

Bioengineering Faculty

Master thesis :

Seamless water quality modelling under extreme weather conditions: Linking land nitrogen outputs to coastal ecosystem health

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Abstract

Tropical cyclones can generate major hydrological and biogeochemical disturbances in coastal ecosystems through extreme rainfall, river discharge, and associated nutrient inputs. These processes are particularly relevant for the Great Barrier Reef (GBR), where coral reefs and seagrass meadows are exposed to variations in coastal water quality. This study investigates how TC-associated nitrogen inputs and hydrodynamic processes shaped phytoplankton dynamics during and following TC Jasper and TC Kirrily in the 2023–24 wet season, and the resulting potential ecosystem exposure. A three-dimensional hydrodynamic model of the GBR region was developed using SLIM3D and coupled to a simplified NPZD biogeochemical model incorporating dissolved inorganic nitrogen inputs from four rivers affected by the two cyclones. Model performance was evaluated against hydrodynamic observations, satellite-derived chlorophyll-*a*, and in situ DIN measurements. Simulated phytoplankton concentrations were then used to quantify cumulative exposure of coral reefs and seagrass meadows, complemented by a River–No-river sensitivity experiment. The hydrodynamic model reproduced the main regional circulation and hydrographic structures, whereas the biogeochemical model underestimated the observed post-Jasper coastal DIN enrichment and phytoplankton response. Nevertheless, contrasting exposure patterns emerged between the two cyclone periods. Exposure increased strongly in the northern GBR following Kirrily, with mean values in Cape York South reaching 123.3 mmol N m⁻² d for coral reefs and 103.7 mmol N m⁻² d for seagrass meadows, compared with 47.7 and 21.4 mmol N m⁻² d, respectively, following Jasper. This intensification coincided with an episodic intrusion of DIN-rich subsurface waters and hydrodynamic conditions potentially favouring local retention. Overall, cyclone-driven phytoplankton exposure in the GBR appears to result from the interaction of riverine nitrogen inputs, the pre-existing biogeochemical state, and hydrodynamic processes controlling nutrient supply, transport, and retention.

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Contributions

	Personal contribution	Other contributors	Artificial intelligence
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Methodology	X ¹	X ^{2,3,4,5}	X ⁶
Investigation	X ¹	X ^{2,5}	X ⁶
Formal analysis	X ¹	X ^{2,4}	X ⁶
Validation	X ¹	X ^{2,4}	X ⁶
Writing – original draft	X ¹		X ⁷
Writing – review and editing	X ¹	X ²	X ^{6,7}
Visualization	X ¹		X ⁶
Supervision	X ¹	X ²	

1. **Charles Lavry** (UCLouvain), master's student and author of the thesis.
2. **Emmanuel Hanert** (UCLouvain), supervisor, provided scientific guidance and advice throughout the project, helped orient methodological and scientific choices, supervised the development of the work, and reviewed parts of the manuscript.
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List of Acronyms

ACE	Accumulated Cyclone Energy
AIMS	Australian Institute of Marine Science
BOM	Bureau of Meteorology
COTS	Crown-of-Thorns Starfish
DIN	Dissolved Inorganic Nitrogen
DO	Dissolved Oxygen
DON	Dissolved Organic Nitrogen
EAC	East Australian Current
ERA5	ECMWF Reanalysis 5th Generation
GBR	Great Barrier Reef
GloFAS	Global Flood Awareness System
GPC	Gulf of Papua Current
IMOS	Integrated Marine Observing System
IPCC	Intergovernmental Panel on Climate Change
JCU	James Cook University
N	Nitrogen
NPZD	Nutrient–Phytoplankton–Zooplankton–Detritus
NRM	Natural Resource Management
NRMSE	Normalised Root Mean Square Error
P	Phosphorus
P90	90th Percentile
PAR	Photosynthetically Active Radiation
PI	Potential Intensity
PN	Particulate Nitrogen
PON	Particulate Organic Nitrogen
QC	Quality Control
RMSE	Root Mean Square Error
SEC	South Equatorial Current
SLA	Sea Level Anomaly
SLIM	Second-generation Louvain-la-Neuve Ice-Ocean Model
SS	Suspended Sediment
SST	Sea Surface Temperature
TC	Tropical Cyclone
WMIP	Water Monitoring Information Portal
WMO	World Meteorological Organization

1 | Introduction

1.1 Extreme weather events and land-based nutrient export

Tropical cyclones (TCs) are intense atmospheric disturbances that develop from a localized low-pressure system forming over warm tropical or subtropical waters. They are characterised by an organized circulation that rotates counterclockwise in the Northern Hemisphere and clockwise in the Southern Hemisphere [Shultz et al., 2005]. The term TC encompasses hurricanes, typhoons and cyclones, which correspond to regional denominations respectively used for the North Atlantic and Northeastern Pacific regions, the Northwestern Pacific region and the Indian Ocean and South Pacific regions [Neumann, 2007].

TCs often combine diverse hazards such as intense winds, heavy rainfall, and storm surges, which in turn trigger many consequential events (Fig. 1). Storm surges, which arise from the combined action of low pressure and extreme winds, constitute the most destructive aspect of TC to coastal regions. They trigger various devastating impacts resulting from an excess of seawater reaching the land, such as flooding and direct damages of coastal areas, or saline contamination of soils. Extreme winds also greatly affect coastal regions, causing damage to structures and vegetation, as well as frequent loss of power and communication, recurrent fire hazards and direct casualties. Finally, the intense rain brought by TC is the factor affecting the most inland regions, as it comes with flooding in many areas, ground subsidence, direct destruction of vegetation and deterioration of water quality [Anthes, 1982]. Taken together, these numerous hazards place TC as the second most dangerous natural disaster worldwide, after earthquakes [WMO, 2022]. On a yearly basis, TC affect between 6 and 12 % of the world population and cause an estimated \$15 billion in damage [Emanuel, 2020; Jing et al., 2024].

A statistically significant decline in the global TC number has been reported since 1990, accompanied by a reduction in overall accumulated cyclone energy (ACE). However, this overall trend is largely influenced by observations from the Western North Pacific and does not reflect a weakening storm intensity [Klotzbach et al., 2022]. In fact, over the same period, the maximum intensity of major TC (classified as Cat. 3 or higher) has increased [Wang et al., 2020]. This is consistent with the global rise in potential intensity (PI) calculated from climatological data, which represent the maximum theoretical intensity reachable in a given environment. Higher PI implies a greater incidence of intense storms [Emanuel, 2020], as evidenced by the observed 15% increase in the probability of

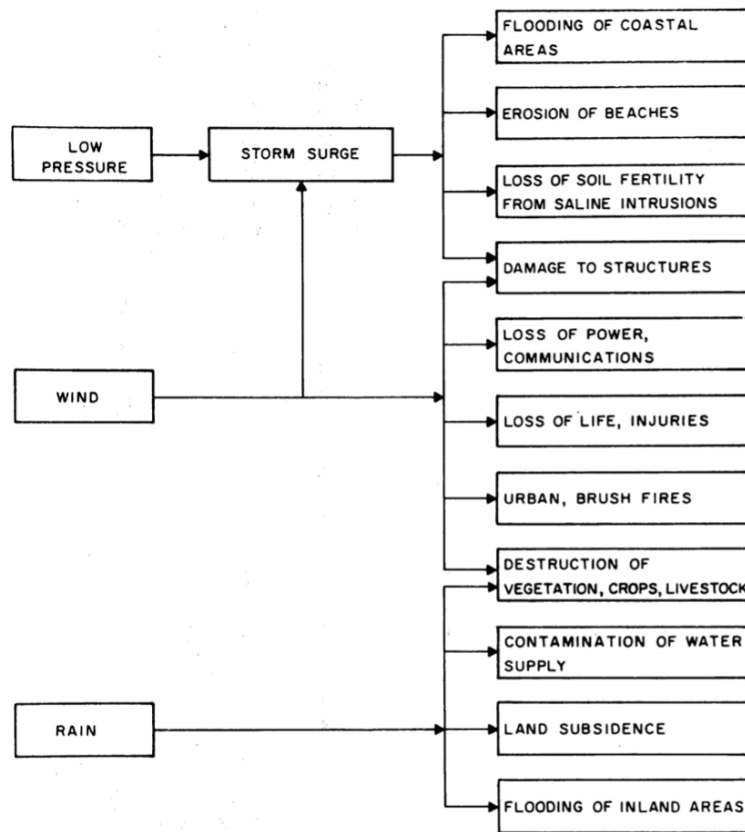


Figure 1: Types of potential damages accompanying TCs [Anthes, 1982].

major TC occurrence over the last 40 years [Dunne, 2020]. Furthermore, over the past 20 years, the global population exposed to TC has nearly doubled, and nearly two-thirds of this increase is attributed to changes in cyclone activity itself, rather than population growth alone [Jing et al., 2024]. Regarding low-intensity TC, no significant trend toward intensification has been identified, particularly in the Northern Hemisphere [Ma et al., 2025]. However, this observation remains relatively secondary, as it is mainly severe TC — although much less frequent — that are responsible for the vast majority of material damage and human losses worldwide [Emanuel, 2020; Pielke et al., 2008].

The evolution of TC activity takes place in the context of ongoing global warming driven by increasing atmospheric greenhouse-gas concentrations. The resulting rise in global temperatures warms the upper ocean and modifies atmospheric large-scale circulation patterns and thermodynamic conditions [Calvin et al., 2023]. Global warming also promotes the occurrence and intensity of marine heat waves, leading to positive SST anomalies [Prochaska et al., 2023]. Although important uncertainties remain regarding the future evolution of TC activity, projected changes in environmental conditions are expected to influence cyclone characteristics. In particular, SST plays a central role in TC formation and intensification, as it controls the ocean–atmosphere heat and moisture fluxes that fuel TC development [Feng et al., 2023]. Future projections suggest an increase in the proportion of storms reaching very high intensity levels (Cat. 4–5), with a projected median increase of approximately 13 %. Global average TC intensity is also expected to slightly increase, together with TC-associated precipitation rates, which may rise by a median of approximately 14 % under ongoing climate warming. Despite this projected intensification of the strongest events, the

overall global frequency of TC events may decrease [Knutson et al., 2019]. In view of the projected changes and the remaining uncertainties surrounding the future evolution of TC activity, continued monitoring and understanding of these events remain essential in a warming climate.

Australia is among the countries most frequently affected by TC, with approximately eleven events recorded every year — accounting for 13% of the worldwide total. About half of these systems reach the severe intensity threshold [Bureau of Meteorology, 2025b]. TC occurrence is nevertheless strongly seasonal, with most events occurring between November and April, when oceanic and atmospheric conditions are most favourable for cyclone formation. On average, three to four cyclones make landfall annually, causing significant impacts on the population, infrastructures and the local ecosystems [Bureau of Meteorology, 2025a]. The ecological disturbances are particularly critical along the northeastern coast of Australia, where the Great Barrier Reef (GBR) hosts highly vulnerable coastal and marine ecosystems. It is estimated that around 25 % of the yearly TCs pass within the GBR Marine Park [Earth Systems and Climate Change Hub, 2021].

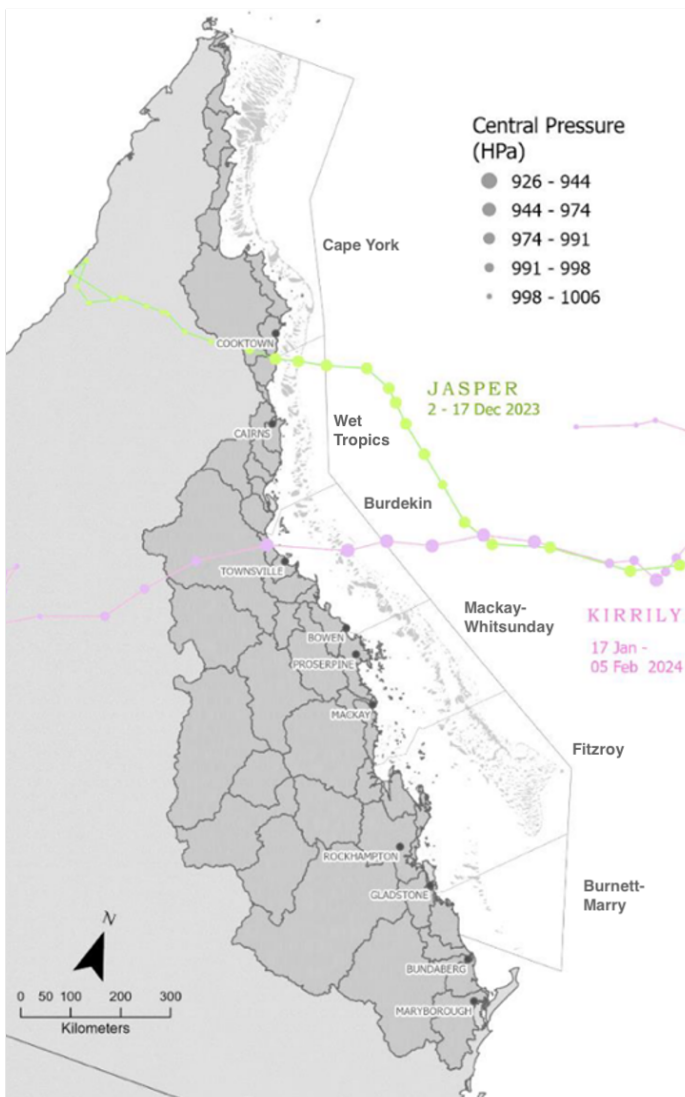


Figure 2: Tracking of Jasper and Kirrily TCs [Moran et al., 2025].

During the 2023–24 TC season, two systems made landfall along Queensland’s eastern coast, where the GBR is located: Cyclone Jasper and Cyclone Kirrily (Fig. 2). The two cyclone trajectories crossed the northern sector of the GBR catchment, encompassing the Cape York, Wet Tropics, and Burdekin Natural Resource Management (NRM) regions [Moran et al., 2025]. Second category TC Jasper crossed the coast about 120 km north of Cairns, on December 13th. Shortly after making landfall, it began to weaken rapidly, dropping below the threshold of a Cat. 1 system [Gurtner, 2024]. The resulting low-velocity tropical low then stalled over the region for five days, with persistent and substantial rainfall leading to major floods and landslides [Beadle, 2024]. Some studies have suggested that Jasper may represent the wettest TC ever recorded in Australia [Emanuel, 2024]. One month later, TC Kirrily made landfall near Townsville on January 25th [Gurtner, 2024]. As it moved inland, the Cat. 1 system lost intensity, forming a tropical depression that persisted more than ten days over the region [Bureau of Meteorology, 2024b].

Sudden and intense precipitation accompanied both TCs, although the rainfall events associated with TC Jasper were markedly more extreme. Locally, daily rainfall totals exceeded 850 mm during TC Jasper, whereas TC Kirrily produced isolated daily falls above 200 mm [Bureau of Meteorology, 2024a]. These extreme precipitation events contributed to marked positive rainfall anomalies during the 2023–24 wet season across several northeastern Australian catchments. Rainfall anomalies were particularly pronounced in Cape York and the Wet Tropics, where wet-season rainfall exceeded the 30-year long-term average, with mean daily anomalies reaching up to 15 mm day^{-1} near the regional boundary. Rainfall in the Burdekin region remained broadly consistent with long-term climatological conditions, despite this region corresponding to the coastal landfall area of TC Kirrily. At the basin scale, the largest positive anomalies in Cape York were observed in the Endeavour and Jeannie basins, where mean daily wet-season rainfall exceeded the historical average by approximately 5 and 7 mm day^{-1} , respectively. Even stronger anomalies occurred in the Wet Tropics, where the Daintree and Mossman basins both recorded increases of more than 11 mm day^{-1} relative to the long-term wet-season mean [Moran et al., 2025].

These precipitations had a direct impact on the volume of freshwater entering the Reef lagoon. In the Cape York and Wet Tropics regions, mean annual river flows respectively reached 2.6 and 2.1 times their long-term median, while total Reef-wide discharge rose to 1.7 times the historical average. These values correspond to the highest discharge levels ever recorded for Cape York and the highest observed in the Wet Tropics region in the past 13 years [Moran et al., 2025]. Although rainfall in the Burdekin region remained close to the long-term seasonal average, the Burdekin River nonetheless exhibited a marked hydrological response to TC Kirrily, owing to the northward extent of its catchment. As a result, river discharge increased sharply during the event, reaching a peak of approximately 210,000 ML day^{-1} in early February [Queensland Government, n.d.].

During high-flow conditions generated by extreme rainfall events, riverine loads of pollutants (sediments, nutrients and pesticides) rise sharply, as surface runoff mobilizes and transports significant quantities of contaminants from adjacent lands to local waterways. For the GBR catchment, such episodic flood pulses concentrate the bulk of annual freshwater, sediment and nutrient exports within brief intervals [Furnas, 2003]. Once discharged into the lagoon, these waters form river plumes that can extend up to hundreds of kilometers alongshore, exposing inshore ecosystems to contaminated waters for periods of days to weeks. The magnitude and composition of these pollutant exports vary substantially among river systems, as they are strongly influenced by freshwater discharge and land-use patterns across the contributing catchments [Haynes et al., 2007]. Large mean annual sediment loads are generally associated with extensive basin areas and drier climatic regions. For nutrients, two distinct export patterns emerge. Particulate phosphorus (PP) and particulate nitrogen (PN) exports generally follow sediment load distributions. In contrast, dissolved inorganic nitrogen (DIN) loads are mainly controlled by basin size and the extent of intensive fertilized cropping areas [Lewis et al., 2024]. Land use varies considerably across GBR catchments. Cape York NRM remains the most preserved region, with approximately 61% of conservation areas. Wet Tropics NRM region presents a significant proportion of intensive agricultural areas, particularly sugarcane and banana cultivation. Finally, the Burdekin region extends further inland and is

largely dominated by extensive grazing lands, covering approximately 88% of its area. Nevertheless, intensive cropping areas, including horticulture and sugarcane, are also present along the coastline [Queensland Land Use Mapping Program, 2020].

The temporal evolution of total nitrogen (N), phosphorus (P) and suspended sediment (SS) daily loads at the outlets of the Barron and Burdekin Rivers clearly shows the influence of TCs on pollutant delivery to the GBR (Fig. 3). The Barron and Burdekin Rivers were strongly affected by TC Jasper and TC Kirrily, respectively. The two time series show a sharp and rapid increase in contaminant exports following the passage of each cyclone. The Barron River displays the largest load peaks: daily sediment loads reach nearly 90,000 t around December 18th, while the maximum daily loads of total nitrogen and phosphorus peak at approximately 400 and 100 t, respectively. For the Burdekin River, a pronounced but more moderate peak is observed around January 27th, consistent with the lower rainfall totals associated with TC Kirrily. Daily sediment loads reach a peak of around 30,000 t, while total nitrogen and phosphorus peak at around 150 and 50 t, respectively.

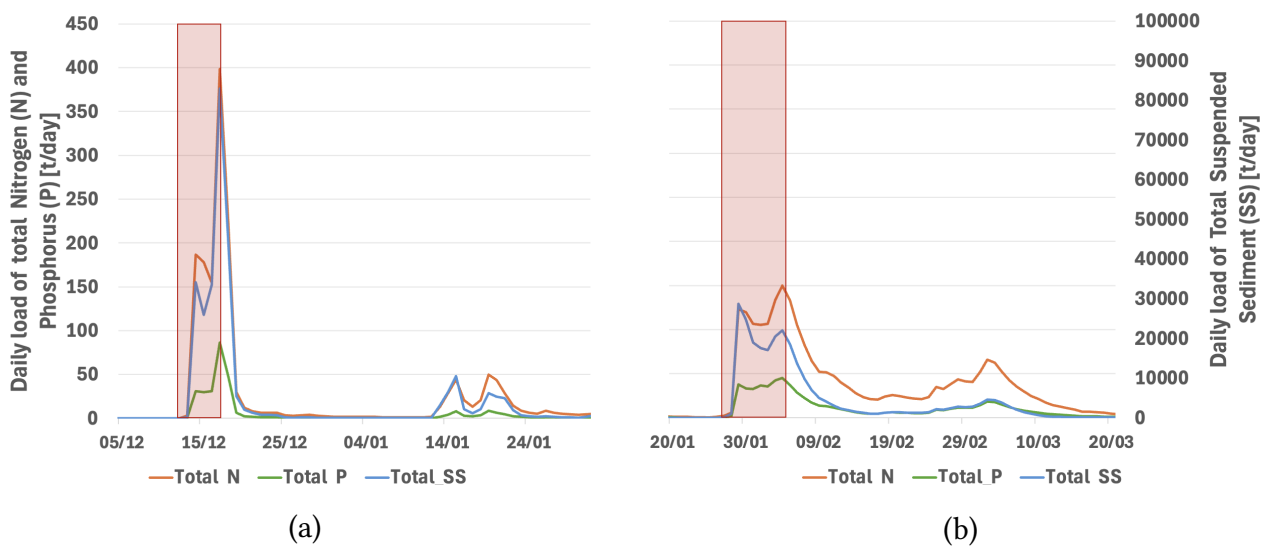


Figure 3: Evolution of daily loads of total nitrogen (N), phosphorus (P) and suspended sediment (SS) at the outlets of the (a) Barron and (b) Burdekin Rivers. The shaded red areas indicate the periods associated with TC influence. Data source: Queensland Government, Department of the Environment, Tourism, Science and Innovation, 2026.

1.2 Coastal ecosystems under pressure

Located in northeastern Australia, the GBR extends for more than 2,300 km along the Queensland coast and contains the world's largest coral reef system [UNESCO, n.d.]. It is one of the most diverse marine ecosystems on Earth, hosting about 1 620 species of fish, more than 400 types of hard coral, six of the world's seven species of marine turtle, one of the richest dugong populations, and numerous other marine species [Richards & Day, 2018]. The GBR is not a single environment, but a mosaic of interconnected habitats, including coral reefs, seagrass meadows, and estuarine systems with mangrove forests [Schaffelke et al., 2005]. Among these, coral reefs and seagrass meadows are the most spatially extensive. The more than 2,900 individual coral reefs span an area exceeding

24,000 km², while seagrass meadows cover an estimated 35,000 km² of the seafloor, though their coverage varies seasonally [McKenzie et al., 2024].

Fig. 4 illustrates the spatial distribution of coral reef and seagrass habitats across the GBR region. Coral reefs form a near-continuous belt along the Queensland coastline, extending from the northern tip of Queensland to Bundaberg in the south. In the northern sector, reefs are densely distributed and remain relatively close to the shoreline, whereas south of Townsville they extend progressively into offshore waters. Coastal seagrass meadows are more heterogeneously distributed along the coastline, with notable concentrations in the north of Cooktown, and in the southern part of the region. Deepwater seagrass meadows are most prominent in two areas: in the central GBR, where they form broad bands interspersed with coral reefs, and in the southern GBR, where a large deepwater seagrass bed lies offshore from the Gladstone region. A third concentration is found at the northern extremity of the GBR, near the coastline.

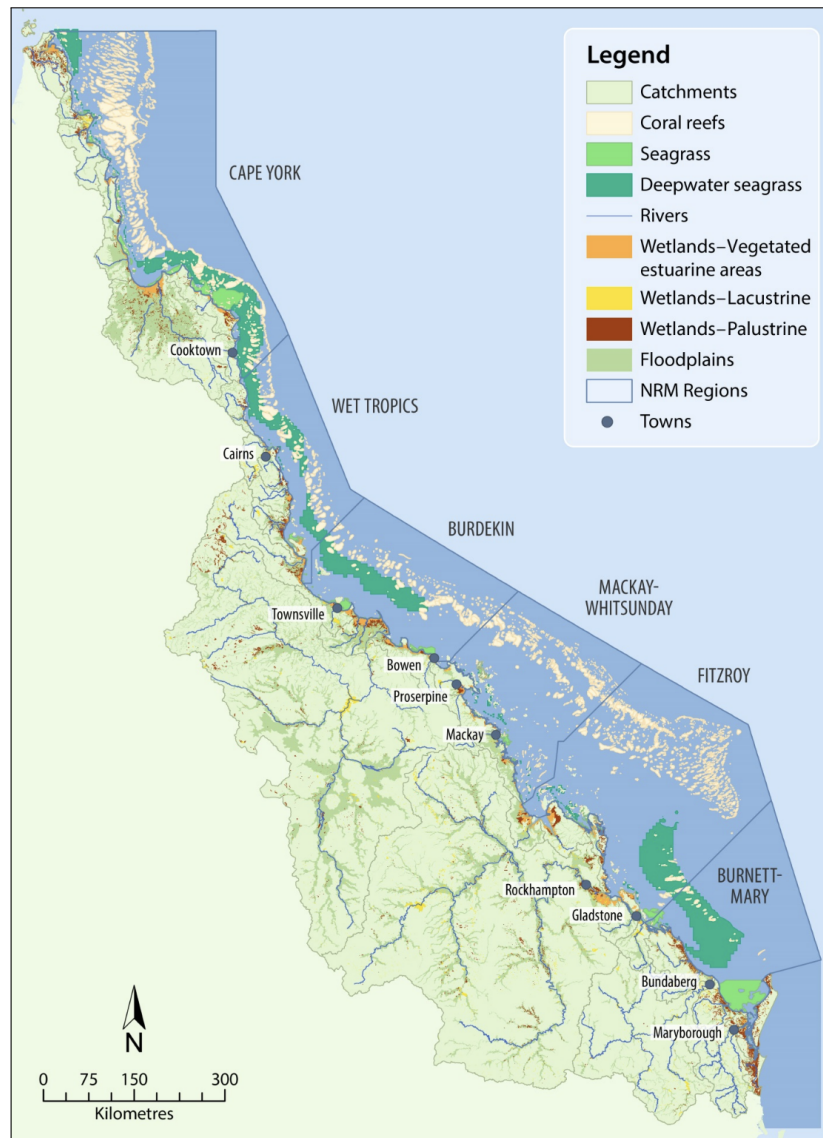


Figure 4: Spatial distribution of coral reefs and seagrass meadows in the GBR region [Schaffelke et al., 2017]

Corals fall into two broad groups: hard corals, also known as reef-building corals, and soft corals. Reef-building corals are responsible for the formation of coral reefs, as their polyps extract calcium from the ocean in order to build a rigid limestone outer skeleton [Reef Authority, 2024]. These polyps live in a mutualistic symbiosis with a microalga called zooxanthellae (*Symbiodinium*), that reside in their tissues. The algae supply an important part of coral's metabolic energy needs by translocating part of its photosynthesis production. In return, polyps provide the zooxanthellae inorganic nutrients and a safe anchor point [Roth, 2014].

Although coral reefs occupy only 7% of the Marine Park and the World Heritage Area, they support an exceptionally diverse array of marine life [Reef Authority, 2022]. By providing feeding, breeding, spawning and nursery grounds, they act as critical habitats for the reproduction, development and persistence of many aquatic species. Furthermore, because they are generally located in coastal environments, coral reefs provide a mechanical protection to the shoreline against erosion by waves, currents and storms [Moberg & Folke, 1999]. Beyond their ecological role, coral reefs are the primary driver of the GBR's status as a major tourist destination. During the 2023-24 season, tourism associated with the site generated an estimated \$8 billion, thereby contributing significantly to the Australian economy. Tourism-related activities also supported roughly 68,000 full-time jobs in the country [Ibrahim et al., 2025].

Seagrasses are flowering plants that grow fully submerged and form vast marine meadows. The GBR region hosts around 20% of the world's seagrass species [GBRMPA, 2024], and its meadows account for an estimated 13–22% of the global seagrass extent [McKenzie et al., 2020]. By providing a shelter and nursery for small fish, prawns, and numerous other marine organisms, seagrass meadows form a hub for biodiversity. They also constitute a major source of food for micro-, meso- and mega-herbivores, such as turtles and dugongs [Waycott et al., 2007]. Moreover, the conditions they create enable the recruitment of many other plant species [Schaffelke et al., 2005]. Beyond supporting marine fauna and flora, seagrass ecosystems also contribute to key biogeochemical processes, acting as nutrient and sediment filters. Indeed, during the process of nutrient absorption, these marine plants contribute to improving seawater conditions [Great Barrier Reef Foundation, 2021]. Finally, despite occupying less than 0.2% of the world's ocean area, seagrass meadows play a major role in the global carbon cycle by storing large quantities of organic carbon, mainly in the soils and sediments underlying the meadows. This high storage capacity results from both their high primary productivity and their ability to trap organic particles from the water column, while the oxygen-poor conditions of seagrass sediments slow decomposition and allow organic carbon to accumulate over long timescales. Globally, they may store up to 19.9 Gt of organic carbon and contribute approximately 10% of the annual organic carbon burial in the oceans [Fourqurean et al., 2012]. In terms of sequestration time, carbon retained in seagrass biomass may persist over years to decades, while carbon buried in sediments can be stored over millennial timescales [Duarte et al., 2005; Lo Iacono et al., 2008].

Although they play an essential role in the resilience of the entire region, these two ecosystems are fragile and show particularly worrying trends [GBRMPA, 2024]. Their capacity to resist and recover from disturbances is compromised by the cumulative effects of local stressors such as land-

based runoff, and broader global pressures such as climate change [McKenzie et al., 2024].

The main pressures affecting GBR corals include rising sea temperatures and marine heatwaves, TCs, Crown-of-Thorns Starfish (COTS) outbreaks, and ocean acidification [McKenzie et al., 2024]. Nearshore reefs, however, are also influenced by adjacent coastal river catchments, which deliver land-derived sediments, nutrients and other pollutants to GBR coastal waters [Schaffelke et al., 2005]. As a result, inshore corals are affected by land-based runoff and the associated degradation of water quality, particularly through increased turbidity, nutrient enrichment, and low-salinity flood plumes [A. Thompson et al., 2022]. These threats have led to variable evolutions of coral cover across the three regions of the GBR (Northern, Central and Southern regions). From 1985 to 2009, all regions showed mostly small fluctuations, with a stable trend for the northern and central regions and a gradually decreasing coral cover for the southern region. After 2010, unprecedented disturbances in both magnitude and frequency affected the GBR, causing a historic global decline in coral cover, with values reaching 10% [Emslie et al., 2024]. However, after years of cumulative impacts, the 2018–2023 period saw fewer disturbances affecting corals and led to substantial overall increase in coral cover up to 30–40% [GBRMPA, 2024]. This fast recovery is attributable to the presence of fast-growing corals, which have the disadvantage of being more sensitive to acute stressors (coral bleaching, TCs, COTS, and flood plumes) [GBRMPA, 2024; Mellin et al., 2019]. Finally, the 2023–2024 season was marked by the most spatially extensive mass coral bleaching ever recorded on GBR and by the occurrence of TC Jasper and Kirrily. The combined effects of these two severe disturbances, along with COTS outbreaks, led to a decline of 14–30% of the coral cover in 2025 compared to 2024 levels, depending on the region [AIMS, 2025].

