateau itself spans near zero to 44.4 %, with shelf-break meadows crediting higher shares than those close to Mahé island (Figure 35b). The proposed control would be hydrodynamic rather than purely geometric. *T. ciliatum* fragments stay positively buoyant for an extended floating phase, drifting at the surface where their fate hinges on the currents they meet, not on their straight-line distance to the shelf break. Shelf bathymetry would set those currents, steering surface flow, organising shelf-edge jets and shaping retention gyres. Distance to deep water would then act through two channels: geometry, by setting whether deep water is reachable in a reasonable transit time, and circulation, by shaping the surface flow that carries particles there. A median NW-monsoon residence time of 94 days over the Mahé Plateau leaves ample room for stranding and remineralisation when surface currents recirculate inshore. Whether these shares translate into actual creditable carbon depends on two further controls examined next: the chosen sequestration boundary and the depth threshold.

For sediment cores and in-meadow methodologies, only the carbon physically in place is credited. Applying the same logic to the exported pool and counting only particles whose terminal fate falls inside the EEZ yields a potentially creditable share. This share is shaped by two factors. The first is administrative: the EEZ boundary is a legal line with no physical counterpart (Luisetti et al., 2020). The second is internal to the LPT model. Storage time lengthens steadily with depth, from about 100 years at 100 m to over 1,300 years at 1,000 m (Stukel et al., 2023), and atmospheric isolation reaches the century scale near 1,000 m (Duarte and Krause-Jensen, 2017; Ricour et al., 2023). The model encodes this continuum through a fixed horizon z'_{seq} , and the 1,000 m value simply picks a line on it. Moving z'_{seq} from 250 to 3,000 m drops the potentially creditable in-EEZ share from 40 % to 24 %, a 16-percentage-point transfer absorbed by lateral export (Figure 37a), so two analysts running identical physics report different national sinks. Beyond these internal choices, no external framework currently credits the exported pool at all.

For now, this flux remains outside the national inventory: no operational Monitoring, Reporting and Verification (MRV) methodology currently credits exported seagrass detritus to its source meadow (IPCC, 2014; Houston et al., 2024). Two obstacles sustain the gap. The scientific literature on POC export quantification remains thin, with no consensus protocol for fluxes leaving the meadow. Attribution then compounds the difficulty: sediment cores keep carbon in place and yield direct

stock estimates, whereas exported POC disperses in the water column and travels with the currents, leaving no routine tracer to link a downstream particle back to a specific meadow. Closing the gap will require both better methodology and more *in situ* measurements of the model inputs, which are examined next.

The estimate presented in this work carries several limitations, grouped into six families that condition the budget. (a) *Input parameters*. Leaf productivity rests on a single *T. ciliatum* measurement at Inhaca, Mozambique (Bandeira, 2002), applied uniformly in time and depth, and correcting its upward bias would plausibly lower the creditable flux. The 17.8 % POC-to-NPP ratio is a cross-species global mean (Duarte and Krause-Jensen, 2017) applied uniformly to every meadow, and the carbon content f_C is taken from the Seychelles seagrass inventory (Rowlands et al., 2024) and applied uniformly to all meadows, with no species-specific variability. The decomposition rate k_0 rests on proxies, beach-stranded leaves decaying in air (Ochieng and Erftemeijer, 1999) and a global seagrass average (Duarte and Krause-Jensen, 2017), so in-water decay could run faster and erode more of the exported carbon than assumed. (b) *Particle behaviour*. Rising velocity is estimated from a buoyancy-drag balance rather than measured (Appendix D), and settling and windage are borrowed from other species for want of *T. ciliatum* values. Decomposition follows a simple first-order law uniform across fragment sizes. (c) *Constant-rate forcing*. Production is held constant year-round and detachment is spread evenly across days, omitting storm pulses and seasonal canopy turnover. (d) *Coverage and substrate*. Meadows are assumed monospecific *T. ciliatum* on high-density polygons, missing substrate classifications default to rock (a 2.7-fold upward bias on the affected elements), and anthropogenic disturbance is not represented. (e) *Hydrodynamics*. In-situ validation data are too sparse to fully constrain the SLIM3D circulation that drives the LPT: the shelf-break ADCP-N record is limited in duration (Figure 26), offshore currents are available only at 10 m (RAMA) and shelf currents only depth-averaged (ADCP). (f) *Statistics*. Interannual scatter (30 % to 40 %) exceeds the seasonal signal, and the 2016–2020 window is too short to average out the underlying climate oscillations, so a longer record would be needed to ensure that the partition reported here is not over-represented by the few years simulated. These limitations apply specifically to the Seychelles case, and testing the framework elsewhere will help separate model-specific from generic uncertainties.