Among the stressors affecting GBR seagrass meadows, TCs, land-based runoff, and thermally-driven stress from rising sea surface temperatures stand out as the most damaging [McKenzie et al., 2024]. Within this set of pressures, GBRMPA (2024) highlights poor inshore water quality as a primary threat, particularly when caused by land-based runoff or the wind-driven resuspension of sediments. This poor water quality affects seagrasses by reducing light availability for photosynthesis, either directly through suspended sediments and turbidity, or indirectly through nutrient-driven algal growth [Furnas, 2003]. From 2005 onwards, an overall decline in the Seagrass Index — a metric combining seagrass abundance and resilience — was observed across the GBR until the early 2010s. Whereas scores were initially generally described as ‘good’ (61–80), they crossed the ‘poor’ threshold (21–40) in all regions, with occasional drops to the ‘very poor’ category (0–20) in some regions [GBRMPA, 2024]. Subsequently, the Reef-wide index peaked around 60 in 2015, before declining over the following four years. Since then, seagrass dynamics have remained relatively stable, with the global index remaining slightly above 40 over the 2019–2024 period [McKenzie et al., 2025]. However, recent trends differ markedly at the regional scale. Cape York, the Wet Tropics and the Burdekin experienced marked declines in the Seagrass Index during the monitoring periods following TC Jasper and TC Kirrily, although the timing and drivers differed among regions. Conversely, Fitzroy and Burnett–Mary showed improving trends, while Mackay–Whitsunday remained relatively stable [Leahy et al., 2026].

1.3 Nitrogen biogeochemical cycle and its impact on marine ecosystems

Among the pollutants delivered during TCs, nitrogen plays a particularly critical role, as it is generally considered the major limiting nutrient for marine primary production in GBR waters [Waterhouse et al., 2024]. This sensitivity to nitrogen reflects the stoichiometric constraint of marine primary producers, known as the Redfield ratio. This ratio describes the relative proportions of carbon, nitrogen, and phosphorus in phytoplankton biomass or in multicellular marine plants. For marine phytoplankton, this ratio is C:N:P = 106:16:1 [Furnas, 2003]. Consequently, nitrogen inputs can stimulate primary production of phytoplankton and fast-growing pelagic macroalgae [Herbert, 1999]. As phytoplankton is responsible for the vast majority of primary production in marine waters [Moore et al., 2013], this stimulation can potentially disrupt the overall balance of coastal ecosystems [Herbert, 1999]. Describing the ecological impacts of nitrogen enrichment on coral reefs and sea-grass meadows therefore requires an understanding of the diverse nitrogen pools present in marine environments and of the processes through which they interact.

Nitrogen occurs in three major forms in marine environments. Dissolved inorganic nitrogen (DIN) encompasses nitrate (NO_3^-), nitrite (NO_2^-) and ammonium (NH_4^+) [Furnas, 2003], and represents the most immediately bioavailable nitrogen form [Brodie et al., 2013]. Dissolved organic nitrogen (DON) corresponds to the dissolved nitrogen associated with organic carbon and is the most important nitrogen pool in reef waters. Finally, particulate organic nitrogen (PON) represents the nitrogen associated with organic particles, such as detritus or living organisms, and is the second-largest pool of fixed nitrogen in shelf waters. Once delivered to coastal waters by rivers, part of this particulate nitrogen can be deposited in coastal sediments, where it may act as a nutrient reservoir [Furnas, 2003]. Together, these forms of nitrogen participate in a complex biogeochemical cycle [Furnas et al., 2011].

To represent the exchanges and transformations among inorganic nitrogen, organic nitrogen and living biomass, simplified Nutrient–Phytoplankton–Zooplankton–Detritus (NPZD) models can be used, in which dissolved nutrients assimilated by phytoplankton are incorporated into biomass, transferred to zooplankton through grazing, and partly transferred to the detritus pool through mortality and other biological losses, before being remineralized back into the available nutrient pool. Such frameworks were notably formalized by Fasham et al. (1990), who proposed a nitrogen-based formulation structured around seven interacting compartments: nitrate, ammonium, DON, phytoplankton, zooplankton, bacteria and detritus. This general NPZD structure therefore provides a conceptual basis for representing, in the GBR setting, how TC-driven riverine nitrogen inputs are assimilated and redistributed among biogeochemical compartments.

By presenting high temperatures and luminosity, the GBR waters provide favourable conditions for rapid phytoplankton growth once nutrients become available [Furnas et al., 2005]. However, during major floods, river plume dynamics strongly control how phytoplankton responds to nitrogen pulses. In young plumes, growth is initially restricted by high turbidity, which limits light

penetration to a shallow surface layer, even as nitrogen is readily available. As suspended sediments subsequently settle and turbidity declines, phytoplankton can assimilate this nitrogen and grow over an increasingly wide depth range, potentially leading to microalgal blooms within a few days [Furnas, 2003]. This accumulating phytoplankton biomass then increases light attenuation in the water column, reducing light-dependent primary production by seagrasses and coral-associated zooxanthellae [Dennison et al., 1993; Rogers, 1979]. Corals then receive fewer photosynthetic products from their symbionts, which can slow coral growth, induce tissue thinning and disrupt reproduction [Fabricius, 2005; Furnas, 2003]. Furthermore, under severe and prolonged light reduction, corals may lose their symbiotic algae, resulting in bleaching and, in the most affected cases, partial tissue mortality [Jones et al., 2020]. For seagrasses, reduced light availability is considered a major mechanism underlying their decline in Australia, because it directly limits effective photosynthesis [Walker & McComb, 1992].

Beyond its effect on light attenuation, nitrogen-driven phytoplankton production can also affect coral reef health by favouring outbreaks of COTS (*Acanthaster planci*), a coral-eating predator [Brodie & Alan W. Mitchell, 2005]. This link arises because COTS larvae depend directly on phytoplankton as a food source. When phytoplankton is abundant, conditions become more favourable for COTS larvae; their development accelerates and their survival rates are significantly increased [Fabricius, 2005].

In addition to these phytoplankton-mediated effects on corals and seagrasses, nitrogen enrichment can also affect coral physiology more directly. First, elevated DIN concentrations can disrupt the coral–algae symbiosis by impairing the ability of the coral host to regulate its population of algal symbionts. This can increase coral susceptibility to temperature-induced bleaching [Devlin & Brodie, 2023]. This mechanism is of particular concern, given that the 2023–24 TC season and associated major nitrogen inputs coincided with the most spatially extensive mass coral bleaching ever recorded on the GBR [AIMS, 2025]. Second, nutrient enrichment can also affect coral health by exacerbating coral disease dynamics [D’Angelo & Wiedenmann, 2014]. Elevated concentrations of nitrogen and other nutrients can increase both disease prevalence and severity [Bruno et al., 2003; Vega Thurber et al., 2014].

It is important to note that nitrogen is not inherently harmful to marine ecosystems. As an essential nutrient, it supports the growth and functioning of primary producers, including seagrasses and the photosynthetic symbionts associated with corals [Kopp et al., 2013; Touchette & Burkholder, 2000]. The ecological issue arises when nitrogen is supplied in excessive amounts and over short timescales. Under these conditions, nitrogen enrichment can disrupt ecosystem balance, with negative consequences for both seagrass meadows and coral reefs [Burkholder et al., 2007; Lesser, 2021].

1.4 Objectives of this thesis

Tropical cyclones are a major source of disturbance to coastal ecosystems. Major TCs have intensified in recent decades, and projections suggest further increases in high-intensity storms and TC-associated rainfall, generating major hydrological and biogeochemical disturbances through extreme rainfall, river discharge, and nutrient inputs. These events particularly affect Queensland, which is home to the GBR, one of the most biodiverse regions on Earth.

This master's thesis therefore seeks to address the following overarching question:

How do TC-associated nitrogen inputs and hydrodynamic processes shape the spatial and temporal dynamics of phytoplankton in the Great Barrier Reef during and following TC Jasper and TC Kirrily, and what potential exposure do these dynamics create for coral reefs and seagrass meadows?

To answer this question, this study will use a three-step approach. First, the hydrodynamics of the GBR region will be simulated using the SLIM 3D ocean model to represent the main circulations. Second, we will analyse the dispersion and evolution of the different nitrogen pools using a simplified biogeochemical model. Finally, simulated phytoplankton distributions are used to quantify the cumulative exposure of coral reefs and seagrass meadows over the two cyclone periods.

2 | Materials and methods

2.1 General approach

This study aims to investigate the impacts of the 2023–24 TC season and associated riverine nitrogen loads on the main coastal ecosystems of the GBR region through the simulation of phytoplankton bloom dynamics. To address this objective, a three-step modelling framework is implemented.

The first step consists of modelling of the hydrodynamics in the GBR region using Second generation Louvain-la-Neuve Ice-Ocean Model (SLIM). SLIM is an unstructured-mesh hydrodynamic model designed to simulate river network flows (SLIM1D), horizontal dynamics in shallow environments (SLIM2D), and more complex three-dimensional circulations (SLIM3D). Given the pronounced bathymetric variability and the complex oceanic processes of the region of interest, SLIM3D is used in this study. The computational domain is therefore discretized using a three-dimensional mesh. Higher resolution is allocated in areas of ecological and hydrodynamic importance such as coral reef and seagrass systems, coastal boundaries, and regions characterised by steep bathymetric gradients. The ocean circulation governing equations are then solved within each mesh element to reproduce the three-dimensional hydrodynamics of the GBR region.

During the second step, nitrogen riverine inputs, transport, and natural processes are simulated by coupling the three-dimensional hydrodynamic simulation with an NPZD biogeochemical model. The NPZD model represents the interactions between nutrients (N), phytoplankton (P), zooplankton (Z), and detritus (D), and allows estimation of their respective concentrations in space and time over the region. This step aims to reproduce the main biogeochemical response of coastal waters to extreme riverine nitrogen inputs, with particular attention to the emergence, timing, and spatial dynamics of phytoplankton blooms.

Lastly, the NPZD outputs are used to assess the exposure of key coastal ecosystems to cyclone-driven phytoplankton blooms. The spatial and temporal extent of simulated blooms is compared with the distribution of coral reefs and seagrass meadows across the study area in order to identify which ecosystems were exposed to bloom conditions, when this exposure occurred, and how long it persisted. In addition, these exposure patterns are interpreted in light of existing literature and available observational data describing the evolution of ecosystem health following the TC events.

2.2 Region of interest

This master’s thesis focuses on the GBR region, a vast marine area located along the northeastern coast of Australia, in the South Pacific Ocean. Covering approximately 346,000 km², it extends offshore from the Queensland coast, with a width ranging from 70 to 250 km. The adjacent catchment area is divided into six natural resource management (NRM) regions: Cape York, Wet Tropics, Burdekin, Mackay Whitsunday, Fitzroy, and Burnett Mary regions [GBRMPA, 2024]. The catchment population is more than 1.25 million people, with the main population centres being Townsville and Cairns [QGSO, 2025]. This area is widely known for hosting the world’s largest coral reef system; however, it extends far beyond this structure alone. It comprises a mosaic of interconnected coastal and marine ecosystems, including seagrass beds, mangrove forests, estuaries, coastal wetlands, as well as offshore pelagic environments and continental islands. This high variety of habitats makes it a place on which thousands of species depend [DC-CEEW, 2021].

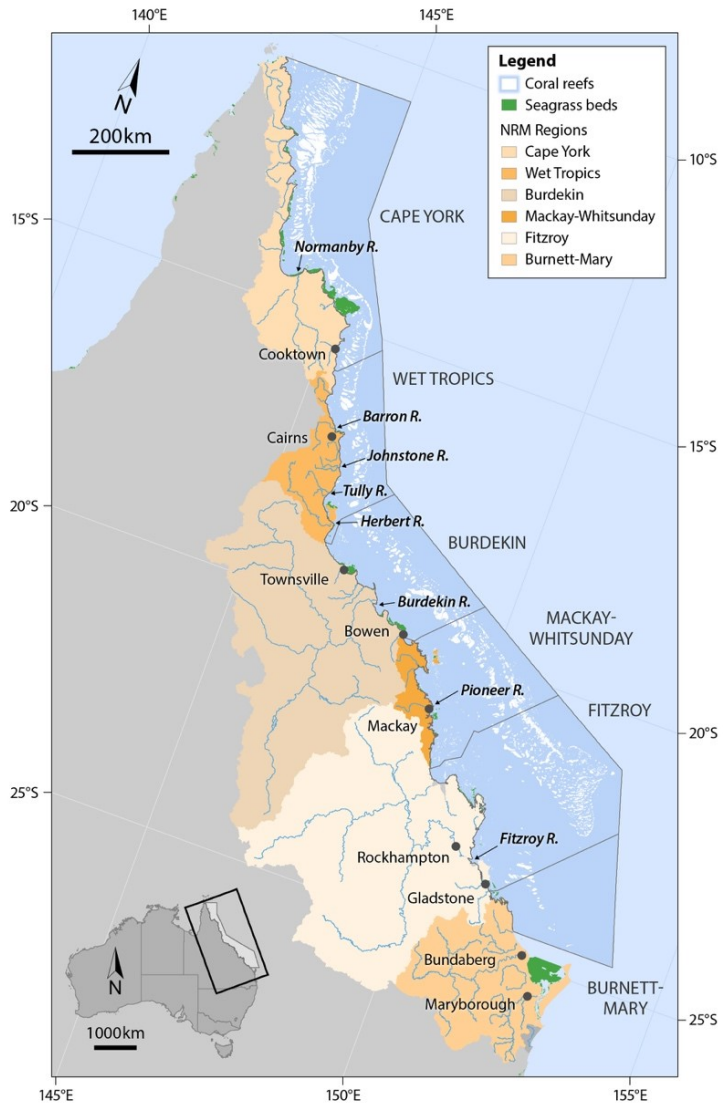


Figure 5: Overview of the GBR region, including the NRM regions and main population centres [Waterhouse et al., 2018].

However, this large ecosystem is increasingly exposed to significant stressors. At the regional scale, anthropogenic activities contribute to the deterioration of water quality, while at the global scale, climate change represents the most significant threat [CSIRO, 2021].

As this thesis specifically investigates the effects of TC-driven nitrogen inputs on coral reef and seagrass ecosystems, particular attention is given to coastal and inner-shelf environments, where riverine nutrient inputs are strongest and where these ecosystems are predominantly located. However, to fully capture the local-scale processes, a comprehensive understanding of the large-scale physical setting of the region is required. The following section therefore provides key background on the regional bathymetry, dominant ocean circulation patterns, and the main river systems discharging into the coastal waters, which together strongly influence the transport and dispersion of nutrients within the GBR system.

2.2.1 Bathymetric profile

The GBR region is characterised by a complex marine morphology composed of several major physiographic structures, including the continental shelf, the continental slope, offshore plateaux, and deep basin systems [Hopley et al., 2007]. Figure 6 illustrates the spatial distribution of these units and highlights the strong bathymetric gradients across the area of interest.

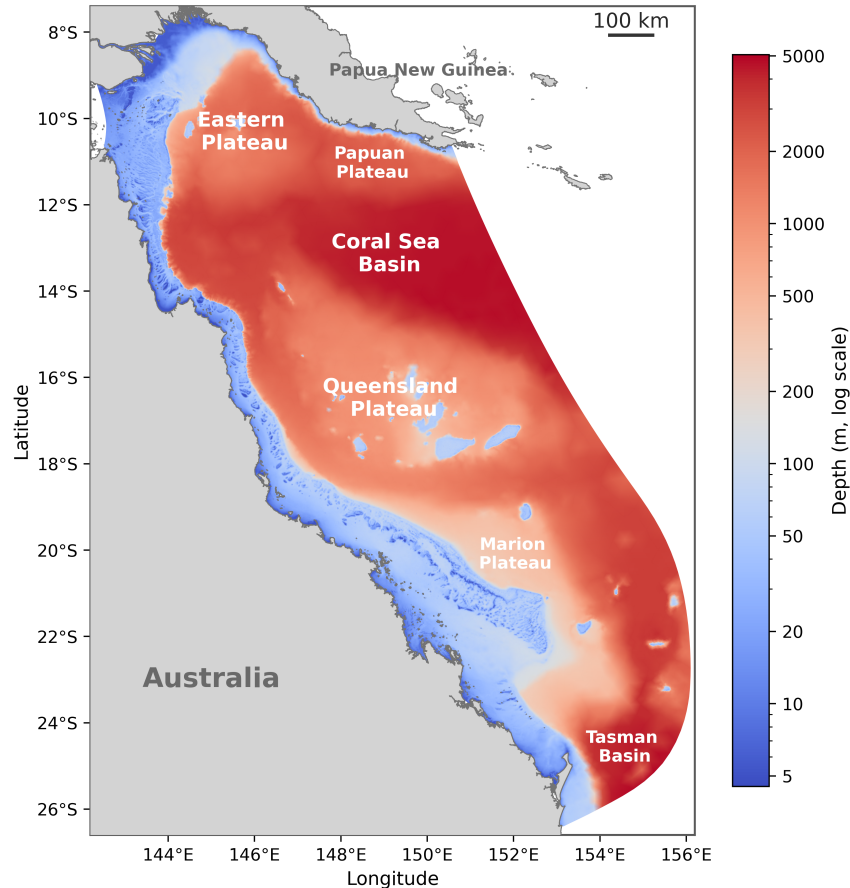


Figure 6: Bathymetry and major physiographic structures of the model's computational domain.

The continental shelf corresponds to the submerged edge of the continent and represents the shallowest part of the system. It is characterised by a gentle depth gradient, with depths generally remaining below 100 m [Beaman, 2008]. Within the GBR region, the continental shelf can be subdivided into three zones: the inner-shelf (from the coast to the 20 m isobath), the mid-shelf (between the 20 and 50 m isobaths), and the outer-shelf (beyond the 50 m isobath) [Hopley et al., 2007]. The length and width of the continental shelf vary along the Queensland coast. In the northern sector, where it is shared between the Australian and Papua New Guinean margins, the shelf is relatively broad. It then narrows to form a constricted section approximately 40-60 km wide up to latitude 16°S. South of this latitude, the shelf progressively widens, reaching a maximum extent of about 270 km. However, further south of latitude 23°S, the continental shelf narrows again [Bridge et al., 2019].

At the seaward edge of the shelf, continental shelf transitions abruptly into the continental slope, which is characterised by a much steeper bathymetric gradient. These slopes contain nu-

merous submarine canyons that play an important role in transporting sediments from the surface toward deeper oceanic environments [Beaman, 2008].

Further offshore, the slope gives way to a complex deep-ocean landscape composed of plateaux and basins with depths ranging from several hundred to several thousand m. In the northern part of the study area, the Eastern and Papuan Plateaux are rapidly replaced by the Coral Sea Basin, the deepest part of the region, reaching depths exceeding 5,000 m. In contrast, the Queensland and Marion Plateaux provide a more gradual transition toward deep oceanic environments in the central and southern sectors. In the southernmost sector, this transition becomes markedly more abrupt, with the continental margin giving way directly to the Tasman Basin [Ceccarelli et al., 2013].

These strong bathymetric contrasts exert a major control on ocean circulation, water mass exchanges, and transport processes throughout the GBR region.

2.2.2 Oceanic circulation

Ocean circulation plays a central role in this master's thesis, as it drives the transport of water masses and regulates the redistribution of heat, salinity, and nutrients across the GBR region. This circulation results from the combined action of wind-driven currents and deeper thermohaline processes associated with temperature and salinity variations, making it inherently variable over time [Delorme & Eddebbbar, 2016]. The resulting seasonal variability of the western Coral Sea and the outer GBR surface circulation is illustrated in Figure 7.

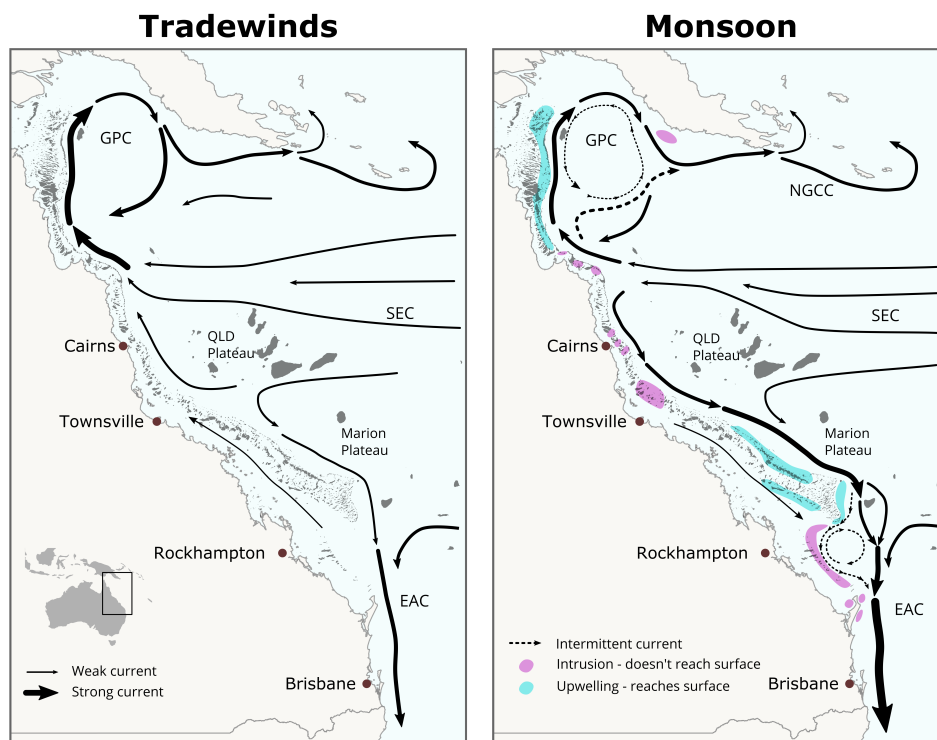


Figure 7: Prevailing surface currents in the GBR and Coral Sea in the Monsoon and Tradewind seasons [C. Steinberg & Lawrey, 2018]

Two dominant circulation regimes characterise the region throughout the year: the trade wind season and the monsoon season. These patterns result in significant variations in both the strength and direction of surface currents. During the trade wind season, surface circulation is relatively organized and homogeneous, whereas the monsoon season is characterised by more complex flow conditions. Moreover, monsoon conditions promote the occurrence of oceanic processes such as upwelling and subsurface intrusions [C. Steinberg & Lawrey, 2018].

More particularly, currents within the GBR lagoon are mainly influenced by two factors: the South Equatorial Current (SEC) and dominant winds [GBRMPA, 2024]. The SEC is a major westward-flowing ocean current that constitutes a major component of the large-scale circulation of the Coral Sea. It enters the region as a series of jets between the Solomon Islands and New Caledonia, interacting with the Queensland Plateau [G. Smith & Spillman, 2019]. Upon reaching the continental shelf of north-eastern Australia, the SEC splits into two main outflows from the GBR region. To the north, the Gulf of Papua Current (GPC) transports approximately 8.2 ± 19.1 Sv toward the Western Pacific Ocean, while generating a partially closed eddy in the northern Coral Sea. To the south, the East Australian Current (EAC) carries a stronger and more stable transport of 14.4 ± 5.8 Sv along the eastern Australian coastline to the Tasman Sea [Schiller et al., 2015]. While the SEC, GPC, and EAC define the large-scale circulation in the region, its surface expression is significantly modulated by seasonal wind patterns. During the austral winter (trade wind season), south-easterly winds consistently exert a north-westward force on the ocean surface, strengthening northward-flowing currents such as the GPC, while opposing and weakening the EAC. Conversely, during the austral summer (monsoon season), surface winds weaken considerably, allowing large-scale oceanic circulation to dominate. Under these conditions, the EAC reaches its seasonal maximum transport [C. Steinberg & Lawrey, 2018].

Wind forcing exerts an even stronger control over circulation on the continental shelf and in coastal waters, where south-easterly trade winds are the primary driver of nearshore currents [GBRMPA, 2024]. Together with the Coriolis effect, these wind regimes primarily govern river plume transport in the GBR region [E. Wolanski, 1994] – a process particularly relevant to this study given that river plumes represent major pathways for nutrients during extreme events such as TC Jasper and Kirrily. In the Southern Hemisphere, the Coriolis force deflects flowing water masses to the left of their initial trajectory, contributing to a systematic northward transport of plume waters at Queensland river mouths [King et al., 2001]. This tendency is reinforced under typical trade wind conditions, where the strong north-westward longshore current on the inner shelf confines the plumes close to the coastline, limiting their cross-shelf dispersion [Wolanski & Ruddick, 1981]. During periods of weakened wind forcing such as the monsoon season, plume waters spread further offshore while generally maintaining a northward trajectory [Devlin, 2005]. The transport of materials within these plumes also depends on their physical properties. Particulate matter generally settles relatively close to river mouths, typically within the first 10 km. In contrast, dissolved nutrients such as DIN can be transported much further, with elevated concentrations detectable over several hundred of km from the coast [Devlin & Brodie, 2005].

2.2.3 Main river systems and study area

Given the importance of river plumes in transporting nutrients across the GBR lagoon, particular attention is paid to the main rivers discharging into the region and contributing to these inputs during the TC events of interest. Thirty-five river systems discharge into the region, collectively exporting an average of 60 million megalitres of freshwater per year. However, discharge volumes vary considerably among catchments [GBRMPA, 2024], and including all watercourses in the hydrodynamic simulation would have introduced unnecessary complexity and limited additional relevance. This thesis therefore focuses on a selected subset of the most influential river systems, which are subsequently used to simulate nitrogen inputs. Rivers were selected based on their hydrological importance within the GBR basin and the magnitude of their response to the TC Jasper and Kirrily. Based on these criteria, four systems were identified as the most relevant: the Daintree, Barron, Herbert, and Burdekin rivers. All are located near the landfall regions of the two TCs (see Figure 8). The corresponding river discharge data and daily DIN loads are described in the forcing sections of the hydrodynamic and NPZD models (Subsections 2.3.3 and 2.4.3, respectively).

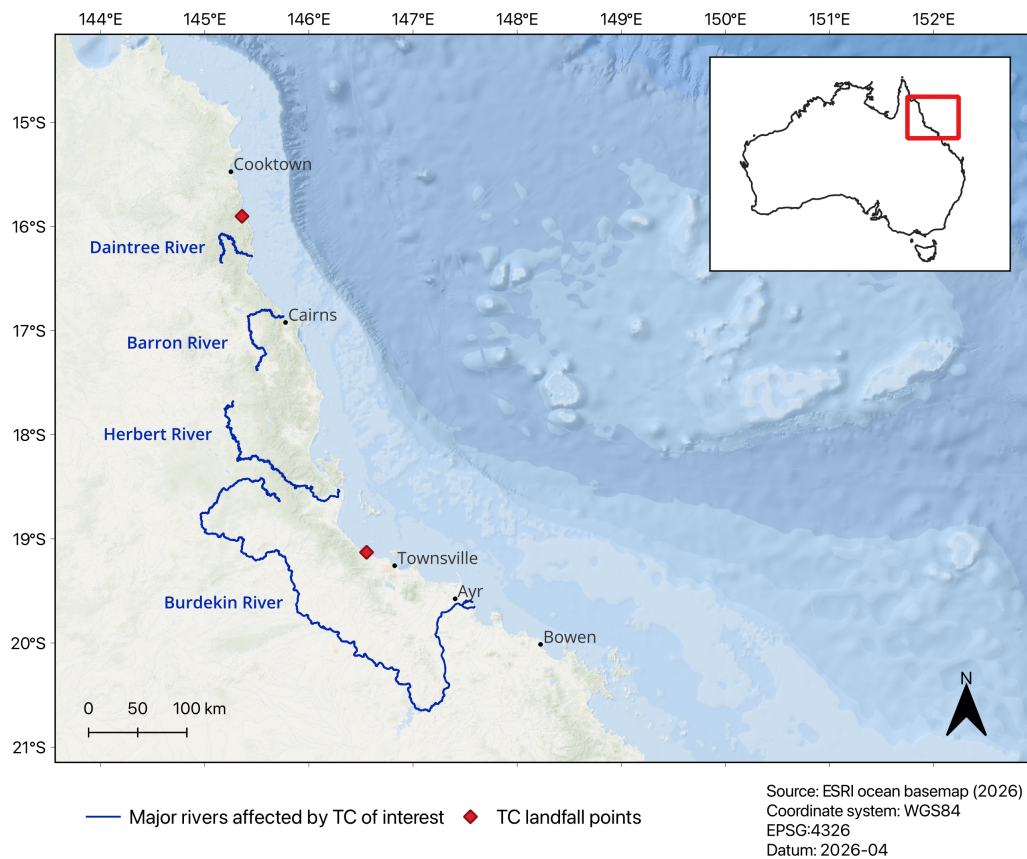


Figure 8: Main river systems contributing to riverine DIN inputs within the study area after TC Jasper and Kirrily.

The four river systems are all located along a relatively restricted section of the Queensland coast (between 16°S and 20°S), making this coastal corridor the region of primary interest of this study. The study area is thus defined to encompass the continental shelf environment, where riverine

nutrient inputs exert the strongest influence, and where coral reef and seagrass ecosystems are most likely to be exposed to TC-driven phytoplankton blooms. Simulation outputs and ecosystem impact assessments are therefore primarily analysed within this spatial framework.

2.3 Hydrodynamic modelling

Ocean currents, as well as key oceanic variables such as temperature, salinity, and sea surface elevation, are simulated by solving the ocean circulation governing equations under the hydrostatic and Boussinesq approximations on an unstructured mesh. This section presents the main concepts underlying the hydrodynamic model used in this study. First, the mathematical formulation of the model is introduced, followed by its numerical implementation within SLIM3D. Then, the forcings and initial conditions applied to the numerical model are detailed, along with the validation methodology. This section is based on the SLIM Wiki, the SLIM-Ocean¹ website, and the work of Delandmeter (2017). It also draws inspiration from the master theses of Scherpereel (2022) and Van Eetvelt (2024).

2.3.1 Mathematical model

Computational domain

The domain on which the three-dimensional hydrostatic equations are applied is denoted as Ω . It is described by three coordinates: x and y represent the horizontal dimensions of a local planar projection of the Earth's surface, while z is the vertical coordinate (see Figure 9). Ω is bounded by three types of boundaries: the fixed lateral boundaries (Γ_l), the bottom boundary (Γ_b), and the moving top surface boundary (Γ_s). Lateral boundaries can either be closed or open, depending on whether they correspond to the coastline or the open ocean. The bottom boundary corresponds to the seafloor, i.e. $z = -h(x, y)$, where $h(x, y)$ is the bathymetry. Finally, the top boundary corresponds to the time-dependent free surface elevation ($z = \eta(t, x, y)$) defined with respect to a reference surface $\Gamma_{s,0}$ ($\eta = 0$) corresponding to a sea at rest. The total water depth is therefore given by $H(t, x, y) = h(x, y) + \eta(t, x, y)$.

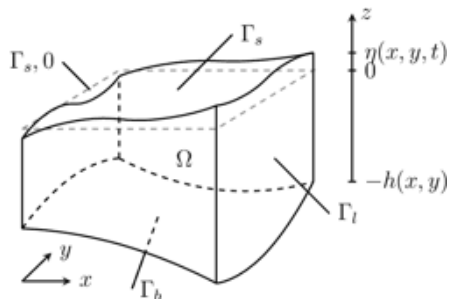


Figure 9: Sketch of the 3D domain Ω defined by its boundaries Γ . The top surface Γ_s is the only moving boundary [Delandmeter, 2017].

¹<https://www.slim-ocean.be/>

Physical assumptions

A full resolution of the governing flow equations would be unnecessarily complex and computationally demanding. For that reason, SLIM3D solves the three-dimensional hydrostatic Boussinesq equations, which are a simplification of the Navier-Stokes equations, where:

- Vertical acceleration is assumed to be negligible (hydrostatic approximation). This assumption states that the gravitational force that is applied to a fluid is compensated by the vertical component of the pressure gradient force it experiences. In other words, the weight of the water column lying above is balanced by the upward force resulting from the pressure difference this water column generates. The Navier-Stokes vertical momentum conservation equation therefore reduces to:

$$\frac{\partial p}{\partial z} = -\rho g, \quad (1)$$

where p is the pressure [Pa], ρ the density of the fluid [kg m^{-3}] defined such that $\rho = \rho_0 + \rho'$, with ρ_0 and ρ' respectively the reference density value of 1000 [kg m^{-3}] and the density variation, and g is the gravitational acceleration [m s^{-2}].

- Density is assumed to be constant, except in the gravity term (Boussinesq approximation). It implies that, since the density variation is much smaller than the reference density, i.e. $\rho - \rho_0 \ll \rho$, it can therefore be neglected in all the terms of the momentum equation, except when associated with the gravitational term g .

Model variables and governing equations

The three-dimensional hydrostatic equations solved by SLIM3D over the domain Ω involve seven hydrodynamic variables (see Table 1). Five are prognostic (u , η , S , T , and p), while two (w and ρ') are diagnostic. Prognostic variables are advanced in time by the model, whereas diagnostic variables are computed from the prognostic fields at each time step. In total, the system consists of five governing equations (momentum, continuity, pressure, and tracers T and S), together with one supplementary equation for the computation of the density deviation.

Table 1: Hydrodynamic model variables

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}

Under assumptions detailed earlier, the momentum, continuity and the sea surface equations (the latter also being called depth-averaged continuity equation) read:

$$\frac{\partial \mathbf{u}}{\partial t} + \nabla_h \cdot (\mathbf{u}\mathbf{u}) + \frac{\partial(w\mathbf{u})}{\partial z} = \nabla_h \cdot (\nu_h \nabla_h \mathbf{u}) + \frac{\partial}{\partial z} \left(\nu \frac{\partial \mathbf{u}}{\partial z} \right) - f \mathbf{e}_z \times \mathbf{u} - \frac{1}{\rho_0} \nabla_h p, \quad (2)$$

$$\frac{\partial w}{\partial z} = -\nabla_h \cdot \mathbf{u}, \quad (3)$$

$$\frac{\partial \eta}{\partial t} = -\nabla_h \cdot \int_{-h}^{\eta} \mathbf{u} dz, \quad (4)$$

where $\nabla_h = \left(\frac{\partial}{\partial x}, \frac{\partial}{\partial y} \right)$ is the horizontal gradient operator, $f = 2\Omega \sin \phi$ is the Coriolis parameter depending on the latitude ϕ and the Earth's rotation speed $\Omega = 7.29 \times 10^{-5} [\text{rad s}^{-1}]$, $\mathbf{e}_z$ is the vertical unit vector pointing upwards, and ν_h and ν are respectively the horizontal and vertical eddy viscosities. ν_h follows the Smagorinsky (1963) closure scheme, which expresses the horizontal eddy viscosity as a function of the local strain rate of the flow:

$$\nu_h = \alpha \Delta^2 \sqrt{2 \left(\frac{\partial u}{\partial x} \right)^2 + 2 \left(\frac{\partial v}{\partial y} \right)^2 + \left(\frac{\partial u}{\partial y} + \frac{\partial v}{\partial x} \right)^2}, \quad (5)$$

where α is a non-dimensional constant and Δ^2 is the horizontal area of the mesh element. The pressure gradient term appearing in the Eq. (2) can be further decomposed into its atmospheric, external and internal components under the hydrostatic assumption:

$$\frac{1}{\rho_0} \nabla_h p = \underbrace{\frac{1}{\rho_0} \nabla_h p_{atm}}_{\text{Atmospheric}} + \underbrace{g \nabla_h \eta}_{\text{External}} + \underbrace{\frac{g}{\rho_0} \nabla_h \int_z^{\eta} \rho' dz'}_{\text{Internal}}, \quad (6)$$

with p_{atm} the atmospheric pressure. Additionally, ρ' is computed from the equation of state described in Jackett et al. (2006), and is a function of temperature T and salinity S . Finally, both tracers are modeled using advection-diffusion equations. The left-hand side of the equations gathers the local time derivative and advective transport, whereas the right-hand side represents subgrid-scale turbulent diffusion.

$$\frac{\partial T}{\partial t} + \nabla_h \cdot (\mathbf{u}T) + \frac{\partial(wT)}{\partial z} = \nabla_h \cdot (\kappa_h \nabla_h T) + \frac{\partial}{\partial z} \left(\kappa \frac{\partial T}{\partial z} \right), \quad (7)$$

$$\frac{\partial S}{\partial t} + \nabla_h \cdot (\mathbf{u}S) + \frac{\partial(wS)}{\partial z} = \nabla_h \cdot (\kappa_h \nabla_h S) + \frac{\partial}{\partial z} \left(\kappa \frac{\partial S}{\partial z} \right), \quad (8)$$

where κ_h and κ are the horizontal and vertical diffusivities. Turbulent closure models are necessary to include turbulent subscale processes that cannot be explicitly resolved by the model. They are accounted for by the turbulent variables ν_h , ν , κ_h , and κ . κ_h is parametrised following Okubo's approach [Okubo, 1971], even though Smagorinsky's scheme could also be applied for κ_h . Lastly, ν and κ are computed using the $\kappa - \epsilon$ turbulent closure model [Rodi, 1987; Kärnä, 2020]. The transport of temperature and salinity is further affected by surface heat and freshwater fluxes, including heat

transfer through turbulent air motion for sensible Q_{sens} and latent Q_{lat} fluxes, radiative forcing for thermal (longwave) Q_{LW} and solar (shortwave) Q_{SW} components, and surface evaporation E and precipitation P . These fluxes are imposed as Neumann boundary conditions to Eq. (7) and (8):

$$\kappa_v \frac{\partial T}{\partial z} = -\frac{(Q_{sens} + Q_{lat} + Q_{LW} + Q_{SW})}{\rho C_p} - Q_{nudg}, \quad (9)$$

$$\kappa_v \frac{\partial S}{\partial z} = -S(P - E) - Q_{nudg}, \quad (10)$$

where C_p is the specific heat capacity of seawater, and $Q_{nudg} = \frac{\Delta z_s}{\tau}(C_0 - C)$ is an additional surface relaxation term implemented as a simple data assimilation method, where C_0 is the observed surface value of the tracer (either T_{obs} or S_{obs}) and C the corresponding simulated value, as a function of a surface mixed layer depth Δz_s and a relaxation period τ . Surface evaporation $E = \frac{-Q_{lat}}{\rho L_e(T)}$ is determined as a function of the latent heat flux and the temperature-dependent latent heat of vaporization $L_e(T)$ defined by Henderson-Sellers (1984). In addition, corrections to the air-sea fluxes based on bulk formulations are implemented to add feedback processes between the evolution of the simulated surface tracers and the atmospheric forcings computed a priori, following the approach of Barnier et al. (1995). These corrections apply to the outgoing heat fluxes and include sensible, latent, and longwave radiation components.

Boundary conditions

As previously mentioned, the three-dimensional domain Ω is delimited by static lateral Γ_l and bottom Γ_b boundaries, and includes a dynamic surface boundary Γ_s . The mesh deformation to track the free surface elevation is achieved through a generic Arbitrary Lagrangian Eulerian (ALE) formulation [Donea et al., 2004]. Impermeability conditions are imposed on Γ_b and the closed lateral boundaries Γ_l , while the evolution of Γ_s is influenced by currents and vertical freshwater fluxes:

$$w - \mathbf{u} \cdot \nabla_h b = 0, \quad \mathbf{x} \in \Gamma_b, \quad (11)$$

$$\mathbf{u} \cdot \mathbf{n}_h = 0, \quad \mathbf{x} \in \Gamma_c, \quad (12)$$

$$w - \frac{\partial \eta}{\partial t} - \mathbf{u} \cdot \nabla_h \eta + s = 0, \quad \mathbf{x} \in \Gamma_s, \quad (13)$$

where $\mathbf{n}_h = (n_x, n_y, 0)$ is the horizontal unit normal vector and $s = (P - E)$ is the water surface source term due to precipitation and evaporation. Given that the upper boundary is located at $z = \eta(x, y, t)$, Eq. (13) implies that Ω is dynamic in the time domain. A full slip condition is used for the tangential velocity on the closed boundaries Γ_{lc} . Additional momentum flux boundary conditions are imposed on the bottom boundary Γ_b and free surface Γ_s to account for the bottom friction τ_b and surface wind τ_s stresses:

$$\nu_v \frac{\partial \mathbf{u}}{\partial z} = \frac{\tau_b}{\rho_0}, \quad \mathbf{x} \in \Gamma_b, \quad (14)$$

$$\nu_v \frac{\partial \mathbf{u}}{\partial z} = \frac{\boldsymbol{\tau}_s}{\rho_0}, \quad \mathbf{x} \in \Gamma_s, \quad (15)$$

where bottom stress is parametrized using a logarithmic profile:

$$\frac{\boldsymbol{\tau}_b}{\rho_0} = C_d \|\mathbf{u}_{bf}\| \mathbf{u}_{bf}, \quad (16)$$

$$C_d = \left[\frac{\kappa}{\ln((z_{bf} - b)/z_0)} \right]^2, \quad (17)$$

with C_d the bottom drag coefficient, κ ($\simeq 0.41$) the von Karman constant, z_{bf} the vertical bottom coordinate, taken here as the middle point of the lowermost element, $\mathbf{u}_{bf}$ the bottom velocity evaluated at z_{bf} , and z_0 the bottom roughness length. The surface wind stress $\boldsymbol{\tau}_s$ is defined as:

$$\boldsymbol{\tau}_s = \rho_{air} C_w \|\mathbf{u}_{10}\| \mathbf{u}_{10}, \quad (18)$$

with ρ_{air} the air density, $\mathbf{u}_{10}$ the wind speed at 10 m above the sea level, and C_w is the wind drag coefficient computed using the Smith and Banke regression: $C_w = 10^{-3} \times (0.63 + 0.066 \|\mathbf{u}_{10}\|)$ [Smith & Banke, 1975; Cushman-Roisin & Beckers, 2011].

Barotropic mode

As a complement to the previous model, an external two-dimensional mode is added to simulate the fast propagating surface gravity waves with depth-averaged equations, increasing computational efficiency. In this context, water transport $\mathbf{Q} = H\mathbf{u}$ is solved with Eq. (19) after the integration of the three-dimensional terms from the momentum equation, represented by S_{3D} in Eq. (20).

$$\frac{\partial \mathbf{Q}}{\partial t} = -gH\nabla_h \eta - \frac{H}{\rho_0} \nabla_h p_a + S_{3D}, \quad (19)$$

with

$$\begin{aligned} S_{3D} = & \nabla_h \cdot \int_b^\eta -\mathbf{u}\mathbf{u} \, dz \\ & + \int_b^\eta \left[\nabla_h \cdot (\nu_h \nabla_h \mathbf{u}) + \frac{\partial}{\partial z} \left(\nu_v \frac{\partial \mathbf{u}}{\partial z} \right) - f \mathbf{e}_z \times \mathbf{u} - \frac{g}{\rho_0} \nabla_h \int_z^\eta \rho' \, d\zeta \right] dz. \end{aligned} \quad (20)$$

2.3.2 Numerical model

Spatial discretization

SLIM3D solves the aforementioned equations on three-dimensional triangular prismatic elements using the Discontinuous Galerkin Finite-Element (DG-FE) scheme and a $P_1^{DG} - P_1^{DG}$ discretization. Unlike continuous Galerkin methods, which enforce solution continuity across element interfaces, the DG methods allow each element to retain its own independent degrees of freedom. As

a result, it is particularly well suited for parallel computing. The $P_1^{DG} - P_1^{DG}$ discretization implies that the model variables are approximated using piecewise linear polynomials within each element.

The use of the FE method requires a mesh for performing hydrodynamic simulations over the domain Ω . First, a two-dimensional unstructured surface mesh is generated by discretizing Γ_s into a set of triangular elements $\mathcal{T}$ with the mesh generator GMSH [Geuzaine & Remacle, 2009] through the seamsh² Python library (Figure 10). Using an unstructured mesh allows the spatial resolution to vary across the study area: higher resolution is allocated in areas of ecological and hydrodynamic importance, while coarser resolution is used offshore. In this work, the mesh resolution is adjusted by considering the distance to the coastline, the proximity to coral reefs and seagrass meadows, as well as bathymetric variations. This results in a mesh of more than 360,000 triangles with minimal and maximal resolutions of ~ 200 m and ~ 50 km.

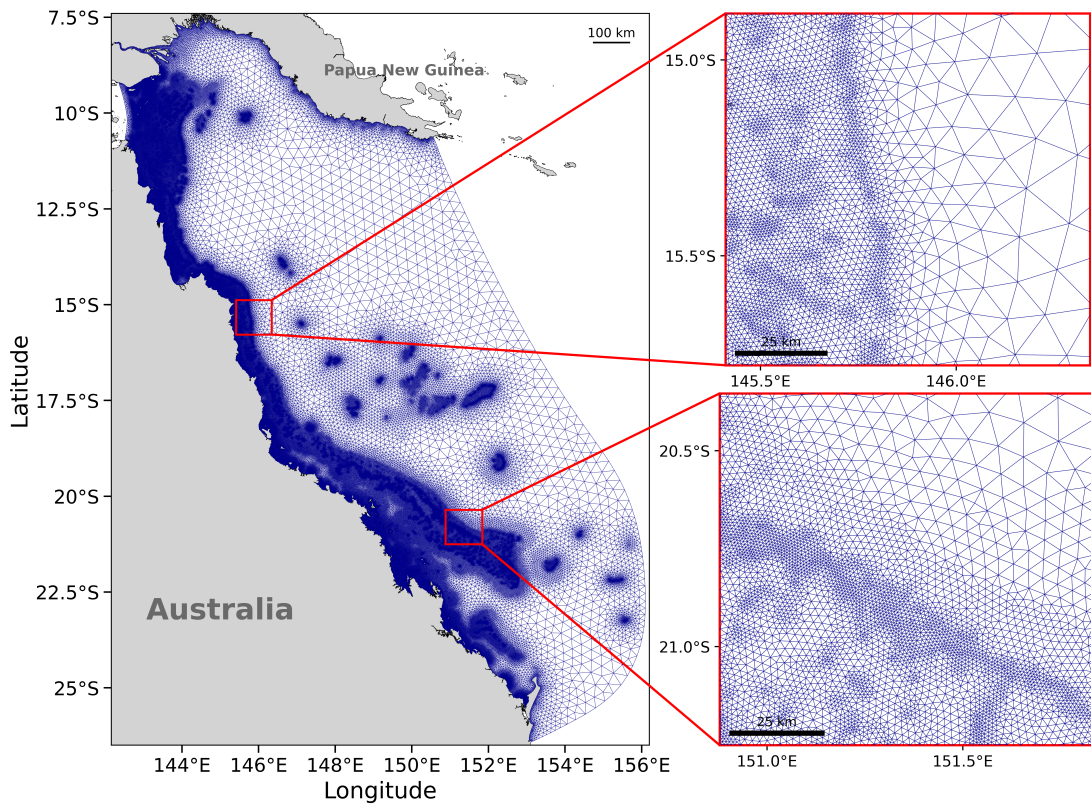


Figure 10: Unstructured two-dimensional mesh of the GBR domain, with close-up views on the shelf break and coral reef areas. The mesh resolution ranges from 200 m to 50 km.

Second, a three-dimensional mesh is generated by vertically extruding the two-dimensional surface towards Γ_b , resulting in a set of prism columns $\mathcal{P}$ (Figure 11). Each column is then subdivided into several prism layers, forming the final three-dimensional discretization of the domain. The ocean state variables are computed at the nodes of the prisms, resulting in six degrees of freedom per element. The prism distribution depends on the local depth and the type of selected coordinate system. Different approaches may be used, including conventional σ -coordinates, stretched S-coordinates, or geopotential z -coordinates at fixed vertical levels. Each of these vertical coordinate systems has its own advantages and weaknesses.

²<https://pypi.org/project/seamsh>

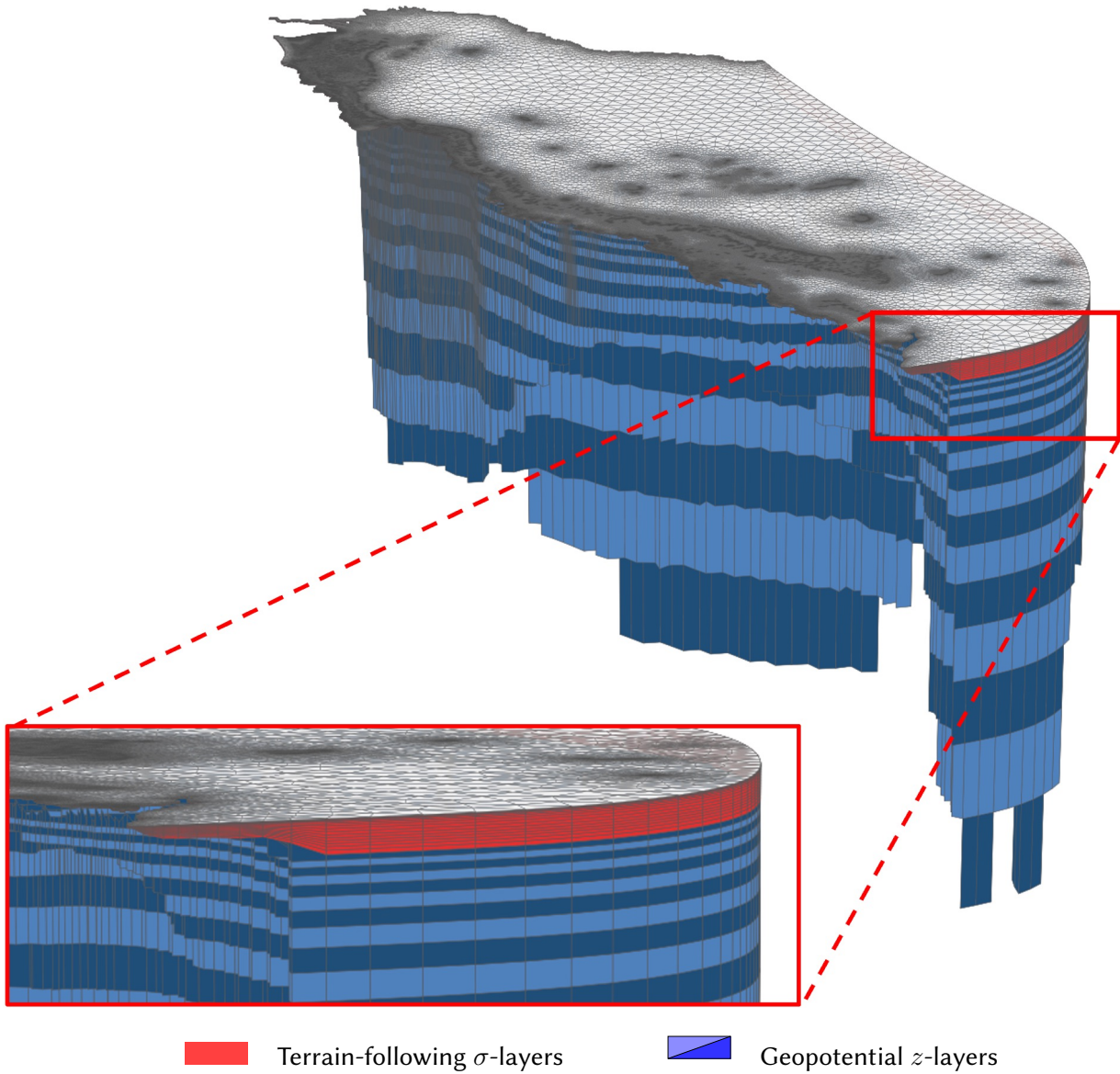


Figure 11: Three-dimensional mesh using a hybrid vertical coordinate system. Terrain-following σ -layers are represented in red, while geopotential z -layers are shown in blue.

Known as terrain-following coordinates, σ -layers follow the bathymetry of the domain, allowing for an accurate representation of the ocean topography. In a traditional σ system, layer thickness is uniformly distributed throughout the water column. Stretching functions can however be applied to generate S -coordinates, in which the vertical layer thickness varies along the water column. These two very similar systems provide a realistic representation of the hydrodynamics inside the domain but can lead to a reduced numerical stability of the model, especially in areas with highly variable bathymetry. For this reason, σ -layers are mainly used in shallow waters and coastal areas, where the topography is relatively smooth.

Z -layers are geopotential layers that traverse the domain at fixed depths, independently of the local bathymetry. As a result, the ocean floor is then represented as a staircase-like structure, and the number of active layers varies across the domain following the local depth. While this system

is computationally robust, the stepped representation of the seafloor induces a poor representation of the near-bottom ocean dynamics. Z -layers are generally avoided in shallow and coastal areas, where bathymetric gradients strongly influence the flow, and are instead mainly used in very steep domains or in the deep ocean.

To combine the hydrodynamic accuracy of terrain-following coordinates with the numerical stability of geopotential layers, a hybrid vertical coordinate system is adopted in this thesis. In this approach, the water column is divided at a prescribed depth threshold: stretched S -coordinates are used in the upper part of the ocean, where an accurate representation of coastal and shelf hydrodynamics is required, while geopotential z -coordinates are applied below, where steep bathymetric gradients would compromise the numerical stability of terrain-following layers. In the present study, stretched S -coordinates are applied within the upper 150 m of the water column, while geopotential z -coordinates are used below. This strategy allows the continental shelf, which is shallow and characterised by relatively smooth bathymetric gradients, to be fully represented by σ -layers. With 15 σ -layers and 20 z -layers, respectively, this configuration results in a total of 35 vertical layers. The resulting vertical discretization is illustrated in Figure 11, where terrain-following layers are shown in red and geopotential layers in blue.

Temporal discretization

Temporal discretization in SLIM3D is based on a mode-splitting strategy that separates fast barotropic (2D external mode) dynamics from slower baroclinic (3D internal mode) processes. The barotropic mode represents depth-averaged gravity waves (uv2D) and the evolution of the free-surface elevation (η), and is integrated using a small time step. In contrast, the baroclinic mode captures the fully three-dimensional internal motions (uv3D, temperature, and salinity) and is advanced with a larger time step. Consequently, several barotropic substeps (ΔT) are performed within each baroclinic time step (Δt). These 2D iterations update the free-surface elevation and depth-averaged velocities, which are then used in the subsequent 3D integration step. The coupling between the two modes is ensured through a split-IMEX Runge-Kutta scheme [Ishimwe et al., 2024]. This mode-splitting approach greatly reduces computational costs compared to applying the small barotropic time step to the full 3D system.

2.3.3 Model forcings and initial conditions

The mathematical and numerical framework described above defines how the ocean state evolves in space and time. However, realistic simulations require the prescription of initial conditions, which define the initial state of the system, and external forcings, which constrain its evolution at the boundaries and at the surface throughout the simulation period. These forcings include both static and time-dependent components.

The bathymetry, which defines the geometry of the domain, is the only static forcing. In

the present work, it is derived from the AusBathyTopo Great Barrier Reef 30 m dataset ³, a high-resolution Digital Elevation Model (DEM) produced by Geoscience Australia in 2020. It was compiled from ship-based multibeam and single-beam echosounder surveys, airborne LiDAR bathymetry, and satellite-derived bathymetry data, resulting in a grid resolution of approximately 30 m (0.0003 arc-degrees). The dataset covers the entire GBR domain through four overlapping tiles, with an additional tile extending coverage to the Torres Strait region.

In contrast to the bathymetry, all remaining forcings are time-dependent and are imposed either at the open boundaries or at the ocean surface. Tidal forcing, including tidal elevation and velocity, as well as sea surface elevation, ocean currents, surface winds, precipitation, temperature, and salinity, are obtained from BARRA2 (Bureau of Meteorology Atmospheric high-resolution Regional Reanalysis for Australia). This Australian regional atmospheric reanalysis provides long-term and spatially complete records of the atmosphere and the surface at high resolution (4.4 km) [Bureau of Meteorology, 2026]. In addition to prescribing the open-boundary conditions, BARRA2 provides the initial temperature and salinity conditions. Finally, the heat fluxes, more specifically surface solar and turbulent heat flux components, are provided by ERA5⁴ (ECMWF Reanalysis 5th Generation). Produced by the European Centre for Medium-Range Weather Forecasts (ECMWF), this dataset has a horizontal resolution of nine km, and an hourly temporal resolution.

Finally, riverine inputs are incorporated into the simulation through prescribed discharge conditions imposed at river boundaries. For the Barron, Burdekin, and Herbert rivers, in situ discharge observations were obtained from the Queensland Water Monitoring Information Portal⁵. For each river, the most downstream monitoring station with available discharge data was selected, thereby providing the closest available estimate of freshwater discharge to the river mouth. However, no discharge measurements were available for the Daintree River. This limitation was addressed using discharge estimates from the Global Flood Awareness System (GloFAS), which provides globally consistent gridded hydrological time series at daily resolution. The reliability of the GloFAS dataset was assessed by graphically comparing in situ freshwater discharge observations with GloFAS-derived estimates for the Barron, Burdekin, and Herbert rivers (see Figure 12).

Although GloFAS occasionally overestimates or underestimates discharge magnitudes, it successfully reproduces the temporal variability and overall hydrological dynamics of the three river systems. In particular, both datasets consistently capture the timing and occurrence of the major discharge peaks associated with TC events, indicating that GloFAS provides a realistic representation of the dominant hydrological variability. Despite remaining discrepancies in peak magnitudes, the overall agreement is considered sufficient for event-scale applications. Consequently, GloFAS is considered an appropriate proxy for the Daintree River within the framework of this study.

³<https://ecat.ga.gov.au>

⁴<https://cds.climate.copernicus.eu>

⁵<https://water-monitoring.information.qld.gov.au>

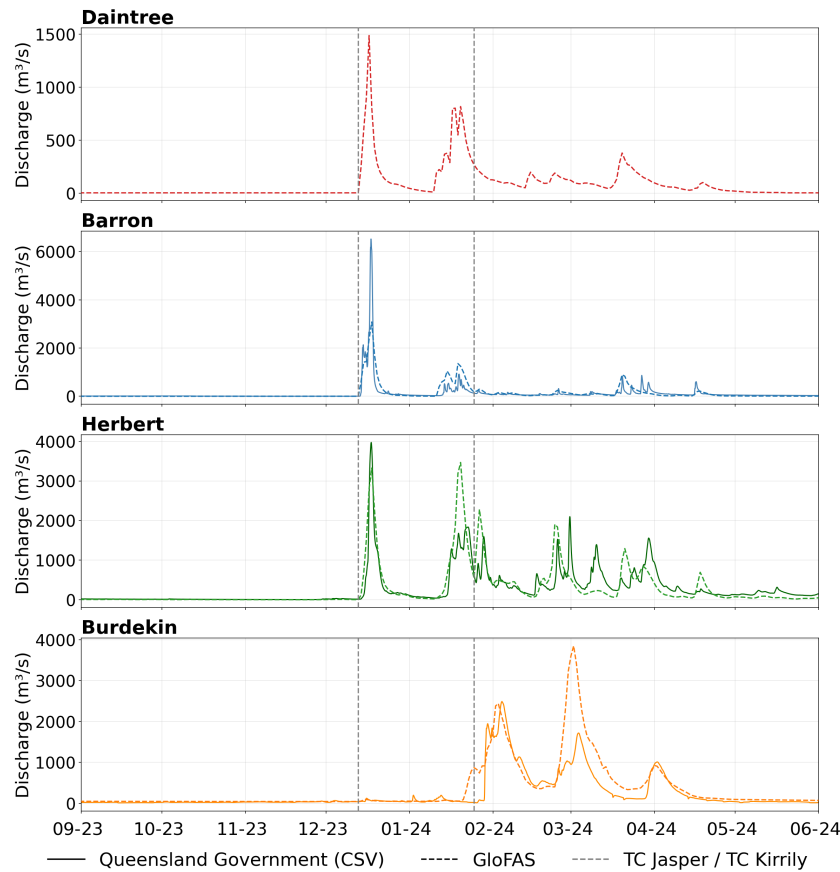


Figure 12: Observed freshwater discharge from the most downstream Queensland Government monitoring stations, where available, and GloFAS-derived estimates for the four river systems.