This work gives a first estimate of the blue carbon exported from Seychelles *T. ciliatum* meadows. It must be read with caution, since the key inputs (decomposition and productivity) rest on sparse data and better measurements would refine it. Even so, the particle view shows that the exported fraction is substantial, and the carbon sequestered this way each year is not negligible. This makes the export pathway worth investigating in other seagrass regions, where the modelling chain (LPT driven by SLIM3D) can be applied with species-specific tuning of a handful of parameters (rise velocity, sinking velocity, windage, detachment phenology), leaving the coupling and the fate accounting unchanged. Replicating the experiment across archipelagos and shelves with contrasted bathymetry (Gullström et al., 2002; Larkum et al., 2018) would make it easier to assess whether the Seychelles' specific bathymetry plays a strong role in the partition. This is already suspected here, since the archipelagic geometry naturally limits coastal stranding compared to a continuous coastline. Two measurements would help most on the Seychelles side: site-level *T. ciliatum* productivity on the Mahé Plateau to

constrain the absolute flux, and in-water decomposition rates for the same species. Beyond the model itself, policy and **MRV** remain a separate bottleneck before any of these fluxes can be credited.

6 Conclusion

Marine ecosystems are now recognised as a complementary lever to emission cuts in mitigating climate change. Among them, blue carbon ecosystems sequester carbon for centuries to millennia in their sediments. Seagrass meadows act as a carbon sink in two ways: locally through burial beneath the canopy, and remotely through the export of detached fragments to sedimentation and the deep ocean. So far, only the buried fraction has been counted in greenhouse-gas inventories. This study addresses this gap for the Seychelles, estimating the share of exported carbon that reaches long-term sequestration and could potentially be added to the national blue carbon budget.

The **SLIM3D** hydrodynamic model was coupled with a Lagrangian particle tracking module developed for this study, and run over the Seychelles Archipelago and the surrounding deep ocean from January 2016 to July 2020. The hydrodynamic core was validated against tide gauges, **ADCP** and **RAMA** data, and reproduces the seasonal reversal between the NW and SE monsoons that controls plateau circulation, with a residual bias at the shelf-break **ADCP**-N station. The Lagrangian module resolves a full particle lifecycle (rising, floating, sinking, decomposition, stranding), with behavioural parameters drawn from sparse *T. ciliatum* data or taken from congeneric seagrasses. 253 million particles were released across the national seagrass map, weighted by the seagrass fraction of each mesh element, and tracked to five terminal fates: deep-ocean sequestration inside the **EEZ** (i), deep-ocean sequestration beyond the **EEZ** (ii), lateral export (iii), coastal stranding (iv) and seafloor sedimentation (v), with in-transit decomposition added as a sixth, non-terminal fate. The relative budget is then scaled to absolute carbon mass through *T. ciliatum* leaf productivity, the **POC** export ratio, the leaf carbon content and the mapped meadow area, with a Monte Carlo analysis propagating parameter uncertainty. With this framework in place, the research question can now be answered.

What fraction of the particulate organic carbon exported from *T. ciliatum* meadows of the Seychelles reaches long-term sequestration?