Under the influence of the prescribed forcings, the hydrodynamic model progressively evolves from its initial conditions towards a dynamically consistent state, during a phase commonly referred to as the spin-up period. The duration of this adjustment phase depends on the characteristic timescales of the dominant circulation processes and the spatial extent of the domain. In the present study, the simulation is initialized on June 1st, 2023, and the first three months are excluded from the analysis, ensuring that the model has reached a dynamically stable state before the TC events of interest occur in December 2023 and January 2024.

2.3.4 Model validation

Model validation constitutes the final step of the hydrodynamic analysis, and consists of comparing SLIM simulation outputs with in situ observations collected across the study domain. This comparison assesses the ability of the model to reproduce the observed ocean dynamics before its outputs are used to drive the NPZD biogeochemical model.

The present validation relies on two complementary observational datasets. A network of fixed mooring stations from the Mooring Network Facility⁶ of the Integrated Marine Observing

⁶<https://catalogue-imos.aodn.org.au>

System (IMOS) provides continuous measurements of current velocity, temperature, and salinity across Australian coastal waters. These measurements are complemented by free-drifting profiling floats from the Argo program⁷, which deliver vertical profiles of temperature and salinity throughout the water column. During the period of interest (September 2023–June 2024), a total of seven IMOS mooring stations and 21 Argo floats – sampling at 201 locations across the study domain – were operational, as illustrated in Figure 13.

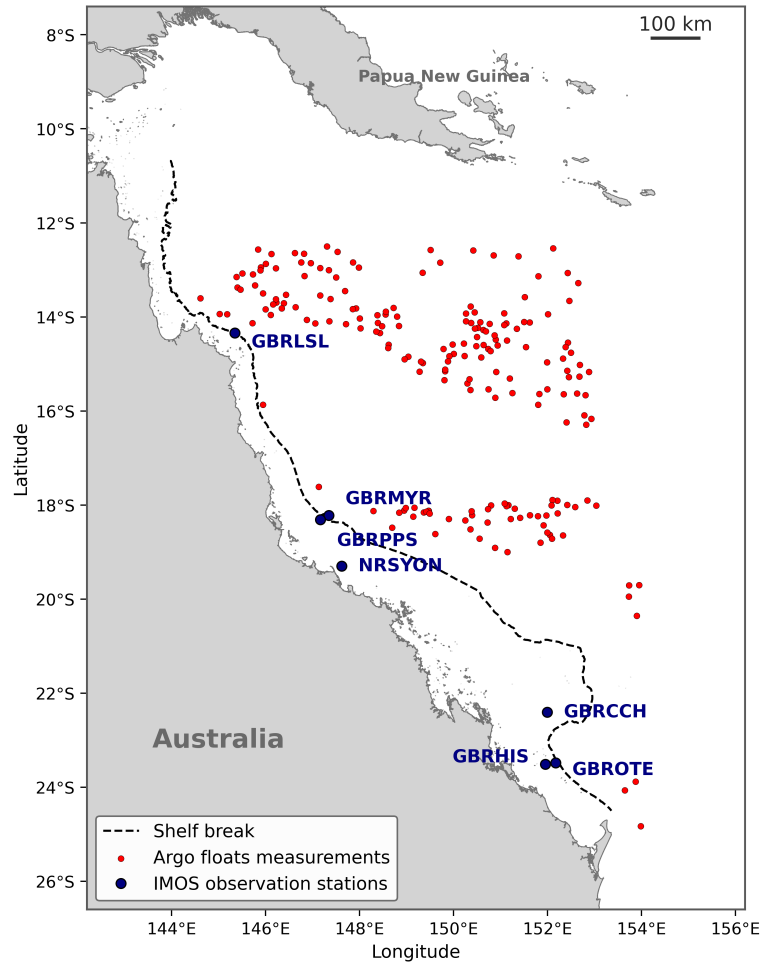


Figure 13: Location of IMOS mooring stations and Argo float observations.

Each IMOS mooring station is equipped with Acoustic Doppler Current Profilers (ADCPs), which measure the zonal, meridional, and vertical components of current velocity throughout the water column. Additional oceanographic sensors provide pressure, temperature, and salinity measurements at fixed instrument depths. These variables are recorded at a temporal resolution of 10 minutes, whereas the SLIM model outputs are available at an hourly resolution. To ensure a consistent and rigorous comparison between datasets, the observational data are linearly interpolated onto the model time grid. Moreover, all variables are accompanied by a quality flag indicating the reliability of the given data. It ranges from 1 to 4, where 1 denotes good-quality data and 4 denotes bad data. In this work, only data associated with a quality flag of 2 or lower were retained. Additional manual filtering was applied to remove outliers associated with instrument deployment and recovery periods.

⁷<https://data.marine.copernicus.eu/>

Argo floats provide complementary observations of the vertical structure of the ocean. These autonomous profiling instruments drift with ocean currents and periodically cycle between the surface and intermediate depths while collecting data. Unlike IMOS moorings, which are constrained to temperature and salinity measurements at fixed instrument depths, Argo floats deliver full-depth profiles of these variables, down to 2000 m. This allows for a more comprehensive validation of the model's thermohaline structure across the water column.

Model performance is assessed through a combination of graphical and statistical analyses. These comparisons focus on sea level anomaly (SLA), horizontal currents, and vertical temperature and salinity profiles. SLA is calculated by subtracting the mean sea level over the validation period from each sea surface elevation time series. Current validation combines complementary analyses of current speed amplitude ($\sqrt{u^2 + v^2}$), direction, and speed distribution. The temporal and vertical variability of current speed amplitude is first assessed throughout the water column using Hovmöller diagrams. Quantitative comparisons are additionally performed at up to three representative depths for each station: the station depth (z_{station}), the mid-depth (z_{half}), and the shallowest depth associated with sufficient quality-controlled measurements (z_{min}). The corresponding depths at each station are summarised in Table 2. Current direction roses and current speed histograms, generated over the entire simulation period at the same representative depths, further characterise the directional and speed distributions of the simulated and observed currents. Vertical temperature and salinity profiles are compared with the corresponding Argo observations to assess the simulated thermohaline structure. Finally, graphical comparisons are complemented by quantitative statistical metrics, including the Root Mean Square Error ($\text{RMSE} = \sqrt{\frac{1}{N} \sum_{i=1}^N (X_{\text{sim},i} - X_{\text{obs},i})^2}$) and the Normalised Root Mean Square Error ($\text{NRMSE} = \frac{\text{RMSE}}{X_{\text{obs,max}} - X_{\text{obs,min}}} \times 100$), expressed as a percentage.

Table 2: Representative depths used for current validation at the seven IMOS stations.

Depth	NRSYON	GBRCCH	GBRHIS	GBROTE	GBRPPS	GBRMYR	GBRLSL
z_{min} (m)	-7.0	-30.0	-10.0	-25.0	-10.0	-40.0	-65.0
z_{half} (m)	-14.5	-42.5	-23.2	-29.5	-35.5	-104.5	-177.2
z_{station} (m)	-28.9	-85.0	-46.4	-59.1	-70.9	-209.0	-354.4

2.4 The NPZD biogeochemical model

This section presents the NPZD biogeochemical model implemented in the present study. First, the governing equations are introduced, with emphasis on the dominant processes driving the evolution of the system. Then, the initial conditions are described, together with the simplifying assumptions required to compensate for the limited availability of observational data. The external forcings are subsequently presented, with particular attention given to riverine dissolved inorganic nitrogen (DIN) inputs. Finally, the validation methodology used to assess model performance is detailed.

The present implementation is a highly simplified biogeochemical model, designed to provide a first-order estimate of the biogeochemical response associated with TCs Jasper and Kirrily. Accordingly, the model is restricted to four aggregated compartments: nutrients (N), phytoplankton

(P), zooplankton (Z), and detritus (D). Their conceptual organization and the nitrogen fluxes linking them are illustrated in Figure 14. Nitrogen is the only nutrient considered within the nutrient pool, as it is generally considered to be the primary limiting element for phytoplankton growth in the GBR waters (Section 1.3). In particular, the model focuses on DIN, comprising nitrite (NO_2^-), nitrate (NO_3^-), and ammonium (NH_4^+), which constitutes the main bioavailable forms of nitrogen assimilated by phytoplankton. Furthermore, no distinction is made among phytoplankton and zooplankton functional groups, which are respectively represented as single aggregated compartments characterised by uniform biological parameters.

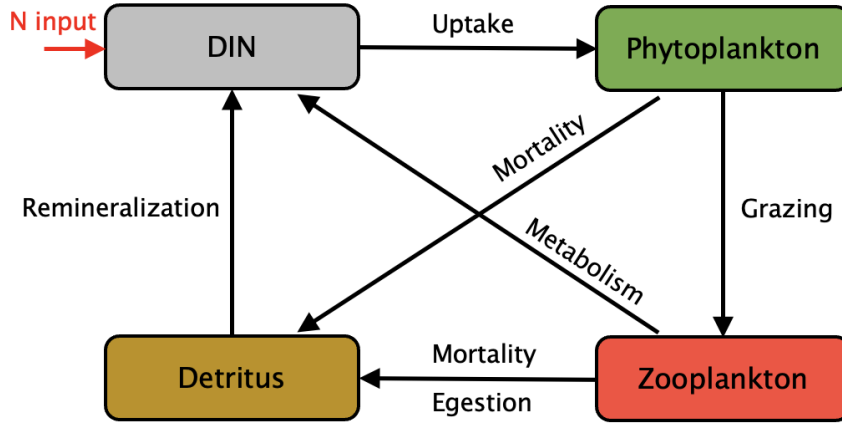


Figure 14: Conceptual representation of the NPZD model structure and nitrogen fluxes among dissolved inorganic nitrogen, phytoplankton, zooplankton, and detritus compartments.

2.4.1 Governing equations

The NPZD model is described by a system of four coupled differential equations representing the temporal evolution of nitrogen concentrations within the state variables (N , P , Z , and D). The variables and parameters involved in these equations are summarised in Table 3 at the end of the section. The present implementation closely follows the NPZD formulations proposed by Tian et al. (2015), whose parameter set constitutes the default reference for this model. This reference was refined using more locally relevant parameter values whenever available and compatible, particularly those from Zhang et al. (2016) and Bureau of Meteorology (2020). At this stage, the equations describe only the local biogeochemical sources and sinks of nitrogen within each grid cell, independently of any spatial transport process. The following equations are solved at each grid cell and model time step.

Nutrients

$$\frac{dN}{dt} = -U_P + \lambda G_Z + \varepsilon f(T)D \quad (21)$$

Phytoplankton

$$\frac{dP}{dt} = U_P - G_Z - m_p P^2 \quad (22)$$

Zooplankton

$$\frac{dZ}{dt} = \gamma G_Z - m_z Z^2 \quad (23)$$

Detritus

$$\frac{dD}{dt} = (1 - \gamma - \lambda)G_Z + m_p P^2 + m_z Z^2 - \varepsilon f(T)D \quad (24)$$

with the phytoplankton uptake term U_P and the zooplankton grazing term G_Z defined as:

$$U_P = \mu_{\max} f(T) f(N) f(I) P \quad (25) \quad G_Z = g_{\max} f(T) Z \frac{P^2}{P^2 + K_P^2} \quad (26)$$

Eq. (21) describes the evolution of the DIN pool (N) as driven by three main processes. The only loss term corresponds to the nitrogen uptake by phytoplankton during primary production, represented by U_P (Eq. (25)). This uptake scales linearly with phytoplankton concentration and with the maximum phytoplankton growth rate $\mu_{\max}$, and is regulated by nutrient availability, light availability, and temperature through the functions $f(N)$, $f(I)$, and $f(T)$. In contrast, the DIN pool is replenished by zooplankton active respiration and detritus remineralization. The former is represented by the fraction λ of the grazing flux G_Z , while the latter regenerates DIN from the detrital pool through a first-order, temperature-dependent process with a rate coefficient $\varepsilon f(T)$.

Eq. (22) describes the evolution of the phytoplankton compartment (P), which is controlled by growth, grazing, and mortality. Phytoplankton growth is represented by the uptake term U_P , whereas grazing losses are represented by G_Z (Eq. (26)). This grazing term scales with zooplankton biomass Z , is modulated by temperature through $f(T)$, and follows a quadratic saturating response to phytoplankton biomass governed by the half-saturation constant K_P . As a result, grazing increases with phytoplankton concentration but progressively approaches an upper limit when P becomes abundant. In addition to grazing, phytoplankton biomass is reduced by a quadratic mortality term, which transfers these losses to the detritus pool.

Eq. (23) characterises the dynamics of the zooplankton compartment (Z). Zooplankton growth results from the assimilation of grazed phytoplankton, represented by the flux γG_Z , where γ is the zooplankton growth efficiency. The only loss term is quadratic mortality, which transfers zooplankton biomass to the detritus pool.

Finally, Eq. (24) describes the evolution of the detritus compartment (D). Detritus is produced by the fraction $(1 - \gamma - \lambda)$ of the grazing flux transferred through egestion and non-assimilated feeding losses, together with phytoplankton and zooplankton mortality. It is removed through temperature-dependent remineralization, represented by $\varepsilon f(T)D$, which converts detrital organic nitrogen back into DIN and thereby recycles nitrogen within the NPZD system.

The effect of temperature on phytoplankton growth, remineralization, and grazing by zooplankton is described by the exponential function $f(T)$. For simplicity and due to a lack of infor-

mation specific to each process, the same temperature parameters are used, resulting in identical temperature responses for all three processes. This function is expressed as:

$$f(T) = e^{-a|T-T_{\text{opt}}|} \quad (27)$$

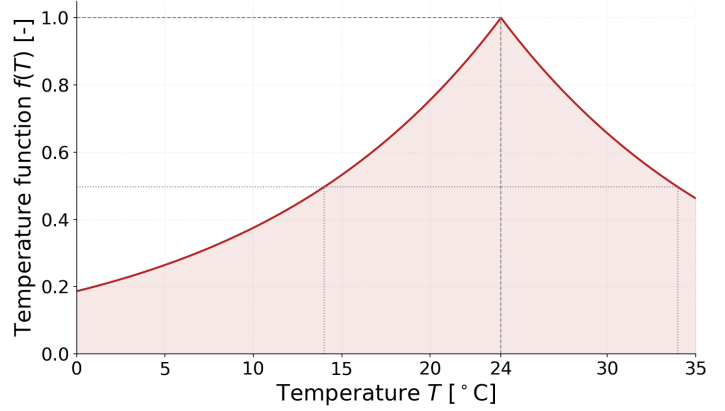


Figure 15: Temperature-dependent biological function $f(T)$.

where a is the temperature coefficient and T_{opt} is the optimal temperature for biological activity, fixed at $0.07 \text{ } ^\circ\text{C}^{-1}$ [Tian et al., 2015] and $24 \text{ } ^\circ\text{C}$ [Chen, 2015], respectively. More specifically, T_{opt} is approximated by the optimal growth temperature of marine phytoplankton at a representative latitude of the GBR region (approximately 17.5°S). This value was estimated graphically from the regression curve established for a diatom of $500 \text{ } \mu\text{m}^3$ cell volume. The function reaches its maximum at T_{opt} and decreases exponentially as temperature deviates from this optimum. With the selected parameterization, the biological response is approximately halved for a $10 \text{ } ^\circ\text{C}$ departure from T_{opt} . Unlike the formulation of Tian et al. (2015), temperature dependence is not applied to the mortality terms, which are instead treated as constant rates, consistent with the simplified nature of the present model.

In addition to temperature, DIN concentration exerts a fundamental control on phytoplankton growth by directly constraining the biomass production. In the model, this effect is represented by the nutrient-dependent phytoplankton growth function, $f(N)$. This function takes the form of a Michaelis–Menten-type formulation, and is expressed as:

$$f(N) = \frac{N}{N + K_N} \quad (28)$$

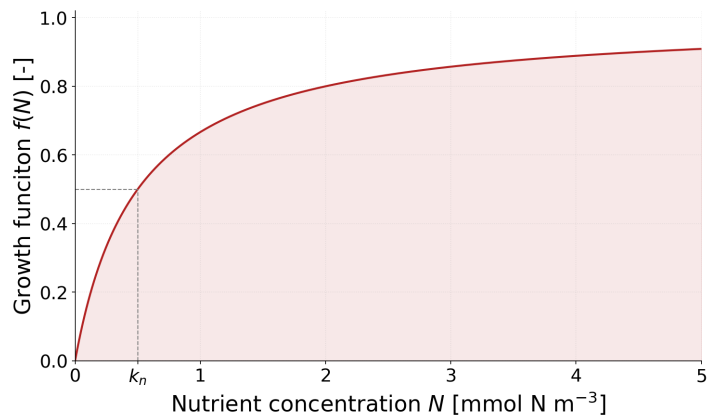


Figure 16: Nutrient-dependent phytoplankton growth function $f(N)$.

where K_N is the half-saturation constant for nutrient uptake, fixed at $0.5 \text{ mmol N m}^{-3}$ [Tian et al., 2015]. This formulation describes a transition from nutrient-limited to nutrient-saturated conditions. At low nutrient concentrations, phytoplankton growth is strongly constrained by nutrient availability. As DIN concentration increases, this limitation progressively weakens and $f(N)$ asymptotically approaches unity.

The uptake of nitrogen by phytoplankton occurs as part of the photosynthetic growth process, which is strongly controlled by photosynthetically active radiation (PAR), denoted $I \text{ [W m}^{-2}\text{]}$. To account for this limitation, a PAR-dependent phytoplankton growth function $f(I)$ is introduced into the growth term. This function follows the formulation of Platt et al. (1980) and is defined as:

$$f(I) = (1 - e^{-\alpha I / \mu_{\max}}) e^{-\beta I / \mu_{\max}} \quad (29)$$

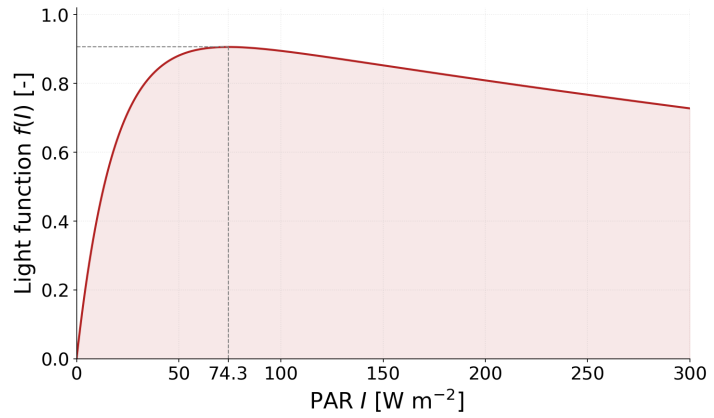


Figure 17: Light limitation function $f(I)$ as a function of photosynthetically active radiation (PAR).

where $\mu_{\max}$ is the maximum phytoplankton growth rate ($3.06 \times 10^{-5} \text{ s}^{-1}$, [Zhang et al., 2016]), α is the light-photosynthesis slope, and β is the light inhibition coefficient, fixed at 1.62×10^{-6} and $3.24 \times 10^{-8} \text{ m}^2 \text{ s}^{-1} \text{ W}^{-1}$, respectively [Tian et al., 2015]. The first exponential term represents the increase in phytoplankton growth under low-light conditions, while the second exponential term accounts for photoinhibition under excessive irradiance. As a result, $f(I)$ increases rapidly at low PAR intensity, reaches an optimum around 74.3 W m^{-2} , and then decreases under excessive PAR. This formulation therefore enables the model to represent both light limitation in deeper waters and photoinhibition under high surface irradiance.

The PAR I is attenuated with depth by water, phytoplankton, and detritus. This attenuation is represented by an exponential formulation, with z denoting depth measured positively downward from the sea surface:

where the parameters $k_W = 0.065 \text{ m}^{-1}$, $k_P = 0.0095 \text{ m}^2 (\text{mmol N})^{-1}$ [Zhang et al., 2016], and $k_D = 0.01 \text{ m}^2 (\text{mmol N})^{-1}$ [Tian et al., 2015] represent light attenuation by water, phytoplankton, and detritus, respectively. I_0 is the surface PAR, fixed here at a representative mean value of 97.1 W m^{-2} . This value was derived from the average daily global solar exposure along the Queensland coast, approximately $19.5 \text{ MJ m}^{-2} \text{ d}^{-1}$ [Bureau of Meteorology, 2020], converted to an equivalent mean irradiance in W m^{-2} . A PAR-to-total-irradiance ratio of 0.43 was then applied to estimate the PAR available at the surface [Fasham et al., 1990]. For the graphical representation, phytoplankton and

$$I(z) = I_0 \exp \left(-k_W z - k_P \int_0^z P dz - k_D \int_0^z D dz \right) \quad (30)$$

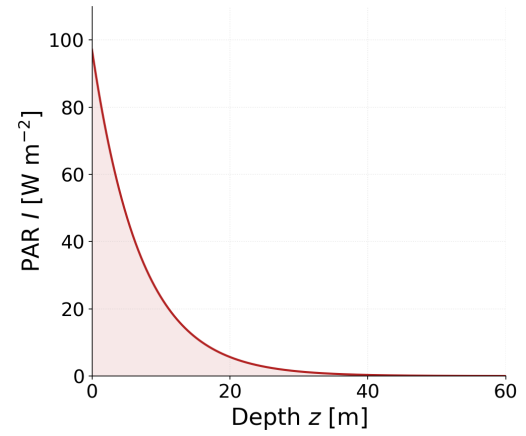


Figure 18: Vertical attenuation of PAR with depth, assuming constant $P = 1.0$ and $D = 0.2 \text{ mmol N m}^{-3}$.

detritus concentrations are assumed to be vertically constant, with $P = 1.0 \text{ mmol N m}^{-3}$ and $D = 0.2 \text{ mmol N m}^{-3}$. The resulting vertical light gradient strongly constrains phytoplankton growth with depth, thereby restricting primary production predominantly to the upper layers of the water column.

Table 3: NPZD model variables and parameters. Source letters indicate parameter sources.

Symbol	Definition	Value	Unit	Source
N	Nitrogen	Variable	mmol N m^{-3}	–
P	Phytoplankton	Variable	mmol N m^{-3}	–
Z	Zooplankton	Variable	mmol N m^{-3}	–
D	Detritus	Variable	mmol N m^{-3}	–
a	Temperature coefficient	0.07	$^{\circ}\text{C}^{-1}$	a
$g_{\max}$	Maximum grazing rate	0.81×10^{-5}	s^{-1}	b
I_0	Surface PAR	97.1	W m^{-2}	c, d
k_P	Light attenuation by phytoplankton	0.0095	$\text{m}^2 (\text{mmol N})^{-1}$	b
k_D	Light attenuation by detritus	0.01	$\text{m}^2 (\text{mmol N})^{-1}$	a
k_W	Light attenuation by pure water	0.065	m^{-1}	b
K_N	Half-saturation constant for nitrogen uptake	0.5	mmol N m^{-3}	a
K_P	Half-saturation constant for grazing	0.25	mmol N m^{-3}	a
m_p	Phytoplankton mortality	9.26×10^{-7}	$\text{s}^{-1} (\text{mmol N m}^{-3})^{-1}$	a
m_z	Zooplankton mortality	6.94×10^{-7}	$\text{s}^{-1} (\text{mmol N m}^{-3})^{-1}$	a
T_{opt}	Optimal temperature	24	$^{\circ}\text{C}$	e
$\mu_{\max}$	Phytoplankton maximum growth rate	3.06×10^{-5}	s^{-1}	b
α	Light–photosynthesis slope	1.62×10^{-6}	$\text{m}^2 \text{s}^{-1} \text{W}^{-1}$	a
β	Light inhibition coefficient	3.24×10^{-8}	$\text{m}^2 \text{s}^{-1} \text{W}^{-1}$	a
ε	Remineralization rate at 0°C	1.736×10^{-7}	s^{-1}	a
λ	Active respiration fraction	0.3	–	a
γ	Zooplankton growth efficiency	0.4	–	a

a = Tian et al. (2015); b = Zhang et al. (2016); c = Bureau of Meteorology (2020); d = Fasham et al. (1990); e = Chen (2015).

2.4.2 Initial conditions

To ensure that the coupled hydrodynamic-biogeochemical system reaches a dynamically consistent state prior to the TC events of interest, the simulation is initialized on June 1st, 2023. Initial conditions must therefore be prescribed for each state variable of the NPZD model at this date. Given the limited availability of observational data across the domain, several simplifying assumptions are required, which are detailed below.

Initial conditions are assumed to be horizontally homogeneous across the entire computational domain. Consequently, spatial variations in the different biogeochemical pools associated with coastal proximity, bathymetry, or local environmental conditions are neglected. Although this assumption limits the physical realism of the initial conditions, it constitutes a necessary compromise to enable a computationally manageable model initialization. Where sufficient data are available, vertical heterogeneity is nonetheless incorporated into the initialization. Profiles of DIN and phytoplankton were derived from outputs of the eReefs biogeochemical model (GBR4 BGC q3b)⁸, sampled at 15 locations distributed along the GBR region at distances ranging from a few tens to several hundred kilometers from the coastline. Since the available eReefs simulations only extend up to 2019, a multi-year averaging approach was applied over the 2011–2018 period to reconstruct representative profiles for June 2023.

Estimating phytoplankton concentrations required an additional assumption, as the only available proxy for phytoplankton biomass is chlorophyll-a concentration (mg chl-a m⁻³). Chlorophyll-a was therefore assumed to be exclusively associated with phytoplankton biomass, neglecting contributions from other autotrophic compartments such as *Trichodesmium* and microphytobenthos. Conversion to phytoplankton nitrogen biomass was carried out in two steps. First, a constant carbon-to-chlorophyll mass ratio of 50:1 was assumed, following Kämpf (2026), and allowing conversion of chlorophyll-a into phytoplankton carbon biomass. Second, a Redfield stoichiometric molar ratio (C:N = 106:16) was used to convert this phytoplankton carbon biomass into nitrogen biomass. This approach results in an approximate conversion factor of 0.63 mmol N (mg chl-a)⁻¹.

$$\text{mg chl-a} \xrightarrow[50 \text{ mg C mg chl-a}^{-1}]{\text{C:Chl-a}} \text{mg C} \xrightarrow[M_C=12 \text{ g mol}^{-1}]{} \text{mmol C} \xrightarrow[N=C \times \frac{16}{106}]{\text{Redfield}} \text{mmol N}$$

The resulting profiles are shown in Figure 19. Prior to their implementation into the NPZD model, they were fitted with analytical functions, ensuring a continuous and smoothly varying representation of concentrations throughout the water column while enabling their extrapolation to greater depths.

No suitable observational data were available for the zooplankton and detritus compartments. However, since zooplankton growth in the model only depends on phytoplankton availability, the initial zooplankton vertical profile was derived from the phytoplankton profile using a constant scaling factor of 0.5. The detritus compartment was initialized at zero concentration throughout the

⁸<https://extraction.eereefs.aims.gov.au>

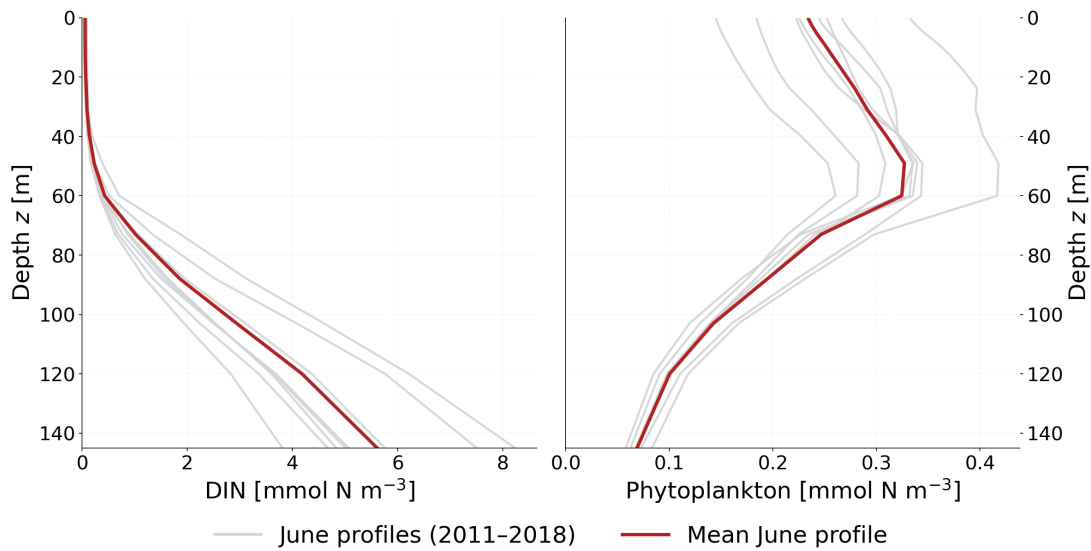


Figure 19: Mean June vertical profiles of DIN and phytoplankton

domain, a simplification expected to have only a limited influence on the simulations as detritus is rapidly produced once phytoplankton and zooplankton mortality begins.

2.4.3 Model forcings

The NPZD model is coupled to the hydrodynamic model through physical transport processes and is forced by prescribed concentrations at the open-ocean boundaries and external riverine inputs of DIN.

Within the ocean domain, the NPZD state variables are transported by the hydrodynamic model through advection and diffusion. Advection redistributes the different compartments according to the simulated ocean circulation, while diffusion represents turbulent mixing acting both horizontally and vertically throughout the water column. In addition to these transport mechanisms, constant downward sinking velocities of 1 m day^{-1} and 10 m day^{-1} were prescribed for phytoplankton and detritus, respectively, following Tian et al. (2015). No explicit sinking velocity was prescribed for zooplankton, consistent with the original formulation and with the fact that zooplankton vertical dynamics are influenced by active transport rather than passive settling alone [D. K. Steinberg & Landry, 2017].

At the ocean open boundaries of the domain, concentrations were prescribed for incoming water masses in all NPZD compartments. Several treatments of incoming NPZD tracers were explored during preliminary simulations, including zero concentrations and concentrations derived from the adjacent model interior. In the retained configuration, inflow concentrations were set equal to the concentration profiles used to initialize the model, thereby assigning background NPZD concentrations to offshore water entering the computational domain. These boundary concentrations were held constant throughout the simulation period, neglecting potential seasonal and event-driven variability in offshore nutrient and plankton concentrations. Although this simplification may in-

introduce a bias in the offshore NPZD fields, its direct influence on the nearshore analyses targeted in this study is expected to be reduced because the open boundaries are located far from the main river-influenced coastal areas. Potential effects transmitted through offshore circulation are nevertheless considered among the model limitations. Finally, riverine DIN inputs are incorporated into the model as external source terms applied to the nutrient compartment (N) at the river boundaries. Daily DIN loads from the four major river systems introduced in Section 2.2.3 (Daintree, Barron, Herbert, and Burdekin Rivers) were obtained from the Queensland Government Water Quality Data Portal⁹. For each river, data from the most downstream available monitoring station were selected, providing the closest available estimate of DIN loads to the river mouth. Their temporal evolution is presented in Fig. 20.

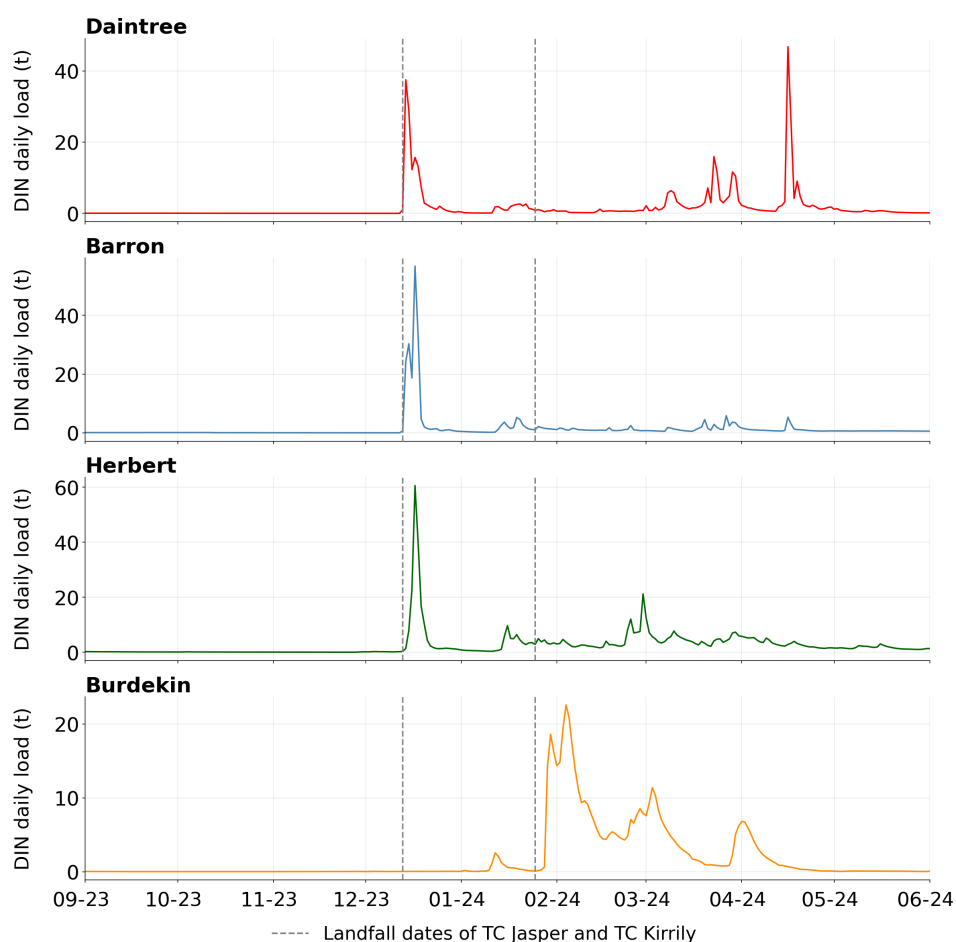


Figure 20: Daily riverine DIN loads from the most downstream available monitoring stations of the four considered river systems.

The four rivers exhibit contrasting temporal patterns in daily DIN loads over the simulation period (Fig. 20). Around TC Jasper, the Daintree, Barron, and Herbert Rivers show abrupt increases, with peak daily loads of approximately 37, 55, and 60 t N, respectively, followed by a rapid decline. Smaller increases occur in all three rivers around mid-January, coinciding with increased freshwater discharge (Fig. 12), but no pronounced DIN-load peak is observed around TC Kirrily. Subsequent variability remains moderate, except for the Daintree River, which reaches approximately 45–50 t N in mid-April. In contrast, Burdekin DIN loads remain close to zero around TC Jasper but increase

⁹<https://apps.des.qld.gov.au/water-data-portal/map>

markedly after TC Kirrily, reaching approximately 20 t N in early February before gradually declining, with smaller secondary increases during March and April.

To quantify riverine DIN inputs associated with each TC, cumulative loads were calculated over three 20-day periods: a reference period from 1 to 20 September 2023, the 20 days following TC Jasper landfall (13 December 2023–1 January 2024), and the 20 days following TC Kirrily landfall (25 January–13 February 2024). The resulting loads are summarised in Table 4.

Table 4: Cumulative riverine DIN loads during the 20-day reference and post-cyclone periods. Values are expressed in tonnes of N.

River	Reference	Post-Jasper	Post-Kirrily
Daintree	0.76	132.07	10.16
Barron	1.72	181.82	25.10
Herbert	3.80	179.77	61.35
Burdekin	0.39	0.77	227.24
Total	6.66	494.43	323.86

The cumulative riverine DIN load increases markedly during both post-cyclone periods compared with the September reference period. A total of 494.43 t N is prescribed during the 20 days following TC Jasper, approximately 74 times the September load of 6.66 t N. Consistent with Fig. 20, this load is concentrated in the three northern rivers, while the Burdekin contribution remains negligible (0.77 t N). Following TC Kirrily, the cumulative load reaches 323.86 t N and is instead dominated by the Burdekin River, which contributes 227.24 t N. The three northern rivers provide smaller but still substantial inputs.

2.4.4 NPZD model evaluation

The NPZD model is evaluated through a combination of internal consistency checks and comparisons with observational data. First, the temporal and vertical behaviour of the simulated NPZD compartments is examined to assess the overall numerical consistency of the biogeochemical simulation. The model is then evaluated against satellite-derived chlorophyll-a observations and in situ DIN measurements collected by James Cook University. Complementary surface and vertical DIN diagnostics are finally used to further characterise simulated nutrient dynamics over the study period.

As a first consistency check, the concentration field of each NPZD compartment is volume-integrated over the entire three-dimensional computational domain at each output time and converted into tonnes of N. The resulting domain-integrated inventories are examined throughout the simulation to identify potential progressive depletion, unbounded accumulation, or abrupt numerical discontinuities. The large-scale vertical distributions of DIN and detritus are additionally examined along a transect extending from the Barron River mouth to the offshore boundary. The transect spans more than 700 km and covers the transition from shallow coastal waters to the deep ocean, reaching depths of approximately 4,000 m. Vertical sections are compared on 1 September 2023, at

the end of the spin-up period, and 30 April 2024, toward the end of the TC season. DIN is used to assess the persistence of the large-scale vertical nutrient structure, while detritus is examined separately because it is initialized at zero and subsequently redistributed through production, transport, and sinking.

Satellite-derived chlorophyll-*a* observations are obtained through NASA Worldview¹⁰. Because satellite coverage varies between dates, particularly due to cloud cover, imagery from MODIS, VIIRS, and OLCI is used according to availability. These products provide near-surface chlorophyll-*a* concentrations expressed in mg m^{-3} . Dates were selected to document pre-cyclone, cyclone, and post-cyclone conditions while prioritizing scenes with sufficient cloud-free coverage over the study area. Twelve dates were retained: 9, 13, 20, 23, 27, and 30 December 2023 for TC Jasper, and 22, 25, and 28 January and 6, 11, and 20 February 2024 for TC Kurrily. For the corresponding dates, simulated phytoplankton concentrations are extracted at 5 m depth. Since satellite chlorophyll-*a* and simulated phytoplankton biomass are not directly equivalent and are expressed in different units, the comparison focuses on large-scale spatial patterns and their temporal evolution around each TC.

A second assessment of the NPZD model is performed using in situ water-quality measurements collected by James Cook University. The dataset comprises 78 measurements collected at 46 sampling locations during four campaigns in December 2023 and March 2024. Among the different water-quality variables available, DIN is calculated as the sum of measured ammonium (NH_4^+), nitrite (NO_2^-), and nitrate (NO_3^-) concentrations. For the present comparison, only observations from 16, 19, and 28 December 2023 are considered in order to evaluate the coastal DIN response immediately following TC Jasper. For each sampling date, three sampling locations distributed across the sampled coastal area are retained, resulting in nine locations for the main analysis (Fig. 21). The selected sampling locations are JCF315, JCF318, and JCF320 on 16 December; JCF301, JCF302, and JCF329 on 19 December; and JCF330, JCF336, and JCF337 on 28 December. Each selected location includes a surface sample (D0) and a subsurface sample (D1), resulting in 18 individual observations, with D1 depths ranging from 3 to 24 m. For each observation, simulated DIN concentrations are extracted at the corresponding sampling location and depth from the model output associated with the sampling date.

In addition to these model evaluation steps, simulated DIN dynamics are examined using complementary surface and vertical diagnostics. First, daily DIN concentrations in the surface layer are mapped from 9 to 14 December 2023 for TC Jasper and from 24 to 29 January 2024 for TC Kurrily to characterise the spatial evolution of nutrient-rich surface waters during each cyclone passage. Then, vertical DIN distributions are examined throughout the simulation period along cross-shelf transects extending from the four river sectors toward the offshore ocean. These vertical sections are analysed together with surface DIN distributions over the full domain and simulation period to identify episodes of upward and landward displacement of DIN-rich subsurface waters. Potential events are identified by visual inspection of the vertical DIN sections as periods showing a coherent upward and landward extension of elevated DIN concentrations from offshore waters toward the continental shelf. To illustrate the temporal evolution of one of these events, the late-January

¹⁰<https://worldview.earthdata.nasa.gov/>

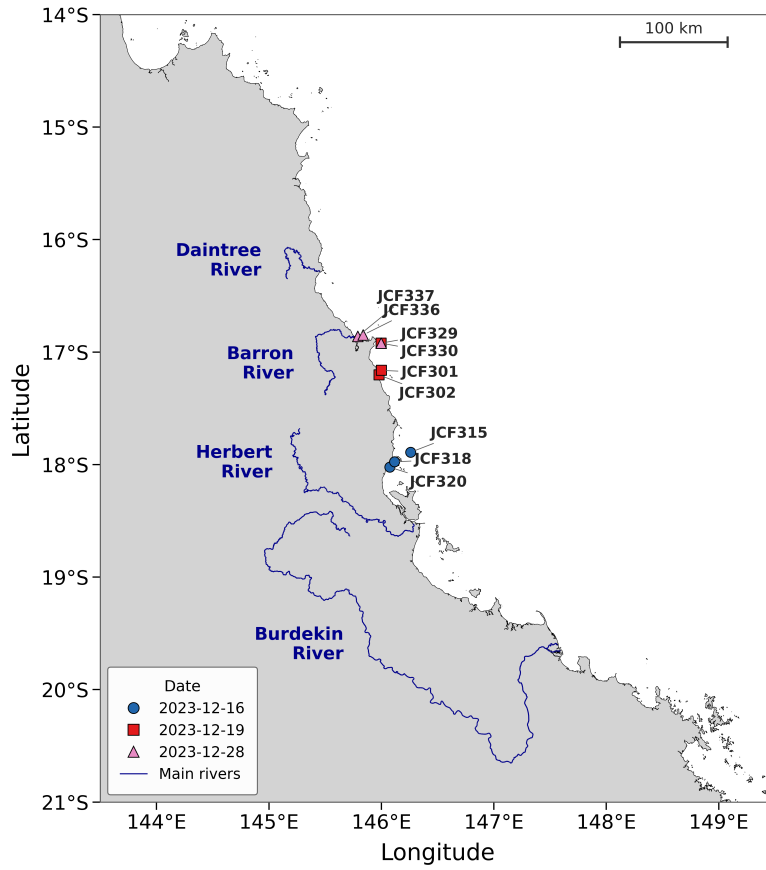


Figure 21: Location of the nine JCU sampling locations retained for the DIN comparison following TC Jasper. Symbols indicate the sampling date.

event along the Barron River–offshore transect is examined in greater detail using vertical sections extracted on 25, 28, and 31 January and 3 February 2024.

2.5 Cumulative phytoplankton exposure assessment

The final step of the modelling framework aims to quantify the cumulative exposure of coral reefs and seagrass meadows to simulated phytoplankton following TC Jasper and TC Kirrily. Phytoplankton concentration is used as a proxy for potential phytoplankton-related light attenuation and is integrated over depth and time to account for its vertical extent and persistence. At each horizontal location, cumulative phytoplankton exposure, E , is calculated as:

$$E(x, y) = \int_{t_0}^{t_1} \int_{-h(x, y)}^{\eta(x, y, t)} P(x, y, z, t) dz dt, \quad (31)$$

where P is the simulated phytoplankton concentration [mmol N m^{-3}], h is the local bathymetric depth, η is the free-surface elevation, and t_0 and t_1 delimit the considered 20-day exposure period, spanning 13 December 2023–1 January 2024 for TC Jasper and 25 January–13 February 2024 for TC Kirrily. The resulting indicator is expressed in $\text{mmol N m}^{-2} \text{ d}$. Because E is integrated over

the full water column, it reflects both the phytoplankton field and the local water-column depth. No normalisation by water depth is therefore applied. Accordingly, comparisons between cyclone periods at the same locations are unaffected by this bathymetric component because the underlying bathymetry is unchanged. Further implications and limitations of the indicator are discussed in Section 4.4.

Coral reef and seagrass meadow distributions are then overlaid on the resulting exposure fields to characterise the exposure of these habitats across the study area. Exposure is summarised separately for coral reefs and seagrass meadows within each NRM region using the area-weighted mean and 90th percentile (P90). The P90 represents the exposure value exceeded by the 10% most exposed habitat area and is used to characterise locally elevated exposure that may not be reflected by the regional mean. Because the available seagrass dataset does not provide complete coverage of offshore meadows, the seagrass analysis is restricted to the mapped habitat distribution.

Finally, the sensitivity of simulated phytoplankton exposure to the prescribed riverine DIN forcing is assessed by comparing two NPZD simulations. The reference simulation includes the DIN inputs from the Daintree, Barron, Herbert, and Burdekin rivers, whereas in the no-river simulation these riverine DIN inputs are removed while the remaining model configuration and hydrodynamic forcing are kept unchanged. Cumulative phytoplankton exposure is calculated over the same post-cyclone periods in both simulations, and their difference is defined as:

$$\Delta E(x, y) = E_{\text{reference}}(x, y) - E_{\text{no-river}}(x, y). \quad (32)$$

Positive values of ΔE therefore indicate greater cumulative phytoplankton exposure when riverine DIN inputs are included, whereas negative values indicate greater exposure in the no-river simulation. This comparison is interpreted as a sensitivity of the simulated phytoplankton response to the prescribed river forcing rather than as a direct attribution of exposure to river-derived DIN.

3 | Results

This chapter presents the results of the three main components of the modelling framework. First, the hydrodynamic model results are presented, followed by the NPZD model and its biogeochemical response to TC Jasper and TC Kirrily. Finally, cumulative phytoplankton exposure is quantified for coral reefs and seagrass meadows.

3.1 Hydrodynamic model

Hydrodynamic model performance is first assessed against the observational datasets described in Section 2.3.4. SLA and ocean currents simulated by SLIM are compared with IMOS observations, while temperature and salinity profiles are compared with Argo measurements. The main validation focuses on NRSYON, GBRPPS, and GBRLSL, representative of the region of interest, and on the two Argo profiles within this region, at (147.13, -17.61) and (145.95, -15.87). Complementary IMOS validation results are presented in Appendix. The simulated circulation during the TC Jasper and TC Kirrily study periods is then characterised using mean current speed and streamlines at 10 m depth.

3.1.1 Sea level anomaly validation

Fig. 22 shows the comparison between observed and simulated SLA at the three selected IMOS stations. The SLIM time series exhibit clear monthly patterns, with two periods of enhanced tidal range separated by approximately two weeks. At NRSYON and GBRPPS, observations and simulations show similar tidal amplitudes and phases throughout the simulation period, despite occasional slight overestimations or underestimations. This results in RMSE values of 0.143 and 0.123 m, respectively. In contrast, GBRLSL shows substantially larger discrepancies. Although SLIM reproduces the main tidal periodicity, the observations display larger amplitudes and stronger high-frequency variability, with peaks reaching approximately 3 m, resulting in an RMSE of 0.624 m.

Validation results for the remaining stations are presented in Appendix. Fig. 38 shows validation patterns similar to those observed at NRSYON and GBRPPS for most stations, with the exception of GBRMYR, where the agreement between observations and simulations is noticeably weaker.

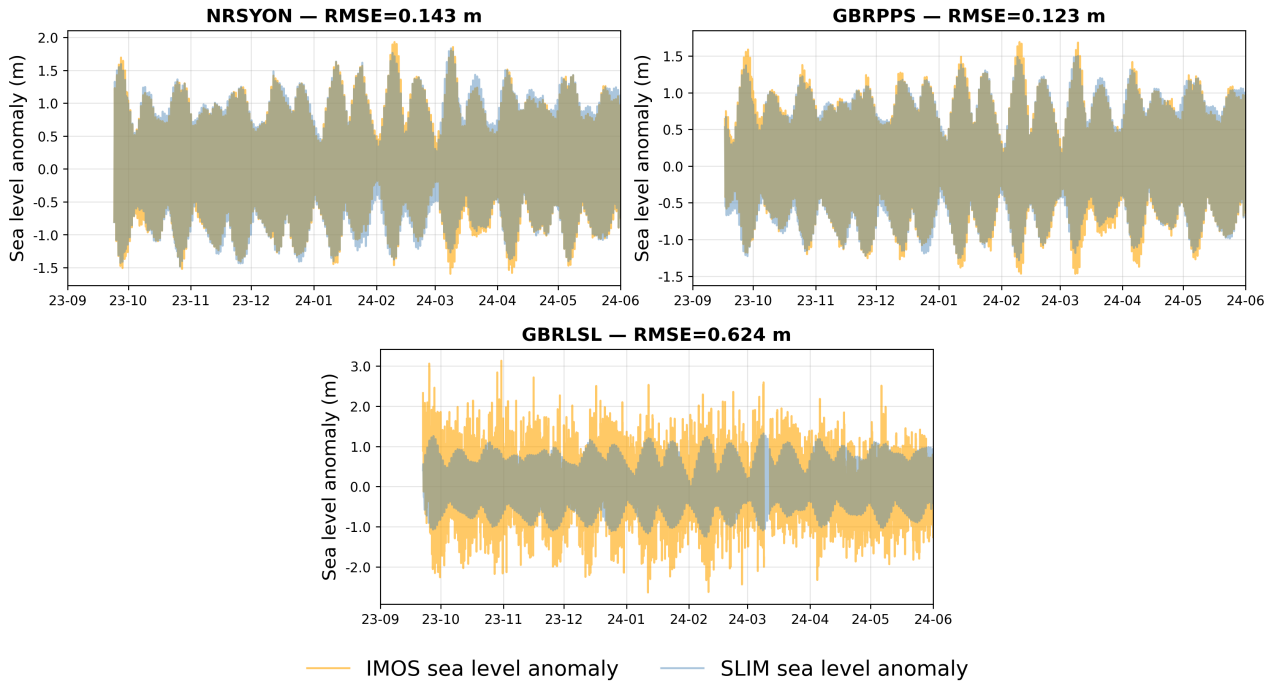


Figure 22: Comparison of sea level anomaly between SLIM and IMOS at the three selected validation stations.

Across the seven stations, RMSE values range from 0.106 to 0.624 m (Appendix, Table 6). NRSYON, GBRCCCH, GBRHIS, GBROTE, and GBRPPS exhibit RMSE values between 0.106 and 0.203 m and normalised RMSE values of approximately 3.5–5%. In contrast, GBRMYR and GBRLSL display RMSE values of 0.412 and 0.624 m, respectively, with normalised RMSE values of 8.2 and 10.8%.

3.1.2 Current validation

Current speed amplitude

Fig. 23 compares the temporal and vertical variability of current speed amplitude between SLIM and IMOS at the three selected stations. Only depth ranges with sufficient-quality observations are shown, excluding the near-surface layer. At NRSYON, SLIM closely reproduces the observed current variability, with periods of enhanced and reduced activity occurring at similar times in both datasets. Current amplitudes also show little vertical structure across the relatively shallow water column. Agreement is weaker at GBRPPS, particularly between September and January, when both datasets show elevated current activity but differ in the timing and magnitude of individual events. From February onward, the alternation between stronger and weaker currents is better reproduced, with amplitudes also closer to the observations. In addition, both datasets consistently show lower current amplitudes below approximately 60 m, indicating that the main vertical structure is captured. GBRLSL exhibits the largest discrepancies, together with an approximately 150 m difference between simulated and observed station depths. Although both datasets show their highest current amplitudes within the upper approximately 200 m, the temporal evolution differs substantially. In particular, SLIM does not reproduce the sustained elevated amplitudes observed

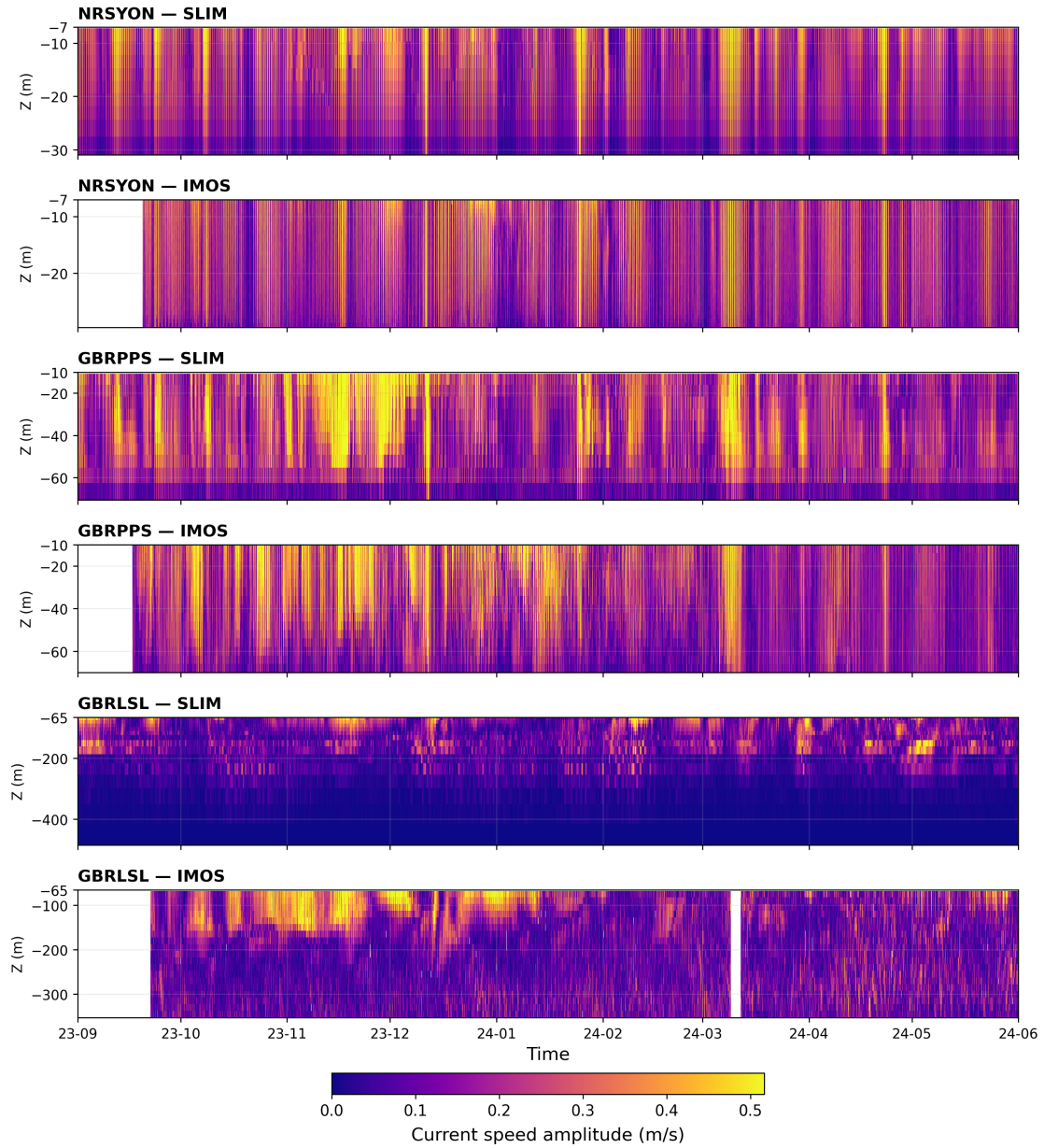


Figure 23: Temporal and vertical distributions of current speed amplitude from SLIM and IMOS at the three selected validation stations.

between October and January and generally underestimates current amplitudes at greater depths.

Complementary Hovmöller diagrams for all validation stations are presented in Appendix (Figures 39–45). Overall, NRSYON, GBRCCCH, GBRHIS, and GBROTE show similar temporal and vertical current-amplitude patterns between SLIM and IMOS. GBRMYR reproduces the main dynamics but generally overestimates current intensity, whereas GBRPPS and GBRLSL exhibit the largest discrepancies between simulations and observations.

The depth-specific validation of current speed amplitude (Appendix, Table 7) is consistent with the patterns identified from the Hovmöller diagrams. NRSYON, GBRCCCH, GBRHIS, and GBROTE generally show the lowest errors, with NRMSE values generally between 8% and 16% across depths. GBRMYR displays intermediate performance, with NRMSE ranging from 14.1% at $z_{\min}$ to 20.5% at

z_{half} . The largest relative errors occur at GBRPPS and GBRLSL, where NRMSE values range from approximately 20 to 25%.

Current direction

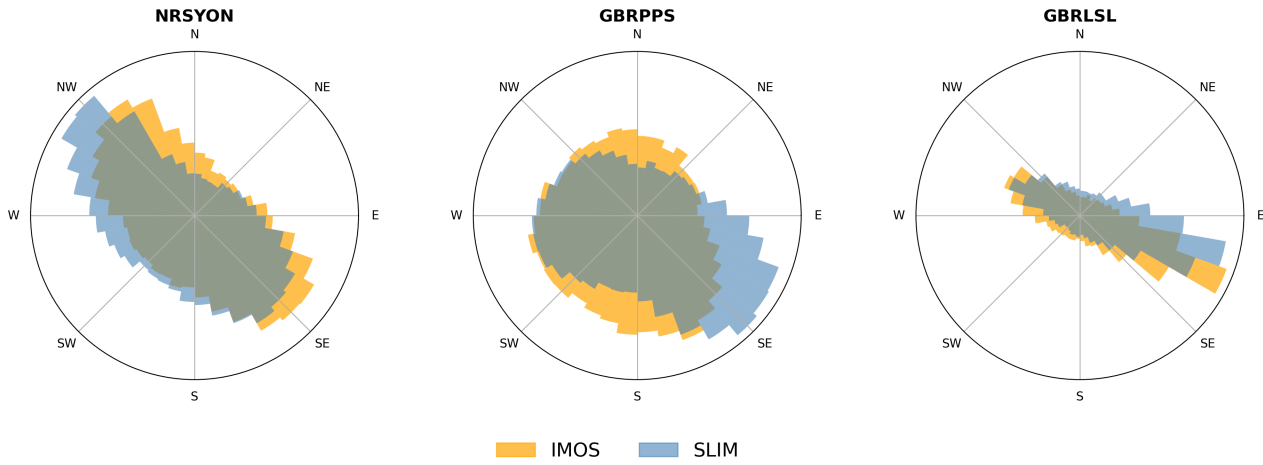


Figure 24: Current direction distributions from SLIM and IMOS at z_{min} for the three selected validation stations.

At the NRSYON station, the direction roses show good agreement between SLIM and the observations, with both datasets displaying a dominant NW–SE bidirectional flow with a similar overall directional distribution. Occurrence frequencies remain substantial over a relatively broad range of directions on either side of this dominant axis, resulting in a broad directional spread. At GBRPPS, both datasets display a relatively broad directional distribution. IMOS observations are relatively evenly distributed across a wide range of directions, with a slight preference for southerly flows. In contrast, SLIM shows a more pronounced concentration toward the E–SE, resulting in an overestimation of occurrence frequencies in this sector, together with a slight underestimation of northerly and southerly directions. At GBRLSL, both datasets display a dominant current orientation along the ESE–WNW axis, with the highest occurrence frequencies directed toward the ESE. Occurrence frequencies in directions outside this axis remain low in both datasets.

The direction roses for the remaining IMOS stations and additional validation depths are presented in Appendix (Figs. 46–48). At z_{min} , GBRCCH, GBRHIS, and GBROTE show close agreement between SLIM and IMOS. GBRMYR shows a larger discrepancy, with observations more broadly distributed toward the S–SE despite a common E–SE preference. Directional patterns remain broadly consistent across depths at most stations, with similar SLIM–IMOS differences. GBRLSL is the main exception, showing stronger depth-dependent changes and increasing disagreement toward greater depths.