Across the 3.5 simulated years, 36.0 ± 2.2 % of the exported carbon settles inside the **EEZ** and could potentially be added to the Seychelles blue carbon contribution, 35.0 ± 6.8 % settles just as durably beyond national borders, and 29.0 ± 5.2 % is lost to coastal stranding and in-transit decomposition (mean $\pm$ interannual σ). Long-term sequestration therefore captures 71 % of the exported pool, equivalent to 12 % of meadow **NPP**.

Meadow location drives the potentially creditable share more than any other control under the model's assumptions. The share spans near zero to 44.4 % across the archipelago, with an area-weighted mean of 34 %. Inner-plateau meadows around Mahé credit only 21 %, their distance from the shelf break trapping particles in shallow water where local currents tend to push them onto the island's coast. Outer-island reefs at Platte Island and Coëtivy, together with the shelf-edge fringe of the Mahé Plateau, reach the upper range as detached leaves cross into deep water within days over short downstream coastlines. Proximity to the shelf break sets most of the inter-meadow contrast. Compared to countries with a continuous continental coast, the archipelagic geometry of the Seychelles is in itself a likely advantage: scattered narrow shelves around individual islands limit

the downstream coastline available for stranding, whereas a continuous coast would give exported POC far more time and surface to wash ashore.

Seasonality has little impact on the simulated allochthonous carbon potentially creditable inside the EEZ, but a larger impact on the lateral export. The in-EEZ potentially creditable share drops by only 3.7 percentage points from the NW to the SE monsoon, holding close to its annual value across both seasons. Lateral export gains 13.6 percentage points in SE as the intensified westward jet flushes particles off the plateau before they settle, so the total deep-ocean fraction rises in SE even as the potentially creditable part barely moves.

In mass terms, annual export from *T. ciliatum* meadows reaches 147 kt C (Monte Carlo median, 95 % CI 44 to 339), of which 104 kt C reach long-term sequestration and 53 kt C (193 kt CO₂e, CI 16 to 122) fall inside the potentially creditable boundary. Against the buried-only national baseline of 510 kt CO₂e per year, the modelled flux would add 12 % at the lower bound, about a third at the median, and nearly double it at the upper bound. No operational MRV framework currently credits exported carbon to its source nation, so this flux remains outside the Seychelles inventory. Under a future framework recognising allochthonous sequestration, and priced at the voluntary blue-carbon range of 15 to 35 USD per tonne CO₂e (S&P Global Commodity Insights, 2025), the median creditable flux would represent 3 to 7 million USD per year (full uncertainty interval 1 to 16 million USD).

Several inputs are still drawn from sparse evidence and could be refined: leaf productivity transposed from a Mozambican site, in-water decomposition inferred from beach-stranded leaves, and behavioural parameters carried over from other seagrass species. *In situ* *T. ciliatum* productivity and in-water decomposition rates measured on the Mahé Plateau would help anchor the absolute estimate. Beyond the Seychelles case, the same SLIM3D–LPT chain could be transferred to other seagrass species through species-specific tuning of the behavioural parameters, leaving the coupling and the fate accounting unchanged. Applying it to other archipelagos could test whether the bathymetric control on the potentially creditable share, suggested here through the link between shelf-break proximity and deep-ocean export across the Mahé Plateau, generalises beyond the Seychelles, and could open the framework to seagrass-rich coastlines facing the same gap in their carbon accounting.