Current speed distributions

At NRSYON, SLIM closely reproduces the overall shape of the observed distribution, with a maximum around $0.15\text{--}0.20\text{ m s}^{-1}$ followed by a progressive decrease toward higher speeds. SLIM slightly underestimates the density between 0.10 and 0.20 m s^{-1} and overestimates it between ap-

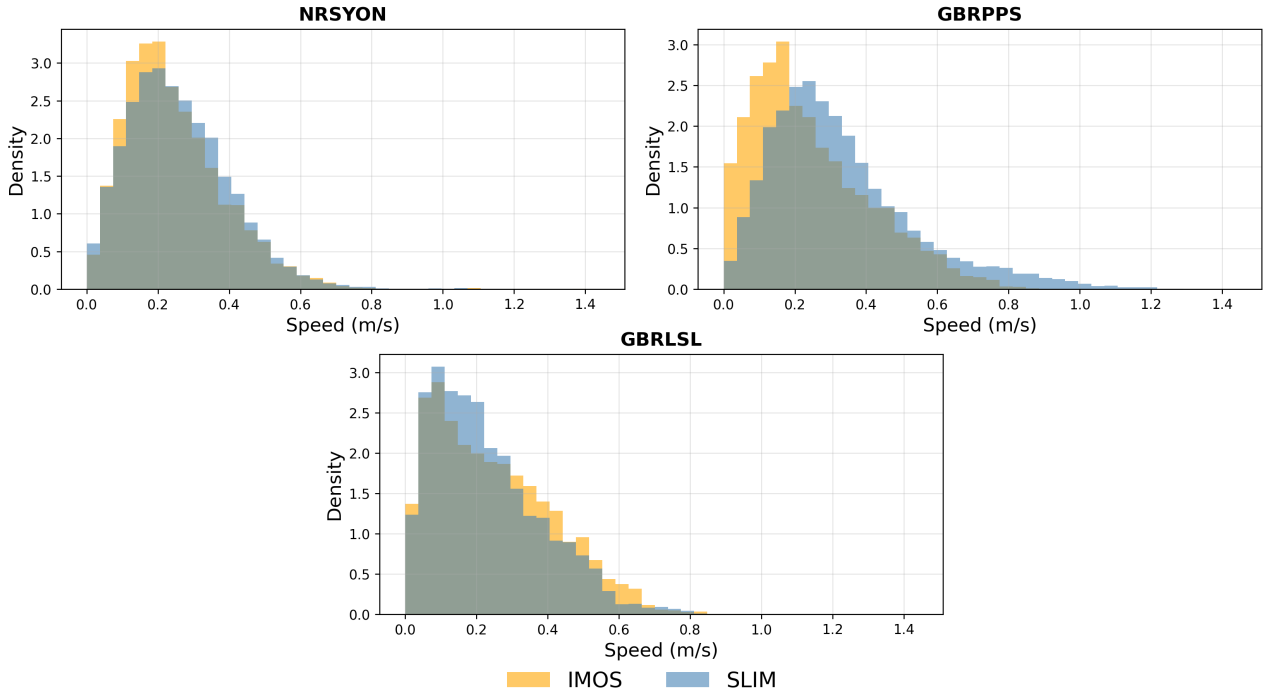


Figure 25: Current speed distributions from SLIM and IMOS at $z_{\min}$ for the three selected validation stations.

proximately 0.25 and 0.50 m s^{-1} . At higher speeds, both distributions progressively converge toward zero. At GBRPPS, agreement is weaker, although both datasets retain a comparable overall distribution. The peak density simulated by SLIM is approximately 0.5 units lower and shifted toward higher current speeds compared with the IMOS observations. SLIM underestimates the density below approximately 0.20 m s^{-1} , while moderate and high speeds are overrepresented, with the simulated distribution extending toward substantially higher speeds than observed by IMOS. At GBRLSL, both distributions are concentrated toward low current speeds, with their highest densities occurring around 0.1 m s^{-1} . SLIM and IMOS are similar at the lowest speeds, although SLIM shows higher densities over approximately 0.10 – 0.30 m s^{-1} . At higher speeds, the two distributions progressively decrease toward zero with relatively limited differences.

Current speed distributions for the remaining IMOS stations and additional validation depths are presented in Appendix (Figs. 49–51). At $z_{\min}$, GBRCCH, GBRHIS, and GBROTE show close agreement between SLIM and IMOS, with similar distribution shapes and peak locations, whereas SLIM generally underestimates current densities at GBRMYR. At z_{half} , the distribution shapes and level of agreement remain broadly similar to $z_{\min}$, but both simulated and observed distributions are generally shifted toward lower current speeds. At z_{station} , SLIM reproduces the observed distributions less closely, particularly at GBROTE and GBRLSL, where low current speeds occur much more frequently than in the observations.

3.1.3 Temperature and salinity validation

Temperature and salinity profiles from SLIM are compared with Argo observations at two profiling locations (Figs. 26 and 27). Both locations lie on the continental slope, and the coarser resolution of SLIM's bathymetry near these steep gradients explains the difference in maximum profile depth.

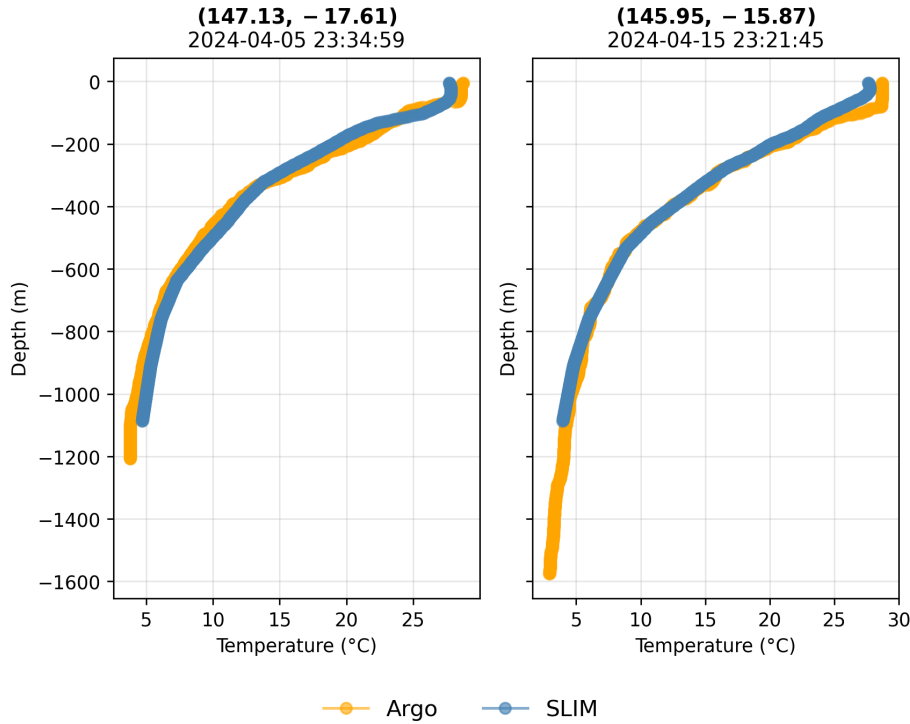


Figure 26: Vertical temperature profiles from SLIM and Argo at the two selected profiling locations.

At the first location, the upper 50 m of the profile are characterised by nearly uniform temperatures of 27–28°C. SLIM closely reproduces the depth of this upper layer, although temperatures are approximately 1°C lower than observed. Below, temperature decreases markedly between approximately 50 and 600 m, with a drop of about 20°C. At greater depths, the decrease becomes more gradual, and SLIM produces slightly warmer values than the observations. A similar vertical temperature structure is observed at the second location, with slightly cooler temperatures in the upper water column, while SLIM places the transition from the upper layer to the stronger temperature gradient at a shallower depth than observed. Below this transition, the simulated profile rapidly converges toward the observations and remains close throughout the remaining simulated depth range. Overall, SLIM reproduces the main features of the observed thermal stratification at both locations, with absolute relative errors of 7.8 and 3.6%, respectively.

The salinity profiles reveal larger discrepancies between SLIM and the Argo observations than those observed for temperature, although the overall magnitude of salinity variation remains modest, spanning approximately 1.5 PSU across the water column. At both locations, the Argo profiles show a similar vertical structure, with salinity increasing from approximately 35 PSU near the surface to a subsurface maximum close to 36 PSU at around 150–200 m. Salinity then decreases markedly between approximately 150 and 500–600 m, reaching about 34.5 PSU, before remaining

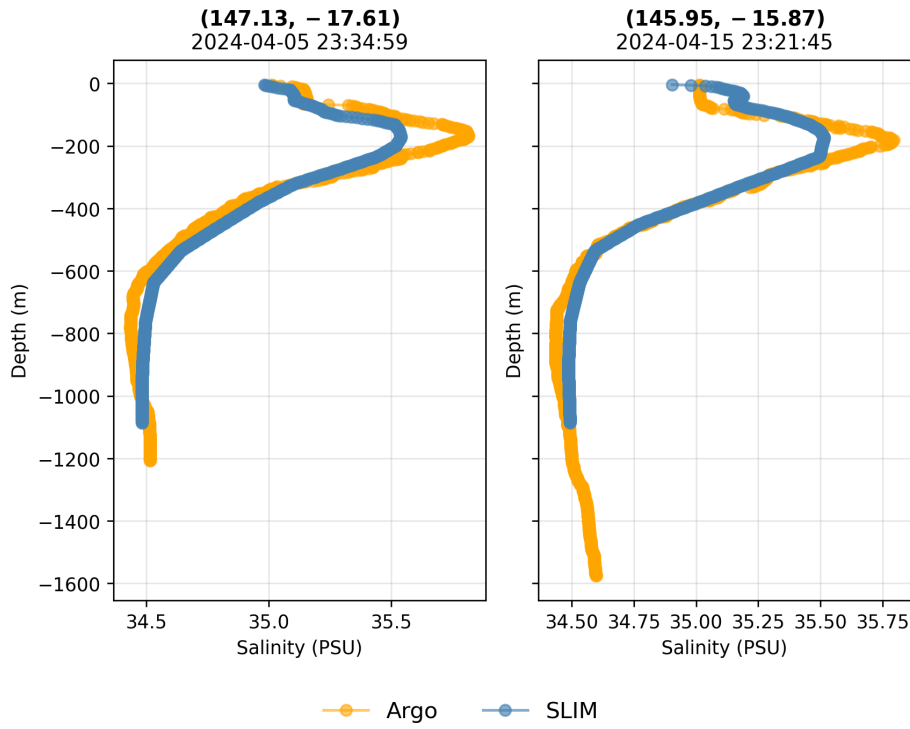


Figure 27: Vertical salinity profiles from SLIM and Argo at the two selected profiling locations.

relatively stable with a slight increase at greater depths. SLIM reproduces this overall vertical pattern but underestimates the subsurface maximum, reaching approximately 35.5 PSU. At greater depths, particularly between approximately 500 and 1000 m, simulated salinity remains higher than observed, although the overall vertical trend is similar. Overall, SLIM successfully reproduces the key structural features of the salinity profiles, with relative errors of 0.22 and 0.16%, respectively.

3.1.4 Circulation patterns around Jasper and Kirrily landfalls

Fig. 28 presents the mean current speed and streamlines at 10 m depth over the study periods associated with TC Jasper and TC Kirrily, extending from five days before to approximately one month each cyclone's landfall. These circulation patterns provide the hydrodynamic context for the advective transport and spatial redistribution of the NPZD tracers examined in the following sections.

Near-surface currents during the Jasper period exhibit a spatially heterogeneous circulation, with a pronounced contrast between the continental shelf and the offshore domain. The dominant feature is a high-speed shelf-break jet, which becomes increasingly well defined south of approximately 17°S, reaching up to 0.8 m s^{-1} . In the western part of the offshore domain, relatively low-speed, quasi-parallel streamlines extend southward toward the shelf edge before joining the jet. Farther east, the circulation is more complex, with several eddy-like recirculation structures that locally modify the flow before it converges toward the jet in the south. Over the continental shelf, current speeds are generally lower, mostly remaining below 0.2 m s^{-1} . In the northern part of the domain, where the shelf is relatively narrow, the inner-shelf current predominantly follows

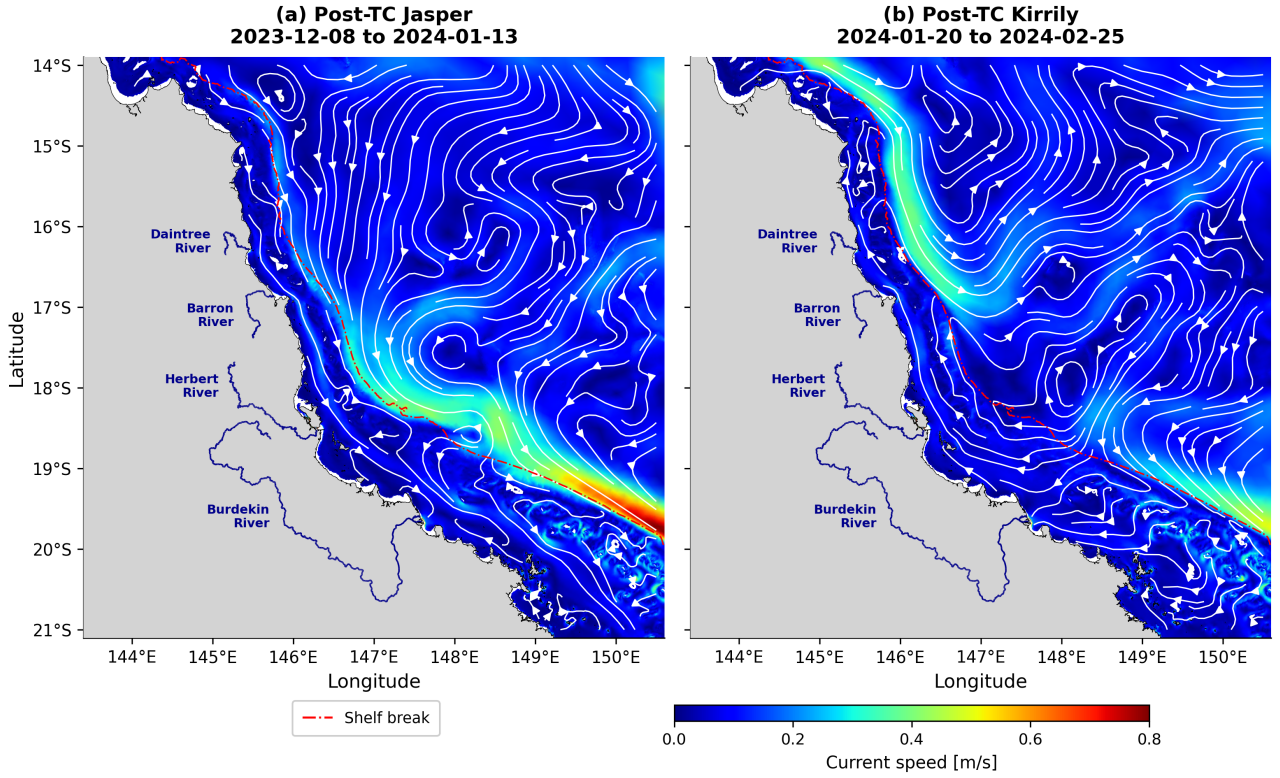


Figure 28: Mean current speed and streamlines at 10 m depth during the study periods associated with (a) TC Jasper, from 2023-12-08 to 2024-01-13, and (b) TC Kirrily, from 2024-01-20 to 2024-02-25.

the coastline toward the southeast. Farther south, as the shelf widens, the inner-shelf circulation becomes more spatially variable, with localized changes in direction and recirculation.

During the Kirrily period, the large-scale circulation displays a more segmented along-shelf organization. Between approximately 14°S and 17°S, an energetic southward jet closely follows the continental shelf break, with velocities locally exceeding 0.4 m s^{-1} . Around 16.5–17°S, it separates from the shelf break and turns offshore, producing a broad region of enhanced velocities extending toward the open ocean. A second shelf-break jet is present farther south, between approximately 19°S and 20°S, where the circulation again converges toward and follows the shelf break. Between these northern and southern branches, the central part of the domain is characterised by several closed and semi-closed recirculation structures. These structures are associated with a northward inner-shelf current that extends to approximately 15°S and remains confined between the coastline and the stronger southward jet. North of approximately 15°S, the inner-shelf flow is instead directed southward, resulting in opposing alongshore current directions within the northern shelf.

3.2 NPZD model

The NPZD model results are first assessed in terms of model behaviour and numerical stability using domain-integrated inventories and vertical DIN and detritus distributions. The biogeochemical response to TC Jasper and TC Kirrily is then examined at complementary spatial scales. Large-scale

near-surface phytoplankton patterns are compared qualitatively with satellite-derived chlorophyll-*a* observations, while simulated coastal DIN concentrations following TC Jasper are compared with water samples collected by James Cook University. Finally, surface DIN dynamics during the cyclone passages and episodic vertical intrusions of DIN-rich deep water are examined.

3.2.1 Model behaviour and numerical stability

Domain-integrated NPZD inventories and vertical DIN and detritus distributions are first examined to assess the general behaviour of the biogeochemical simulation.

Temporal evolution of domain-integrated NPZD inventories

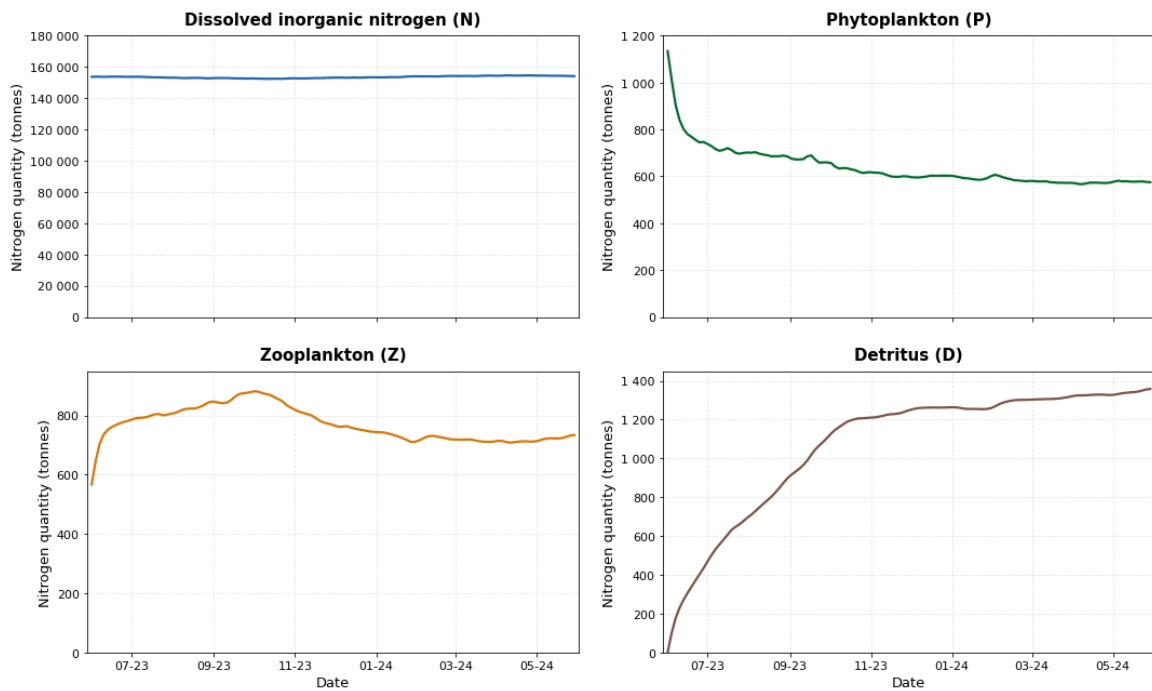


Figure 29: Temporal evolution of the domain-integrated nitrogen inventories in the four NPZD compartments.

As shown in Fig. 29, DIN is the largest compartment throughout the simulation, with an inventory remaining close to 1.5×10^5 tonnes and displaying only limited variations relative to its overall magnitude. During the first weeks of the simulation, the phytoplankton inventory decreases rapidly from approximately 1,100 to 750 tonnes, followed by a more gradual decline to approximately 580 tonnes toward the end of the period. In contrast, the zooplankton inventory initially increases from approximately 560 to 760 tonnes and subsequently reaches a maximum of about 900 tonnes during October. It then decreases progressively to approximately 750 tonnes. Finally, the detrital inventory increases from its initial value of zero throughout the simulation. Its increase is most rapid during the first months, reaching approximately 1,200 tonnes by November, after which the rate of accumulation decreases. Across the analysed period, the four inventories display continuous temporal variations without abrupt discontinuities, while temporal changes generally become less pronounced toward the end of the simulation.

Persistence of the large-scale vertical DIN structure

Preliminary simulations showed a strong sensitivity of the domain-integrated inventories to the treatment of NPZD tracers at the open boundaries. Assigning nearby interior concentrations to inflowing water produced an extreme and unrealistic accumulation of DIN, whereas prescribing zero concentrations produced a progressive depletion of all four compartments. Neither pattern was observed in the retained configuration.

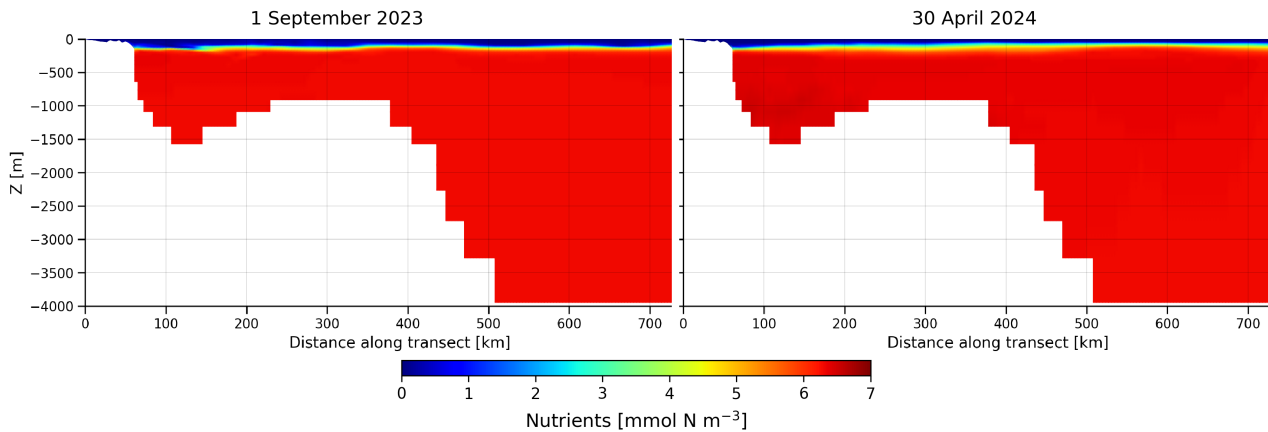


Figure 30: Vertical distribution of DIN along the Barron River–offshore transect on 1 September 2023 (left) and 30 April 2024 (right).

The two transects display a similar large-scale vertical DIN distribution (Fig. 30). At both dates, DIN concentrations are low in the upper water column and typically remain below 1 mmol N m^{-3} over the continental shelf. A pronounced vertical transition occurs at approximately 150–200 m depth, separating the low-DIN upper layer from deeper waters with concentrations exceeding approximately 6 mmol N m^{-3} . Below this transition, concentrations remain high and show limited variation throughout most of the deeper water column. Both the depth of the transition and the overall vertical DIN distribution remain similar between the two dates.

Detritus accumulation

The detrital compartment was analysed separately as a diagnostic of model behaviour for two reasons. First, its initial concentration was set to zero, allowing the progressive development of the detrital pool to be followed throughout the simulation. Second, detritus was assigned a sinking velocity of 10 m d^{-1} , making its vertical redistribution a useful indicator of model behaviour. The implementation of this settling term required particular attention, as preliminary formulations produced numerical instabilities.

Fig. 31 shows a marked contrast in the vertical distribution of detrital nitrogen between the two dates. On 1 September 2023, concentrations remain very low throughout most of the transect, with only weak spatial variations visible within roughly the upper 1,000 m of the water column and negligible concentrations at greater depths. By 30 April 2024, detectable detrital concentrations extend throughout a much larger fraction of the water column. Concentrations are generally higher near the seafloor, with the most pronounced accumulation occurring within a bathymetric depression located approximately 100–200 km from the coast, where values approach 1 mmol N m^{-3} .

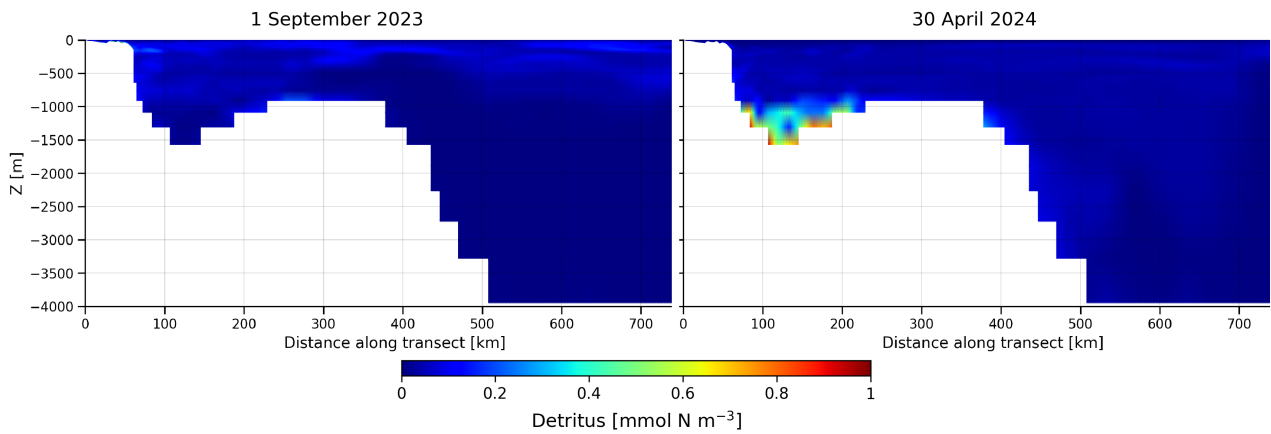


Figure 31: Vertical distribution of detritus along the Barron River–offshore transect on 1 September 2023 (left) and 30 April 2024 (right).

3.2.2 Observed chlorophyll-*a* and simulated phytoplankton comparison

Satellite-derived chlorophyll-*a* observations and simulated phytoplankton concentrations are compared to assess the large-scale spatial patterns and temporal evolution of phytoplankton around TC Jasper and TC Kirrily.

Satellite-derived chlorophyll-*a* observations following the TC events

Across both cyclone periods, satellite-derived chlorophyll-*a* displays a recurrent cross-shelf organization, with the highest concentrations in nearshore waters, spatially variable intermediate concentrations ($0.1\text{--}1\text{ mg m}^{-3}$) across the continental shelf, and low concentrations offshore (Fig. 32). Before TC Jasper, shelf concentrations are generally around 0.5 mg m^{-3} . Following landfall, elevated chlorophyll-*a* concentrations expand toward the mid-shelf across much of the northern study area, with local maxima approaching 20 mg m^{-3} on 23 December. The intermediate-concentration zone also becomes more spatially continuous, with fewer low-concentration patches across the shelf. By the end of December, this expanded zone contracts, although elevated values persist over parts of the outer shelf and within the northern coastal embayment. Before TC Kirrily, the 22 January image already shows a broader coastal band of elevated chlorophyll-*a* than before TC Jasper, together with a heterogeneous zone of intermediate concentrations extending across much of the shelf. Following TC Kirrily, the clearest change occurs on 28 January in the central and southern sectors, where the coastal band broadens and the intermediate-concentration zone forms a more continuous band across the shelf. Compared with TC Jasper, however, the post-cyclone expansion of the high-concentration coastal band is less pronounced. During February, the elevated chlorophyll-*a* signal becomes progressively more confined to coastal waters.

Simulated surface phytoplankton following the TC events

In the simulations, phytoplankton concentrations are generally intermediate over the continental shelf ($0.03\text{--}0.3\text{ mmol N m}^{-3}$) and lower and more heterogeneous beyond the shelf break (Fig. 33). Before TC Jasper, a substantial part of the domain beyond the shelf break is characterised

by concentrations below $0.005 \text{ mmol N m}^{-3}$, with no marked maximum along the coast. Following its landfall, the main spatial change occurs offshore, where intermediate concentrations progressively expand between 13 and 23 December, becoming spatially continuous with the intermediate-concentration field over the shelf. By the end of December, offshore concentrations decrease, while a localized coastal maximum develops near the Daintree and Barron river mouths, reaching approximately 1 mmol N m^{-3} . Before TC Kirrily, the coastal–offshore contrast resembles pre-Jasper conditions on the shelf, except for a coastal maximum observable in the northern sector. After the cyclone, the southern offshore sector shows an increase in the spatial extent of intermediate phytoplankton concentrations, while in the north, intermediate concentrations become increasingly widespread across both the shelf and adjacent offshore waters. This pattern reaches its greatest offshore extent on 6 February. At this date, the northern coastal maximum also reappears, together with a relatively elevated and spatially homogeneous area of phytoplankton concentrations extending across the northern shelf, with values around $0.2 \text{ mmol N m}^{-3}$. Concentrations subsequently decrease in both coastal and offshore waters by 11–20 February.

Comparison of observed and simulated patterns

Overall, simulated surface phytoplankton shows a broad cross-shelf organization similar to that observed in satellite-derived chlorophyll-*a*, with intermediate concentrations over the shelf and lower values offshore. However, the post-cyclone evolution differs markedly between the two datasets. Following TC Jasper, satellite observations show a rapid and extensive expansion of high chlorophyll-*a* concentrations from the coast toward the mid-shelf in the northern sector, whereas the simulation mainly displays increasing concentrations offshore, with little immediate change in the coastal band. Coastal phytoplankton concentrations increase later in the simulation, but the relative increase from pre-cyclone shelf conditions is smaller and more spatially restricted than in the satellite observations. A similar discrepancy is observed following TC Kirrily: satellite imagery shows a clear broadening of the coastal high-concentration band, while the simulated response is dominated by an offshore expansion of intermediate concentrations, with more limited coastal changes. In addition, the simulation develops a relatively homogeneous band of intermediate phytoplankton concentrations across the northern shelf, particularly by 6 February, whereas this shelf-wide feature is less clearly expressed in the corresponding satellite-derived chlorophyll-*a* observations. Thus, the model captures the persistent shelf–offshore spatial structure, but differs in the relative magnitude, timing, and spatial extent of the post-cyclone coastal response.

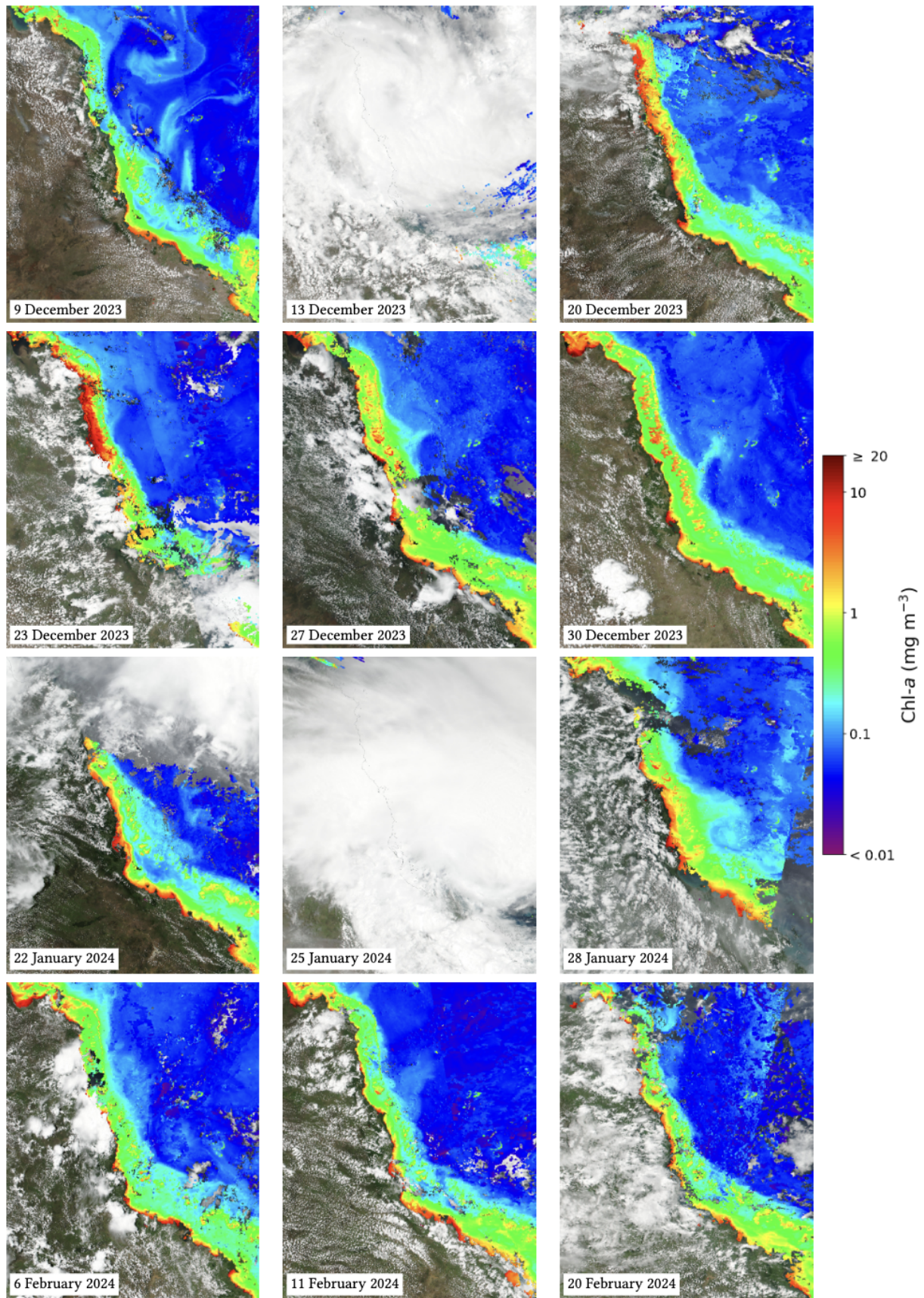


Figure 32: Temporal evolution of satellite-derived surface chlorophyll-*a* concentrations before and after TC Jasper (first six panels) and TC Kirrily (last six panels). The selected dates document conditions immediately before, during, and after each cyclone. Chlorophyll-*a* concentrations are displayed on a logarithmic colour scale.

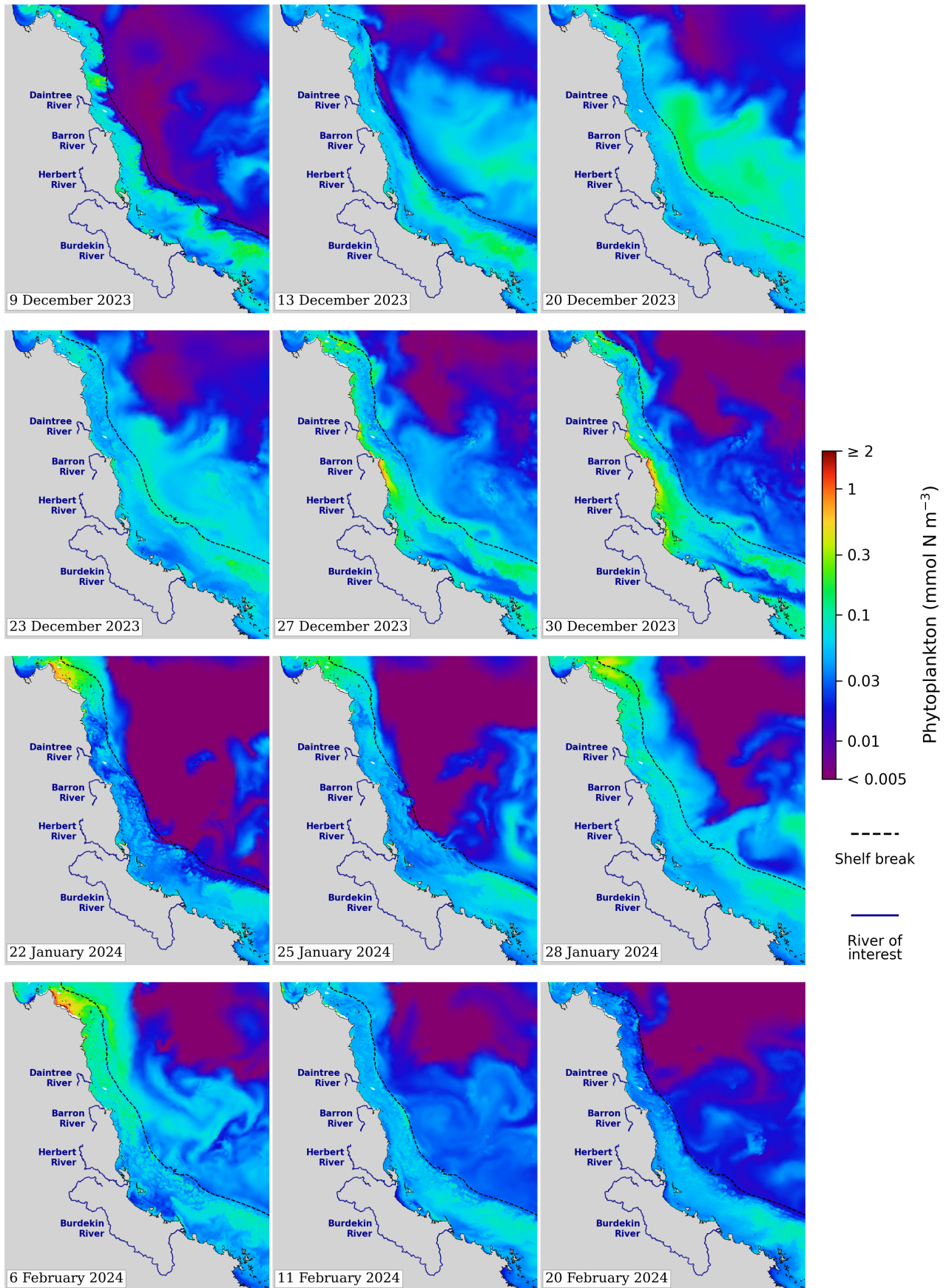


Figure 33: Temporal evolution of simulated phytoplankton concentrations at 5 m depth before and after TC Jasper (first six panels) and TC Kirrily (last six panels). The panels correspond to the dates presented in Fig. 32, and concentrations are displayed on a logarithmic colour scale.

3.2.3 Observed and simulated coastal DIN concentrations

Fig. 34 compares observed and simulated DIN concentrations at the nine selected observation points for the three December sampling dates. Simulated concentrations are additionally shown on an expanded scale in Appendix (Fig. 52).

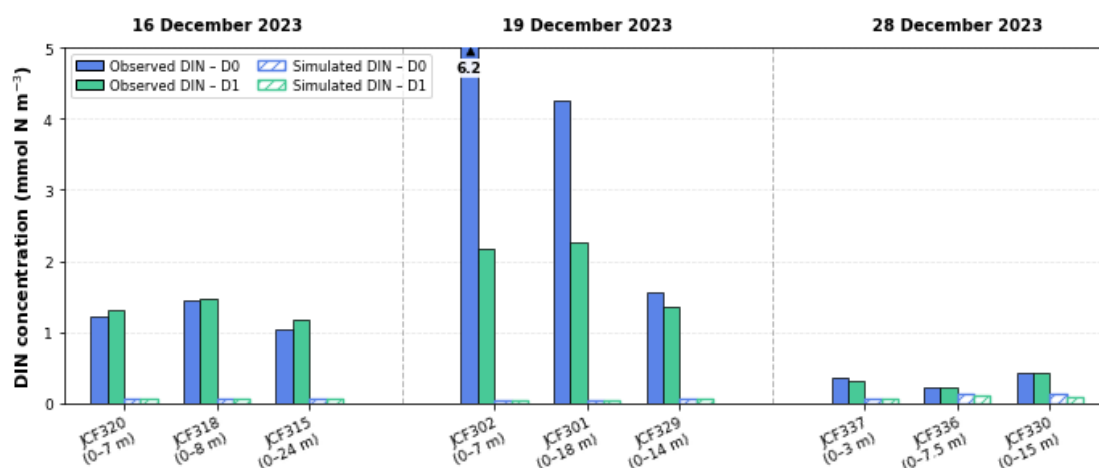


Figure 34: Observed and simulated DIN concentrations at the nine selected sampling locations on 16, 19, and 28 December 2023. Observations and simulations are shown at the surface (D0) and at the corresponding subsurface sampling depth (D1).

The samples collected on 16 December north of the Herbert River show DIN concentrations of approximately 1–1.5 mmol N m⁻³, with limited differences between D0 and D1 (Fig. 34). Simulated concentrations range from approximately 0.06 to 0.08 mmol N m⁻³ across these sampling locations and show little variation between depths (Appendix, Fig. 52). On 19 December, the sampling locations south of the Barron River display the highest observed surface concentrations, particularly JCF302 and JCF301, with 6.2 and 4.3 mmol N m⁻³, respectively. At these two locations, surface concentrations are substantially higher than those measured at D1, whereas the difference between depths is smaller at JCF329. Simulated DIN concentrations on this date increase slightly from JCF302 to JCF329, from approximately 0.05 to 0.06 mmol N m⁻³, while differences between D0 and D1 remain limited. Finally, the samples collected near the Barron River mouth on 28 December show lower observed concentrations, generally below 0.5 mmol N m⁻³, with only minor differences between D0 and D1. On this date, simulated DIN concentrations reach their highest values among the nine selected sampling locations at JCF336 and JCF330, approaching 0.13 mmol N m⁻³, with D0 concentrations exceeding D1 at both locations. Overall, simulated DIN concentrations remain substantially lower than the corresponding observations at all sampling locations, while the observed spatial and temporal variability across sampling locations and dates is not reproduced. The discrepancy is particularly pronounced on 19 December, when the observed surface DIN concentration of 6.2 mmol N m⁻³ is approximately 120 times higher than the simulated value of 0.05 mmol N m⁻³.

3.2.4 Surface DIN response to TCs and episodic vertical DIN dynamics

The previous results show marked temporal and spatial variability in simulated phytoplankton during the periods following the cyclone events, despite comparatively low simulated coastal DIN concentrations. To further characterise the simulated nutrient dynamics, two complementary analyses were performed. First, daily surface DIN distributions were examined over the model domain during the passages of TC Jasper and TC Kirrily to characterise the immediate surface nutrient response to each cyclone. Second, vertical DIN distributions were analysed throughout the simulation along cross-shelf transects associated with the four river sectors to identify episodic upward and landward displacements of DIN-rich subsurface waters. The late-January event along the Barron River–offshore transect is presented below as the clearest example of this process.

Daily surface DIN distributions show an overall increase in concentration and in the spatial extent of relatively high-DIN waters across the domain during both cyclone passage (Appendix, Fig. 53). These increases develop preferentially along and around the cyclone trajectories and extend well beyond the immediate vicinity of the river mouths. The response is more pronounced during TC Jasper, for which surface DIN increases from approximately $0.005\text{--}0.1\text{ mmol N m}^{-3}$ before the passage to values locally exceeding 0.3 mmol N m^{-3} , while the area occupied by these higher concentrations expands markedly by 13–14 December. A similar but weaker evolution occurs during TC Kirrily, with progressively higher concentrations extending over a broader area in the region crossed by the cyclone. From 27 January onward, particularly high surface DIN concentrations also develop north of the Daintree River, locally reaching or exceeding 2 mmol N m^{-3} .

In parallel, the vertical DIN distribution along the Barron River–offshore transect reveals episodic intrusions of DIN-rich deep water toward shallower parts of the continental shelf. The sequence from 25 January to 3 February 2024 was selected as the clearest example of this process.

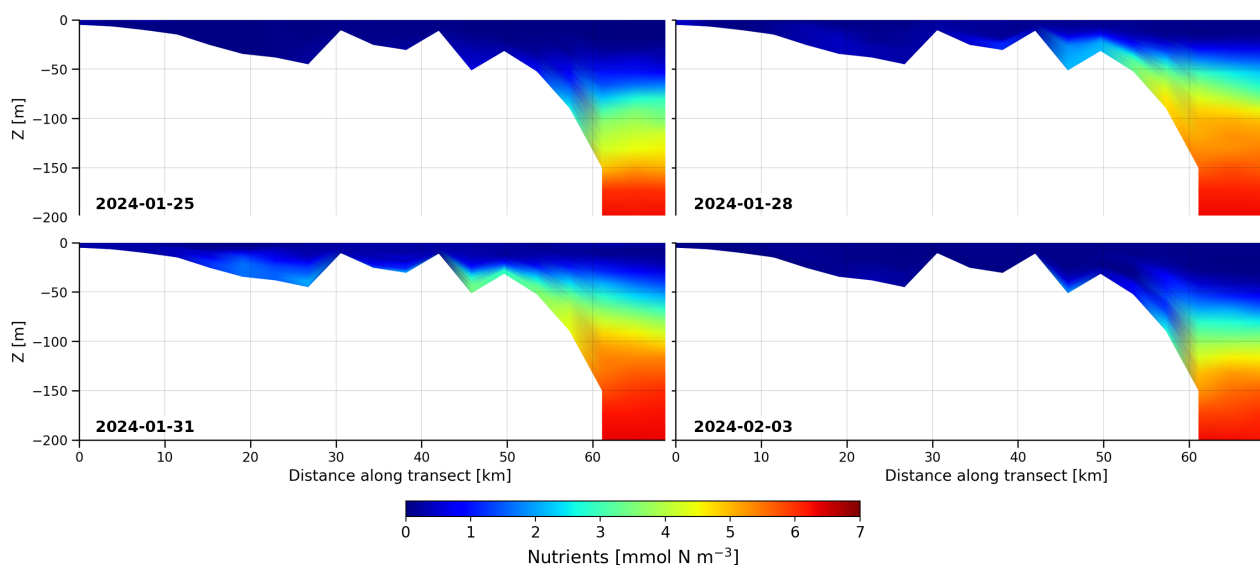


Figure 35: Temporal evolution of the vertical DIN distribution along the Barron River–offshore transect on 25, 28, and 31 January and 3 February 2024.

The four transects show a marked temporal evolution of the cross-shelf DIN distribution (Fig. 35). On 25 January, DIN concentrations are very low throughout the coastal and inner-shelf waters, while elevated concentrations are confined to the offshore end of the transect, with values above 5 mmol N m^{-3} restricted to depths greater than approximately 150 m. By 28 January, intermediate DIN concentrations extend upward and landward, reaching depths of approximately 50 m around 50 km along the transect. Simultaneously, lower but still elevated concentrations spread farther across the shelf toward the coastal sector. This landward extension is most pronounced on 31 January, when concentrations of approximately 2 mmol N m^{-3} extend nearly to the coastal end of the transect. By 3 February, the pattern weakens, with elevated DIN concentrations again becoming more restricted to the deeper offshore part of the transect.

Similar upward intrusions of DIN-rich deep water were identified over the full simulation period by examining surface DIN concentrations together with vertical transects across the four river sectors. Five marked events were identified in the region of interest, with approximate dates of 20 and 28 November, 24 December, 11 January, and 28 January, although their timing and intensity vary spatially across the region. Among these events, the late-January event shows the most pronounced upward and landward displacement of DIN-rich water.

3.3 Cumulative phytoplankton exposure as a proxy for potential light attenuation

This section examines cumulative water-column phytoplankton exposure as a proxy for potential light attenuation following TC Jasper and TC Kirrily. Spatial exposure patterns are first characterised, before assessing the influence of riverine inputs by comparing the reference simulation with a simulation without river inputs.

3.3.1 Spatial patterns of cumulative phytoplankton exposure

Exposure was calculated over the 20 days following each cyclone, from 13 December 2023 to 1 January 2024 for TC Jasper and from 25 January to 13 February 2024 for TC Kirrily. Coral reef and seagrass meadow distributions were overlaid on the resulting exposure maps to assess their spatial correspondence with simulated phytoplankton exposure. The exposure analysis does not encompass the full seagrass distribution presented in Fig. 4 because the available habitat data provide incomplete coverage of offshore meadows.

During the 20 days following TC Jasper, cumulative phytoplankton exposure exhibits a pronounced cross-shelf gradient (Fig. 36). Values generally remain below $25 \text{ mmol N m}^{-2} \text{ d}$ along the coast, particularly in the Herbert and Burdekin sectors, and frequently reach $100 \text{ mmol N m}^{-2} \text{ d}$ toward the seaward part of the shelf. Two coastal areas depart from this general pattern, with comparatively higher exposure north of the Daintree River and south of the Barron River. At the

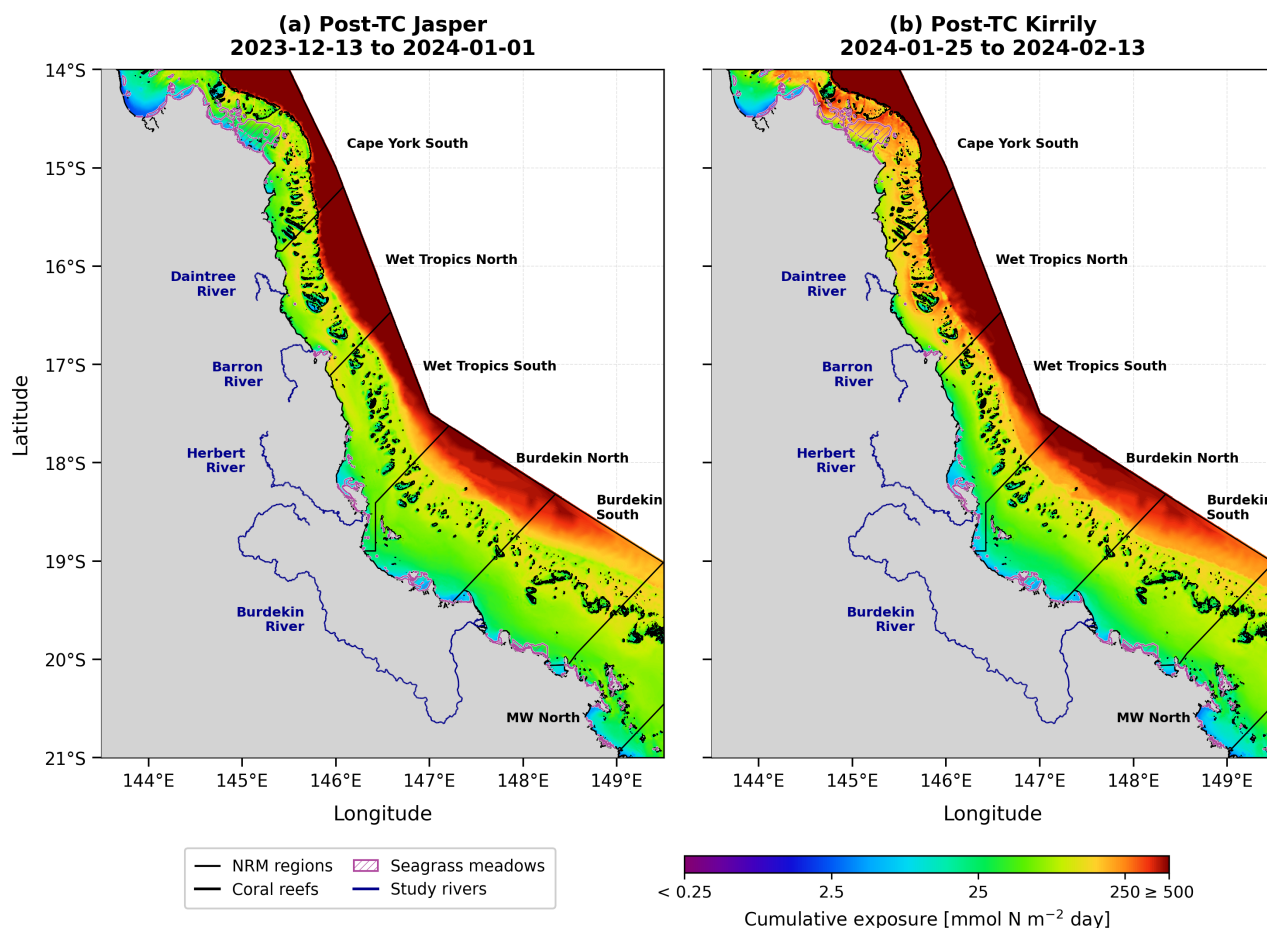


Figure 36: Cumulative water-column phytoplankton exposure during the 20 days following (a) TC Jasper, from 2023-12-13 to 2024-01-01, and (b) TC Kirrily, from 2024-01-25 to 2024-02-13.

local scale, reef-dense areas show pronounced heterogeneity, with recurrent lower exposure values around many reefs compared with the surrounding waters. At the broader scale, however, most coral reefs are located toward the outer shelf and overlap a similar range of intermediate exposure values. In contrast, seagrass meadows are predominantly located in nearshore waters, where exposure is generally lower. During the 20 days following TC Kirrily, the overall cross-shelf distribution remains similar to that observed after TC Jasper, with lower cumulative exposure in coastal waters and higher values toward the offshore part of the shelf. However, offshore exposure is generally higher than during the post-Jasper period in the northern part of the study area (Wet Tropics North and Cape York South), where more extensive areas reach 250–500 $\text{mmol N m}^{-2} \text{ d}$. Farther south, exposure remains comparatively lower and closer to the levels observed following TC Jasper. Coral reefs consequently occur across a broader range of exposure levels than during the post-Jasper period, from comparatively lower values in the central and southern regions to much higher values in the north. Seagrass meadows remain predominantly associated with lower nearshore exposure, except for the northernmost mapped area, which overlaps comparatively high exposure values.

Table 5 quantifies the regional exposure patterns for coral reefs and seagrass meadows. Overall, coral reefs show higher cumulative exposure than seagrass meadows across all regions and during both cyclone periods. A broad latitudinal contrast is observed between the two events. In the two northernmost regions, mean coral reef exposure is higher following TC Kirrily, increasing from

Table 5: Regional cumulative water-column phytoplankton exposure of coral reefs and seagrass meadows during the 20-day periods following tropical cyclones Jasper and Kirrily.

2*NRM region	Coral reefs				Seagrass meadows			
	Jasper		Kirrily		Jasper		Kirrily	
	Mean	P90	Mean	P90	Mean	P90	Mean	P90
Cape York South	47.7	128.3	123.3	289.2	21.4	49.7	103.7	217.2
Wet Tropics North	41.3	82.1	75.2	145.0	19.9	45.1	19.2	38.7
Wet Tropics South	32.1	56.8	38.1	72.2	10.3	15.7	6.0	8.4
Burdekin North	42.7	67.8	42.4	75.0	7.7	12.7	7.7	12.6
Burdekin South	46.1	70.1	38.6	62.7	17.8	33.9	12.0	22.7
MW North	37.3	64.1	30.5	56.4	9.6	20.3	6.8	13.4

Exposure is expressed in $\text{mmol N m}^{-2} \text{ d}$. Mean represents the average exposure across the mapped habitat area. P90 denotes the area-weighted 90th percentile, i.e. the exposure value exceeded by the 10% most exposed habitat area.

47.7 to 123.3 $\text{mmol N m}^{-2} \text{ d}$ in Cape York South and from 41.3 to 75.2 $\text{mmol N m}^{-2} \text{ d}$ in Wet Tropics North. Seagrass exposure also increases strongly in Cape York South (21.4 to 103.7 $\text{mmol N m}^{-2} \text{ d}$) but remains nearly unchanged in Wet Tropics North (19.9 and 19.2 $\text{mmol N m}^{-2} \text{ d}$). Wet Tropics South and Burdekin North show smaller and less consistent differences between the two periods, with coral reef means of 32.1 versus 38.1 and 42.7 versus 42.4 $\text{mmol N m}^{-2} \text{ d}$, respectively. Farther south, TC Jasper produces higher mean exposure for both ecosystem types, for example 46.1 versus 38.6 $\text{mmol N m}^{-2} \text{ d}$ for coral reefs and 17.8 versus 12.0 $\text{mmol N m}^{-2} \text{ d}$ for seagrass meadows in Burdekin South. Cape York South records the highest mean exposure for both ecosystem types during both cyclone periods. Following TC Kirrily, P90 values reach 289.2 $\text{mmol N m}^{-2} \text{ d}$ for coral reefs and 217.2 $\text{mmol N m}^{-2} \text{ d}$ for seagrass meadows, substantially exceeding the corresponding means and highlighting the high exposure reached within the most exposed 10% of the habitat area.

3.3.2 Influence of riverine inputs on cumulative exposure

Cumulative phytoplankton exposure is compared between the reference and no-river simulations over the same 20-day post-cyclone periods. Differences are calculated as reference minus no-river exposure values.

The difference in cumulative phytoplankton exposure between the reference and no-river simulations remains relatively limited during the 20 days following TC Jasper, with differences generally ranging between approximately -50 and $50 \text{ mmol N m}^{-2} \text{ d}$ across much of the domain (Fig. 37). The spatial pattern is nevertheless heterogeneous, with alternating positive and negative differences. Negative values are mainly concentrated in a coastal area of the central domain, together with two offshore sectors at the northern and southern extremities of the study area. Moderate positive differences occur over much of the remaining shelf, with the strongest values observed in the northern offshore sector. In contrast, the post-Kirrily period is clearly dominated by positive differences. The strongest and most continuous signal occurs in the northern part of the domain, where positive values extend from approximately 14°S to 18°S and locally approach $200 \text{ mmol N m}^{-2} \text{ d}$.

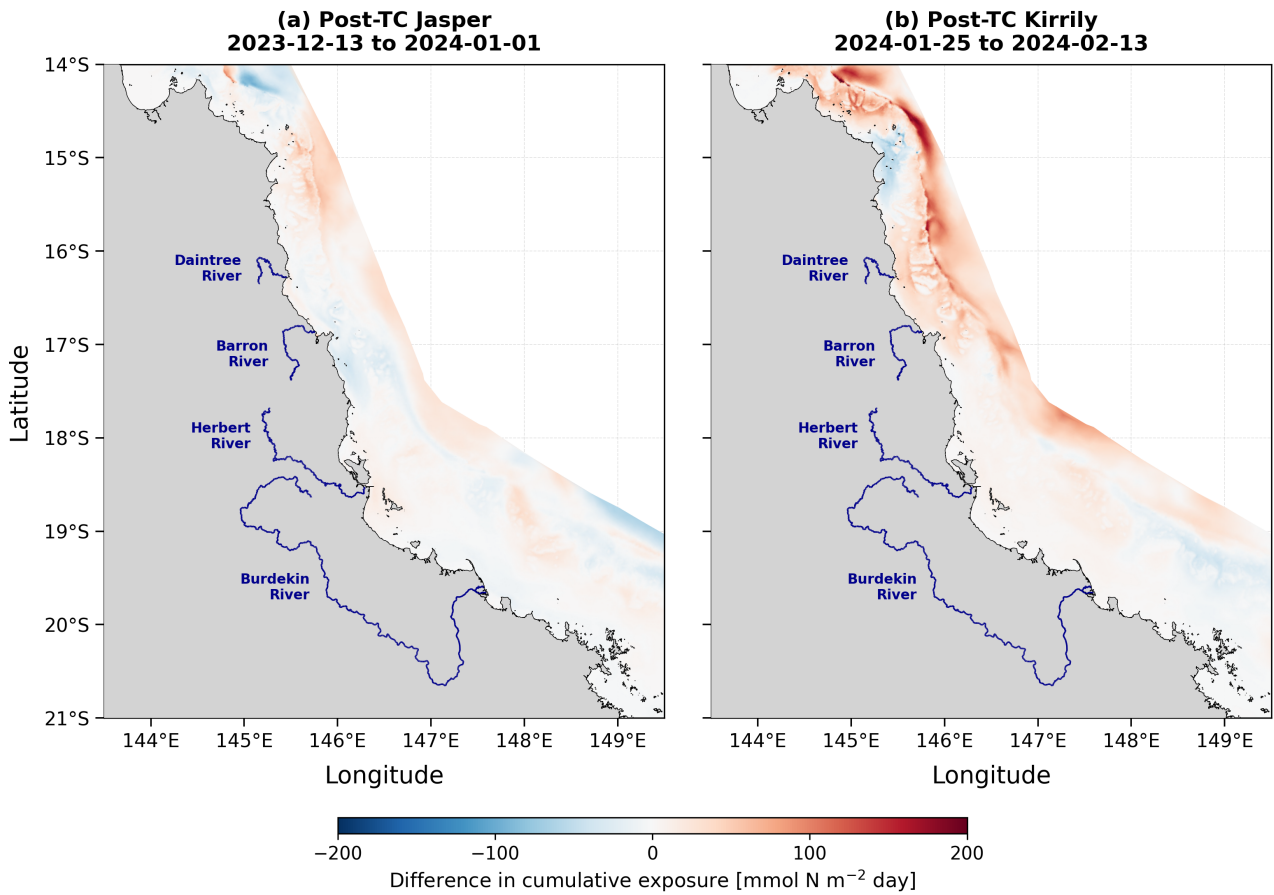


Figure 37: Difference in cumulative water-column phytoplankton exposure between the reference and no-river simulations during the 20 days following (a) TC Jasper and (b) TC Kirrily. Positive values indicate higher cumulative exposure when riverine inputs are included.

d. Negative differences remain locally visible near the coast around 15°S and farther south, where the overall contrast between the two simulations becomes weaker. Overall, the post-Kirrily period shows substantially larger and more consistently positive differences than the post-Jasper period.