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7.1 Bibliography

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A Bathymetric data correction

Sounding depths are interpolated onto each mesh node using the compactly supported weighting function

$$\phi(d) = \begin{cases} \exp\left(-\frac{1}{1 - (d/r)^2}\right), & d < r, \\ 0, & d \geq r, \end{cases} \quad (28)$$

where d is the distance between a mesh node and a sounding and $r = 5,000$ m is the radius of influence. For each mesh node, the interpolated depth is the weighted average of all soundings falling within r ,

$$h_{\text{RBF}} = \frac{\sum_{j \in N} \phi(d_j) h_j}{\sum_{j \in N} \phi(d_j)}, \quad (29)$$

where N is the set of soundings within radius r of the node and h_j the measured depth of sounding j . Mesh nodes with no sounding within r cannot be interpolated and retain the raw GEBCO depth. To avoid a discontinuity at the boundary between the two zones, the RBF estimate and the GEBCO value are blended as

$$h = \alpha h_{\text{RBF}} + (1 - \alpha) h_{\text{GEBCO}}, \quad \alpha = \min\left(\frac{d_{\text{edge}}}{0.5 r}, 1\right), \quad (30)$$

where d_{edge} is the distance from the interpolated node to the nearest node that falls back on GEBCO. Deep inside the sounding coverage, $d_{\text{edge}} \geq 0.5 r$ and $\alpha = 1$, so the node takes the pure RBF estimate. Close to the edge of the coverage, α decreases linearly down to zero, ensuring a smooth transition to the GEBCO background.

B Distance-based mesh size function

The coastline and seagrass boundary size functions increase linearly with the distance d to the nearest boundary, from a minimum size imposed close to the boundary to a maximum size far away:

$$\text{size}_{\text{dist}}(d) = \begin{cases} \text{size}_{\text{min}}, & d \leq d_{\text{min}}, \\ \text{size}_{\text{min}} + \frac{d - d_{\text{min}}}{d_{\text{max}} - d_{\text{min}}} (\text{size}_{\text{max}} - \text{size}_{\text{min}}), & d_{\text{min}} < d \leq d_{\text{max}}, \\ \text{size}_{\text{max}}, & d > d_{\text{max}}. \end{cases} \quad (31)$$

Elements are kept fine within d_{min} of the boundary and coarsen linearly up to size_{max} beyond d_{max} . Values of $(d_{\text{min}}, d_{\text{max}}, \text{size}_{\text{min}}, \text{size}_{\text{max}})$ for the coastline and seagrass functions are reported in Table 2.

C Bathymetric gradient mesh size function

The third mesh size function refines elements at bathymetric slope breaks based on the local gradient of water depth:

$$\text{size}_{\text{grad}}(\mathbf{x}) = \begin{cases} 15\,000 \text{ m}, & |\nabla h| \leq 10^{-5}, \\ \max\left(1\,600 \text{ m}, 0.8 \frac{h}{|\nabla h|}\right), & |\nabla h| > 10^{-5}, \end{cases} \quad (32)$$

where h is the local water depth and $|\nabla h|$ its horizontal gradient magnitude. Where the gradient exceeds 10^{-5} , the element size scales with $h/|\nabla h|$, the horizontal length scale of the slope, ensuring that the abrupt shelf break of the Mahé Plateau is adequately resolved, which is critical for capturing the export of POC towards the deep ocean.

D Rising velocity estimate

No direct measurement of w_{rise} exists for detached *T. ciliatum* fragments. A first-order estimate is derived from a buoyancy–drag balance at terminal velocity for a flat plate rising broadside to the flow:

$$\Delta\rho\,t_\ell\,A\,g = \frac{1}{2}\,C_D\,\rho_w\,A\,w_{\text{rise}}^2 \implies w_{\text{rise}} = \sqrt{\frac{2\,\Delta\rho\,t_\ell\,g}{C_D\,\rho_w}}, \quad (33)$$

where $\Delta\rho = \rho_w - \rho_{\text{leaf}}$ is the density deficit between seawater and the leaf tissue, t_ℓ the blade thickness, A the projected area, C_D the drag coefficient, and ρ_w the seawater density. Using the degassed *P. oceanica* tissue density $\rho_{\text{leaf}} \approx 960 \text{ kg m}^{-3}$ (Johnson et al., 2017) as an interspecific proxy gives $\Delta\rho \approx 65 \text{ kg m}^{-3}$ with $\rho_w = 1,025 \text{ kg m}^{-3}$. With $t_\ell \approx 5 \times 10^{-4} \text{ m}$ (Brodersen et al., 2018) and $C_D = 1.28$ for a flat plate normal to the flow (Hoerner, 1965), Eq. (33) yields $w_{\text{rise}} \sim 2 \times 10^{-2} \text{ m s}^{-1}$.

Given the reliance on an interspecific density proxy, a generic blade thickness, and an idealised drag coefficient, a conservative $\pm 50\%$ envelope is adopted as a type B uncertainty around the central estimate (Morgan and Henrion, 1990), giving $w_{\text{rise}} \in [1 \times 10^{-2}, 3 \times 10^{-2} \text{ m s}^{-1}]$. For meadow depths of 9 m to 50 m on the Mahé Plateau, this corresponds to surface arrival times of 5 min to 85 min, two to three orders of magnitude shorter than the multi-week floating phase. Most seagrass detritus models therefore initialise particles directly at the surface (Hanuise et al., 2025). Particles are here released from the seabed to explicitly resolve the rising phase.

E Mean rotation angle

On the Seychelles Plateau, Alma Carolina Castillo-Trujillo et al. (2021) analysed bottom-mounted ADCPs using the complex-correlation framework introduced by Pijush Kundu (1976). The same framework is adopted here, applied to the comparison between observed and modelled horizontal currents. Expressed through their eastward (u) and northward (v) components, the mean rotation angle is defined as

$$\arg(\rho_c) = \arctan\left(\frac{\overline{u_o\,v_m - v_o\,u_m}^t}{\overline{u_o\,u_m + v_o\,v_m}^t}\right), \quad (34)$$

where $\overline{\cdot}^t$ denotes the time average. The numerator is the time-mean of the cross product between observed and modelled vectors and the denominator the time-mean of their scalar product. The sign of $\arg(\rho_c)$ tells how the modelled current is on average oriented with respect to the observed one: a positive value means it points to the left of the observation, and a negative value to the right. $\arg(\rho_c)$ is reported on the rose panels (k, l) of Figure 26.

ABSTRACT

Marine ecosystems are increasingly recognised as a complementary lever to emission cuts in climate change mitigation. Seagrass meadows sequester carbon locally through burial beneath the canopy and remotely through the lateral export of detached fragments to the deep ocean and sedimentation. National inventories credit only the buried share. This study aims to quantify the missing lateral pathway for *Thalassodendron ciliatum* meadows in the Seychelles, coupling the multiscale hydrodynamic model SLIM¹ with an offline Lagrangian particle tracking module developed for this thesis that releases 253 million *T. ciliatum* propagules between 2016 and 2020.

Long-term sequestration captures 71 % of the exported carbon under the model's assumptions, equivalent to 12 % of meadow net primary production. Within this pool, 36.0 ± 2.2 % settles inside the Exclusive Economic Zone (EEZ), 35.0 ± 6.8 % settles beyond national borders with comparable residence times, and 29.0 ± 5.2 % is lost to coastal stranding and in-transit decomposition. Meadow location shapes the within-EEZ share more than the monsoon does, ranging from near zero to 44.4 % across the archipelago.

Estimated through a Monte Carlo approach on meadow production, the median within-EEZ flux reaches 53 kt C per year (193 kt CO₂e, full range 59–447 across production parameters). This flux is not yet recognised in carbon accounting frameworks, yet it is estimated to be substantial: it would raise the buried-only national baseline of 510 kt CO₂e by 38 %, approach the national transport sector, and reach 40 % of the energy sector. Under emerging voluntary blue-carbon markets, such a flux could also be valorised. More broadly, this thesis underlines why carbon export from seagrass meadows warrants closer investigation as a component of blue-carbon accounting.

¹ Second-generation Louvain-la-Neuve Ice-ocean Model.