4 | Discussion

This discussion interprets the main results of the coupled hydrodynamic–biogeochemical modelling framework in relation to the objectives of this thesis. First, the performance of the hydrodynamic model is assessed in the context of the known circulation and hydrographic characteristics of the GBR region. The biogeochemical response is then discussed, with particular attention to the representation of post-cyclone DIN and phytoplankton dynamics and to the respective roles of riverine and oceanic nutrient supply. Finally, the cumulative phytoplankton exposure of coral reefs and sea-grass meadows is interpreted, before discussing the main limitations of the modelling framework and perspectives for future development.

4.1 Hydrodynamic model performance

Overall, the hydrodynamic validation shows that SLIM reproduces the main physical characteristics of the GBR and adjacent Coral Sea, although model performance varies spatially and with depth. The dominant sea-level variability is well captured at most stations, while current amplitudes, flow orientations and speed distributions are generally reproduced, with larger discrepancies at some deeper and bathymetrically complex locations (GBRSL and GBRMYR IMOS stations). The simulated temperature and salinity profiles also reproduce the main regional hydrographic structure. At the broader scale, the near-surface circulation is broadly consistent with established circulation patterns of the GBR, although some regional features, particularly around 15°S, are not fully reproduced.

The SLA validation indicates that SLIM reproduces the dominant sea-level variability over most of the study area, including the recurrent ~14-day modulation of tidal amplitude associated with the spring–neap cycle documented across the GBR [Mawson et al., 2022; Pickard et al., 1977]. At five of the seven IMOS stations, simulated and observed time series show similar tidal amplitudes and phases, with NRMSE values ranging from 3.6 to 5.1% (Fig. 38, Appendix, Table 6). The main exceptions are GBRMYR and GBRSL, where NRMSE values increase to 8.2 and 10.8%, respectively. These two stations correspond to the deepest deployments considered in the validation. At both locations, and particularly at GBRSL, SLIM simulates a comparatively regular sea-level signal with a smaller range of amplitude, whereas the IMOS observations exhibit substantially greater high-frequency fluctuations. The larger discrepancies therefore appear to reflect not only differences in

the simulated tidal signal, but also variability present in the observations that is not reproduced by the model.

Current validation indicates that SLIM reproduces the main characteristics of the observed circulation at most stations, although performance varies spatially and with depth. Current amplitudes are generally best reproduced at NRSYON, GBRCCH, GBRHIS and GBROTE, with NRMSE values mostly ranging from approximately 8 to 15%. GBRMYR shows intermediate agreement, whereas the largest discrepancies occur at GBRPPS and GBRLSL, where NRMSE values reach up to 25% (Appendix, Table 7). GBRMYR and GBRLSL are the only two stations located directly at the shelf break and are both considerably deeper than the other validation stations. Their greater depth places a larger fraction of the water column within the geopotential z -layer portion of the vertical grid, where steep bathymetric gradients are less accurately resolved. This is reflected in differences between the simulated and observed station depths of approximately 30 m at GBRMYR and 150 m at GBRLSL. The combination of greater water depth, steep shelf-break bathymetry and imperfect representation of the local depth may therefore contribute to the larger discrepancies observed at these stations, particularly at GBRLSL. The current-direction analysis nevertheless shows that the dominant flow orientations are generally well represented. At GBRMYR and GBRLSL, the dominant directions at both $z_{\min}$ and z_{half} broadly follow the local orientation of the shelf break in both IMOS and SLIM, while at the coastal NRSYON station, the dominant NW–SE axis is approximately parallel to the coastline. At GBRLSL, directional agreement becomes weaker at greater depth, consistent with reduced model accuracy in the deepest layers at this station. The current-speed distributions reveal contrasting patterns at the two shelf-break stations. At GBRMYR, noticeable discrepancies in distribution shape are already present at both $z_{\min}$ and z_{half} . At GBRLSL, in contrast, the distributions are relatively well reproduced at these two depths, while agreement deteriorates markedly at z_{station} , where simulated currents are strongly concentrated toward very low speeds and exhibit little variability compared with the observations. Overall, SLIM reproduces the dominant current amplitudes, directions and speed distributions over most of the study area, with greater uncertainty at some deeper and bathymetrically complex locations.

At the Argo validation locations, SLIM reproduces the observed temperature profiles well, capturing the warm upper waters and the progressive decrease through the thermocline. Agreement is slightly weaker for salinity: the observed subsurface maximum reaches approximately 35.8 PSU at 150–200 m, while SLIM reproduces it at a similar depth but with a lower magnitude of about 35.5–35.6 PSU. This structure is broadly consistent with the example hydrographic profile presented by Furnas & A. W. Mitchell (1996) for the central GBR region, which shows a subsurface salinity maximum slightly above 35.6 PSU at around 150–200 m depth. The corresponding temperature profile similarly shows warm surface waters of about 28 °C and a pronounced decrease to below 15 °C by 300 m. At a broader regional scale, the validated GBR4 simulations of Schiller et al. (2015) show a comparable vertical temperature and salinity structure in the Coral Sea adjacent to the GBR shelf break. Together with the Argo comparison, these regional references support the physical realism of the hydrographic structure reproduced by SLIM near the GBR shelf break.

Beyond the pointwise validation presented above, the simulated near-surface circulation pro-

vides a broader view of the regional flow organization reproduced by SLIM across the GBR and adjacent Coral Sea. In both periods, locally enhanced current speeds occur near the shelf break, with the associated flow predominantly directed southward, although their location and intensity differ between Jasper and Kirrily. This circulation pattern is in line with the prevailing summer flow regime reported for the GBR [C. Steinberg & Lawrey, 2018]. In the southern part of the domain, it is also consistent with Choukroun et al. (2010), who identified a strong southward EAC along the shelf edge south of approximately 19°S, with surface velocities exceeding 0.5 m s^{-1} . However, around 15°S, Choukroun et al. (2010) identified the formation of the northward-flowing NQC as the NVJ reaches the GBR shelf, a circulation pattern that is not clearly reproduced here. Closer to the coast, simulated currents predominantly follow the orientation of the coastline, although their direction varies spatially and between the two periods. This alongshore organization is consistent with Luick et al. (2007), whose tracer experiments showed that coastal waters predominantly move north and south along the GBR and tend to remain within a nearshore coastal band. These comparisons should nevertheless be interpreted with caution, as the reference patterns represent prevailing or multi-year circulation, whereas the present fields correspond to short periods influenced by tropical cyclones.

4.2 NPZD model

Overall, the retained NPZD configuration is numerically stable and reproduces several expected large-scale biogeochemical features, but its representation of the cyclone-driven coastal response is substantially weaker. In particular, the model markedly underestimates both the magnitude and spatial extent of the observed coastal phytoplankton response, as well as the magnitude and temporal evolution of post-cyclone coastal DIN enrichment, suggesting that an insufficient representation of bioavailable riverine nitrogen may be an important contributor to this mismatch. At the same time, broader variations in simulated DIN point to an additional oceanic contribution to nutrient availability, including cyclone-related vertical redistribution of subsurface nutrients and a distinct intrusion consistent with episodic upwelling.

Model behaviour and large-scale biogeochemical realism

The first set of diagnostics indicates that the retained NPZD configuration is numerically stable and reproduces several expected large-scale biogeochemical features. Domain-integrated inventories evolve smoothly throughout the simulation (Fig. 29), following an initial adjustment phase for phytoplankton and zooplankton. This stability is particularly relevant given the strong sensitivity of preliminary configurations to NPZD open-boundary conditions, which produced either unrealistic DIN accumulation or progressive depletion. The constant boundary concentrations retained here represent the best compromise found to achieve this relative stability; they likely introduce some bias into the domain-wide budget, but this is expected to have limited influence on the nearshore signal examined below. The model also maintains a coherent vertical DIN structure between September and April (Fig. 30), with nutrient-poor surface and shelf waters overlying DIN-rich deeper waters, separated by a transition around 150–200 m, consistent with observations from the GBR and ad-

jacent Coral Sea [Andrews & Furnas, 1986; Andrews & Gentien, 1982]. Detritus accumulates from its initial value of zero and is progressively redistributed toward depth (Fig. 31), consistent with the oceanic biological pump, through which plankton-derived particulate organic matter is transported from surface waters toward the deep ocean [Turner, 2015].

Post-cyclone phytoplankton and DIN response

The qualitative comparison between satellite-derived chlorophyll-*a* observations and simulated surface phytoplankton (Figs. 32 and 33) reveals both broad agreement and substantial differences in the post-cyclone response. The model reproduces the general shelf–offshore organization, with relatively higher phytoplankton concentrations over the continental shelf and lower concentrations offshore. However, following TC Jasper, satellite observations show a pronounced expansion of elevated chlorophyll-*a* from coastal waters toward the mid-shelf in the northern sector, whereas the simulated response is dominated by an offshore expansion of intermediate phytoplankton concentrations and only limited coastal enhancement. Following TC Kirrily, the observed coastal high-concentration band broadens mainly in the central and southern sectors, while the simulation again shows comparatively limited coastal changes and a broader offshore response. By 6 February, a distinct, spatially homogeneous band of intermediate phytoplankton also emerges across the northern shelf, a pattern with no clear counterpart in the satellite record. Overall, the model therefore captures the background shelf–offshore phytoplankton structure but does not reproduce the magnitude or spatial distribution of the observed post-cyclone coastal response. This discrepancy is particularly relevant because terrestrial runoff represents a major source of new nitrogen to GBR nearshore waters [Devlin & Brodie, 2005], and nutrients delivered by river plumes can rapidly support phytoplankton development when light conditions are favourable [Furnas et al., 2005]. The weak simulated coastal response therefore suggests that insufficient representation of riverine bioavailable nitrogen may be one contributor to the mismatch.

The comparison with in situ DIN measurements further supports this interpretation. Following TC Jasper, simulated coastal DIN concentrations remain substantially below the observations in the Barron and Herbert river sectors, and the pronounced post-cyclone peak observed on 19 December is not reproduced. At the location showing the largest mismatch, the observed surface DIN concentration was approximately two orders of magnitude higher than the simulated value. By 28 December, observed DIN had declined substantially, consistent with phytoplankton uptake, dilution and dispersion of the initial pulse, whereas the simulation failed to reproduce this decline and instead showed its highest DIN values among the sampled locations on that date. The model therefore underestimates not only the magnitude of the early post-Jasper coastal DIN pulse, but also its observed temporal evolution.

Taken together, these results suggest that insufficient representation of bioavailable nitrogen in coastal waters may contribute to the weak simulated post-cyclone response. Three aspects of the river forcing may contribute to this limitation and are discussed below.

Riverine nitrogen forcing

First, the river forcing includes only four rivers. Although these rivers responded strongly to TC Jasper and TC Kirrily, they represent only a small subset of the river systems draining into the GBR. Indeed, monitoring and modelling identify the Herbert, Burdekin, Fitzroy, Johnstone, Mulgrave–Russell, Tully and Haughton basins as the largest DIN exporters to the GBR [Prosser & Wilkinson, 2024], several of which are absent from the model. To provide an order of magnitude, GBR-wide annual DIN export has been estimated at approximately $16,000 \text{ t yr}^{-1}$, whereas the four basins represented here together account for approximately $3,600 \text{ t yr}^{-1}$, or about 23% of this estimate [Kroon et al., 2012]. Over the 20-day post-cyclone periods, the cumulative DIN loads introduced by these four rivers amounted to approximately 494 t following TC Jasper and 324 t following TC Kirrily — only 0.33% and 0.22% of the domain-integrated DIN inventory ($1.5 \times 10^5 \text{ t}$, Fig. 29). This comparison should nevertheless be interpreted cautiously, as most of the domain inventory lies in deeper, offshore waters, whereas riverine inputs are delivered directly into nutrient-poor coastal waters and may therefore be locally important.

Second, even for the four rivers included in the model, the imposed DIN loads may underestimate the actual loads reaching coastal waters. DIN export from GBR catchments is strongly influenced by the extent of intensive fertilized agriculture, particularly sugarcane cultivation [Lewis et al., 2024]. In the Wet Tropics, where three of the four rivers considered here are located, substantial areas of sugarcane occur on coastal floodplains downstream of gauging stations and may therefore not be fully captured in monitored loads [Prosser & Wilkinson, 2024]. Furthermore, part of the drainage from each basin reaches coastal waters through small waterways rather than through the main river channel [Waterhouse et al., 2012], and may therefore not be captured by monitoring at the main river channel. This potential underestimation may be amplified during floods, when over-bank flow can bypass gauging stations and stage–discharge relationships can underestimate flows above bank-full conditions [Wallace et al., 2012]. Because DIN load estimates depend directly on river discharge [Wallace et al., 2009], these flow biases translate into underestimated nitrogen loads. The proportion of annual flow potentially bypassed or under-read at gauges has been estimated at approximately 35%, 24%, 8% and 7% for the Burdekin, Herbert, Daintree and Barron rivers, respectively [Wallace et al., 2012]. Consistent with the potential magnitude of this effect, an analysis of 13 flood events in the Tully–Murray catchments estimated that annual average nitrogen export to the GBR lagoon was underestimated by 47% when flood discharge was not properly accounted for [Wallace et al., 2009]. Such systematic underestimation of gauge-based discharge and DIN loads could therefore contribute to the low coastal DIN concentrations simulated by the model relative to the observations.

Third, the model represents riverine nitrogen only as DIN, excluding particulate nitrogen (PN) and dissolved organic nitrogen (DON) pools that may subsequently contribute to bioavailable nitrogen during plume transport. These pools are substantial at the GBR scale, with annual exports estimated at approximately 15,000 t of DON and 52,000 t of PN, compared with about 16,000 t of DIN [Kroon et al., 2012]. Processes including microbial mineralisation of particulate nitrogen, microbial and photo-oxidative transformation of DON, and desorption of particle-bound ammonium from eroded sediment can progressively increase the bioavailable nitrogen pool, including through DIN

generation [Burford et al., 2024]. In two Burdekin flood plumes, Garzon-Garcia et al. (2021) estimated potential additional DIN generation equivalent to approximately 9–30% of the end-of-catchment DIN load. Importantly, this additional bioavailable nitrogen can be released over timescales ranging from hours for desorption to several days for organic-N mineralisation [Garzon-Garcia et al., 2018; Garzon-Garcia et al., 2021]. These timescales are particularly relevant in flood plumes, where high turbidity can initially limit phytoplankton growth until suspended sediments settle and light availability improves [Furnas, 2003]. A forcing based only on end-of-catchment DIN therefore neglects both the additional nitrogen regenerated during plume transport and its temporal evolution. This delayed nitrogen supply may be particularly relevant once plume turbidity decreases and light availability becomes less limiting, and could contribute to the coastal chlorophyll-*a* increases observed several days after cyclone landfall.

Despite these limitations in riverine nitrogen supply, substantial variations in DIN occurred across the simulation domain, including during both cyclone passages, and were associated with enhanced simulated phytoplankton concentrations. These broader-scale patterns suggest that non-riverine processes also contributed to nutrient availability, motivating an examination of oceanic nutrient supply and vertical redistribution.

Oceanic nutrient supply and vertical redistribution

During the passage of both TC Jasper and TC Kirrily, simulated surface DIN concentrations increased preferentially along and around the cyclone trajectories, extending well beyond the immediate vicinity of the river mouths and suggesting that riverine inputs alone are unlikely to explain this response (Appendix, Fig. 53). Such a pattern is consistent with the known effects of TCs on upper-ocean nutrient redistribution. Strong cyclone winds can weaken stratification, deepen the mixed layer and enhance vertical mixing and entrainment, thereby transporting nutrient-rich subsurface waters toward the surface [P. A. Thompson et al., 2023]. Following TC Kai-Tak in the South China Sea, Lin et al. (2003) documented mixed-layer deepening to approximately 90 m together with strong upwelling, allowing nutrient-rich waters from below a nutricline located at about 50 m to reach the euphotic zone. The simulated DIN response during Jasper and Kirrily therefore supports a contribution from cyclone-induced vertical redistribution of subsurface nutrients, although the present diagnostics do not allow the respective roles of vertical mixing, entrainment and upwelling to be quantified.

The Barron River–offshore transect indicates an episodic upward and landward displacement of DIN-rich offshore waters, which progressively extend onto shallower parts of the shelf between 25 and 31 January before retreating by 3 February (Fig. 35). This pattern closely resembles documented intrusive upwelling events in the central GBR, during which cool, nutrient-rich Coral Sea waters are uplifted over the continental slope and subsequently intrude onto the shelf near the bottom [Furnas & A. W. Mitchell, 1996; Benthuisen et al., 2016]. Such intrusions can propagate tens of kilometres landward over periods of approximately one to several weeks [Andrews & Furnas, 1986], and large events can import nitrate and phosphate stocks equivalent to two to six times those normally present in outer-shelf waters [Furnas & A. W. Mitchell, 1996]. The timing of the simu-

lated event is also consistent with independent observations from the northern GBR. Emslie et al. (2025) reported a sharp decrease in seawater temperature across the Cooktown–Lizard Island shelf (14–15.8°S) in late January and early February 2024, which they considered most likely to result from an intrusion of cooler upwelled water. Although these observations were obtained farther north and cannot directly validate the event simulated along the Barron transect, they indicate that an upwelling-related intrusion likely occurred in the adjacent sector during the same period. Such nutrient-rich intrusions can subsequently support phytoplankton development. Andrews & Gentien (1982) observed enhanced phytoplankton development associated with an intrusion extending landward from the shelf break, while Furnas & A. W. Mitchell (1996) identified these episodic nutrient inputs as an important driver of summer fluctuations in outer-shelf phytoplankton biomass and primary production. Consistent with these responses, several days after the simulated DIN intrusion reached its greatest landward extent, a relatively elevated and spatially homogeneous phytoplankton field developed across much of the northern shelf, including the Cooktown–Lizard Island sector (Fig. 33). This spatial and temporal correspondence is consistent with a phytoplankton response to enhanced oceanic nutrient supply associated with intrusive upwelling, although the contribution of this process cannot be isolated from other nutrient sources and biogeochemical processes.

4.3 Cumulative phytoplankton exposure

Interpretation of cumulative phytoplankton exposure

The cumulative phytoplankton exposure indicator was developed as a proxy for potential light attenuation associated with phytoplankton accumulation. As the model underestimates both coastal DIN enrichment and the associated phytoplankton response following each cyclone (Section 4.2), the resulting exposure estimates are not expected to accurately capture the true magnitude and spatial distribution of phytoplankton response. Direct validation against independent observations is also not possible, as available satellite and field measurements provide surface or spatially and temporally discrete observations rather than the vertically and temporally integrated quantity represented by the indicator. Moreover, cumulative phytoplankton exposure is not a standard indicator in the literature, further precluding comparison with published values.

Following TC Jasper, regional contrasts in cumulative exposure remained relatively limited, with Cape York South recording the highest mean exposure for both coral reefs and seagrass meadows (Fig. 36, Table 5). Consistent with Section 2.2, the indicator also fails to reproduce the pronounced coastal bloom identified from satellite chlorophyll-*a* observations after Jasper (Fig. 32); although these are surface observations, a bloom of that magnitude would be expected to raise cumulative exposure if adequately reproduced. The spatial influence of the contemporary riverine DIN forcing is not clearly identifiable from the absolute exposure field. Although the Daintree, Barron and Herbert rivers, all located within the Wet Tropics region, supplied almost the entire post-Jasper DIN load (493.7 of 494.4 t), exposure remained characterised primarily by a broad cross-shelf gradient rather than by a distinct response within this river-influenced region. This gradient is partly

influenced by bathymetry, as the vertically integrated, non-depth-normalised formulation of E tends to produce higher exposure values over deeper water columns. Following TC Kirrily, however, the regional pattern changed markedly, with a pronounced intensification in the northern part of the domain. Cape York South again recorded the highest mean exposure, with values more than doubling for coral reefs and increasing nearly fivefold for seagrass relative to the post-Jasper period, while exposure in several southern regions remained similar to or lower than after Jasper (Fig. 36, Table 5). This northern intensification contrasts with the distribution of contemporary riverine DIN inputs, which was dominated by the Burdekin River (227.24 of 323.86 t). The post-Kirrily exposure pattern therefore cannot be readily explained by the spatial distribution of the imposed river forcing alone. The northern high-exposure area is, however, consistent with other physical processes identified during the same period. It overlaps the Cooktown–Lizard Island sector, where Emslie et al. (2025) reported a probable intrusion of upwelled water in late January and early February 2024, and is also consistent with the upward and landward displacement of DIN-rich subsurface waters simulated during this period. In addition, simulated near-surface coastal currents converge around 15°S, where opposing alongshore flows meet (Fig. 28), potentially favouring the local retention of phytoplankton biomass and nutrient-rich waters. Together, intrusive nutrient supply and circulation-driven retention provide a plausible explanation for the pronounced northern exposure following Kirrily, although this spatial consistency does not validate the exposure indicator itself.

Contribution of riverine inputs to phytoplankton exposure

The comparison between the reference and no-river simulations isolates the influence of the prescribed riverine forcing within the model, but it should not be interpreted as the direct effect of riverine DIN delivered during each 20-day post-cyclone period. Because both simulations share the same hydrodynamics but have evolved with different river forcing since the beginning of the simulation, differences between the two simulations reflect both contemporary river inputs and the accumulated influence of earlier inputs on the coupled N–P–Z–D system. The mechanisms underlying these differences are difficult to isolate with the present diagnostics, and the interpretations proposed below should therefore be regarded as plausible explanations rather than demonstrated causal relationships.

Following TC Jasper, River–No-river differences remain relatively limited and spatially heterogeneous across most of the shelf (Fig. 37). This relatively small difference may partly reflect the limited cumulative influence of riverine forcing before Jasper, as the major DIN pulse occurred only during the post-cyclone period and the two simulations had therefore not yet developed strongly divergent biogeochemical states. Following TC Kirrily, in contrast, much larger positive differences develop in the northern part of the domain. Importantly, this is also the region where the reference simulation shows the pronounced increase in cumulative exposure described above, and it overlaps the sector affected by the late-January intrusion of DIN-rich subsurface waters. This spatial correspondence suggests that the stronger post-Kirrily divergence may partly reflect an interaction between the legacy of earlier riverine inputs and the subsequent oceanic nutrient supply. Although the Jasper pulse did not generate a pronounced simulated coastal bloom, it may have progressively altered the N–P–Z–D state of the reference simulation during the following weeks, while these in-

puts were absent from the no-river simulation. By the onset of Kirrily, pre-existing differences in phytoplankton biomass and the other biogeochemical compartments could therefore have modified the response to renewed nutrient availability. In particular, phytoplankton uptake scales with existing biomass through $U_P = \mu_{\max} f(T) f(N) f(I) P$, providing a mechanism by which moderate differences in phytoplankton biomass may be amplified when nutrient availability increases. The late-January intrusion may consequently have acted on already divergent biogeochemical states, contributing both to the larger River–No-river differences and to the pronounced exposure simulated in the northern sector. Overall, the experiment indicates that simulated exposure reflects not only contemporary riverine inputs but also their accumulated influence on the biogeochemical system and its interaction with subsequent oceanic nutrient supply. It should therefore be interpreted as a sensitivity of the present model to the four prescribed rivers rather than as a direct estimate of the actual contribution of riverine DIN to phytoplankton exposure in the GBR.

4.4 Limitations and perspectives

The modelling framework developed in this study involved several assumptions and methodological choices that influence the interpretation of the results. Six main limitations were identified and are discussed below, together with potential directions for future development.

The NPZD formulation used in this study provides a simplified representation of the biogeochemical system and was not specifically calibrated for the GBR region. Only one nitrogen pool is considered, while phytoplankton and zooplankton are represented by single aggregated compartments with uniform parameters. The base formulation and original parameter set were developed for the Gulf of Maine, although parameters were adapted using more locally relevant information where available [Tian et al., 2015; Zhang et al., 2016; Chen, 2015; Bureau of Meteorology, 2020]. Although this formulation reproduces several expected large-scale biogeochemical features, its ability to represent the magnitude, timing and spatial variability of post-cyclone responses remains limited. Future work could therefore include a more regionally constrained calibration of the biological parameters, together with a more detailed representation of key biogeochemical processes where sufficient observations are available.

A further limitation is the absence of suspended-sediment dynamics and associated turbidity. During flood events, high nutrient concentrations can coincide with strongly turbid conditions, which may delay phytoplankton development until suspended particles settle and light availability improves [Furnas, 2003]. Their omission may therefore affect the timing and magnitude of the simulated phytoplankton response, while also neglecting a direct source of light reduction for coral reefs and seagrass meadows. Future developments should incorporate suspended-sediment transport and its influence on underwater light availability.

Riverine nitrogen inputs are only partially represented in the present configuration. As discussed in Section 4.2, only four rivers are represented in the present configuration, and their inputs are prescribed exclusively as DIN from monitored loads, which may not capture the full amount

of bioavailable nitrogen reaching coastal waters during floods. This incomplete representation of riverine nitrogen inputs may therefore contribute to the underestimated coastal nutrient enrichment and weak phytoplankton response reproduced by the model. Future applications should broaden the represented river systems and nitrogen pools and improve estimates of event-scale nutrient delivery.

The formulation of the cumulative phytoplankton exposure indicator also presents several structural limitations. Because phytoplankton is integrated over the full water column without normalisation by depth, exposure partly depends on bathymetry, complicating comparisons between habitats occurring at different depths. Temporal integration also masks exposure dynamics, as short intense events and weaker persistent events may produce similar cumulative values. Finally, the indicator represents phytoplankton accumulation rather than actual light reduction or ecological stress. Future developments should therefore better account for habitat depth and exposure intensity and duration, with the longer-term objective of developing a damage-oriented indicator linking reduced light availability to potential ecological impacts on coral reefs and seagrass meadows.

The characterisation of oceanic nutrient supply also remains limited. Several upward displacements of DIN-rich subsurface waters were identified, but the late-January event was the clearest and the only one independently supported by observations of a probable upwelled-water intrusion in an adjacent sector [Emslie et al., 2025]. The smaller events are more difficult to attribute and may involve different combinations of mixing, advection and upwelling. Future work should combine nutrient and hydrographic diagnostics with cross-shelf fluxes or passive tracers, together with more targeted hydrodynamic assessment at the shelf break, to better identify and quantify these processes.

Finally, the constant open-ocean boundary conditions used for the NPZD tracers represent an additional source of uncertainty. While this configuration provided the most numerically stable compromise, it neglects seasonal and event-scale variability in offshore biogeochemical conditions and may therefore influence shelf-break nutrient exchanges. Future implementations could reduce this uncertainty by prescribing time-varying boundary fields, for example from the eReefs GBR4 model output.

5 | Conclusion

TCs are a major and intensifying source of disturbance to coastal ecosystems, particularly in the GBR, one of the most diverse marine ecosystems on Earth, which is frequently affected by these events. During the 2023–24 TC season, TC Jasper and TC Kirrily crossed the GBR catchment within less than two months, bringing extreme rainfall and elevated river discharge to coastal waters. This study aimed to assess the impact of TC-associated nitrogen inputs on phytoplankton dynamics and coastal ecosystem exposure in the GBR using a coupled hydrodynamic–biogeochemical modelling approach.

The first step consisted in simulating the hydrodynamics of the GBR region using the 3D ocean model SLIM3D, evaluated against IMOS mooring and Argo float observations. The resulting hydrodynamic model was then coupled to a simplified NPZD biogeochemical model representing riverine DIN inputs from four rivers strongly affected by TC Jasper and TC Kirrily. This coupled model simulated the transport of nitrogen across the shelf and its assimilation by phytoplankton, with performance evaluated against satellite-derived chlorophyll-a and in situ DIN observations. Finally, simulated phytoplankton concentrations were used to quantify cumulative exposure of coral reefs and seagrass meadows over the post-cyclone periods. A second simulation without riverine DIN inputs was used to assess the contribution of river forcing to this exposure.

Together, these steps allowed us to address the overarching question of this thesis: **How do TC-associated nitrogen inputs and hydrodynamic processes shape the spatial and temporal dynamics of phytoplankton in the GBR during and following TC Jasper and TC Kirrily, and what potential exposure do these dynamics create for coral reefs and seagrass meadows?**

Despite the NPZD model's stable and internally consistent large-scale behaviour, its performance was more limited in reproducing the post-cyclone coastal response. The observed post-Jasper DIN enrichment and rapid coastal phytoplankton increase visible in satellite observations were substantially underestimated by the model. This likely reflects, at least in part, the incomplete representation of riverine nitrogen forcing, as only four rivers are represented, monitored loads likely underestimate true event-scale export, and nitrogen inputs are limited to the DIN pool. Nevertheless, contrasting exposure patterns emerged between the two cyclone periods: regional differences remained relatively limited after TC Jasper, whereas exposure increased sharply in the northern study area after TC Kirrily, particularly in Cape York South. This northern intensification coincided

with an episodic upward and landward intrusion of DIN-rich subsurface waters along the Barron transect, consistent in timing with independent observations of an upwelling-related intrusion near Cooktown–Lizard Island. Simulated near-surface currents also converged around 15°S, potentially favouring local retention of nutrients and phytoplankton biomass. Together, these results suggest that phytoplankton exposure is not governed by riverine nitrogen delivery alone, but emerges from its interaction with oceanic nutrient supply and hydrodynamic redistribution.

These dynamics produced the highest simulated ecosystem exposure in Cape York South following TC Kirrily, where mean cumulative exposure reached 123.3 mmol N m⁻² d for coral reefs and 103.7 mmol N m⁻² d for seagrass meadows, compared with 47.7 and 21.4 mmol N m⁻² d, respectively, following Jasper. These values should, however, be interpreted as indicators of potential phytoplankton-related pressure rather than direct ecological impacts, as the metric does not directly quantify underwater light attenuation or ecosystem response.

The sensitivity analysis provided further insight into the contribution of riverine nitrogen inputs. River–No-river differences remained limited and heterogeneous after Jasper but became substantially larger and predominantly positive in the northern sector after Kirrily. Because the simulations had evolved with different river forcing since their beginning, these differences include both contemporary inputs and the accumulated influence of earlier riverine inputs on the biogeochemical state. The stronger post-Kirrily response may therefore partly reflect an interaction between this biogeochemical legacy and the subsequent oceanic intrusion. Overall, TC-driven phytoplankton exposure in the GBR appears to emerge from the combined influence of riverine nitrogen inputs, their biogeochemical legacy, and hydrodynamic processes controlling nutrient supply, transport, and retention.

6 | References

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7 | Appendix - Hydrodynamic and biogeochemical validation

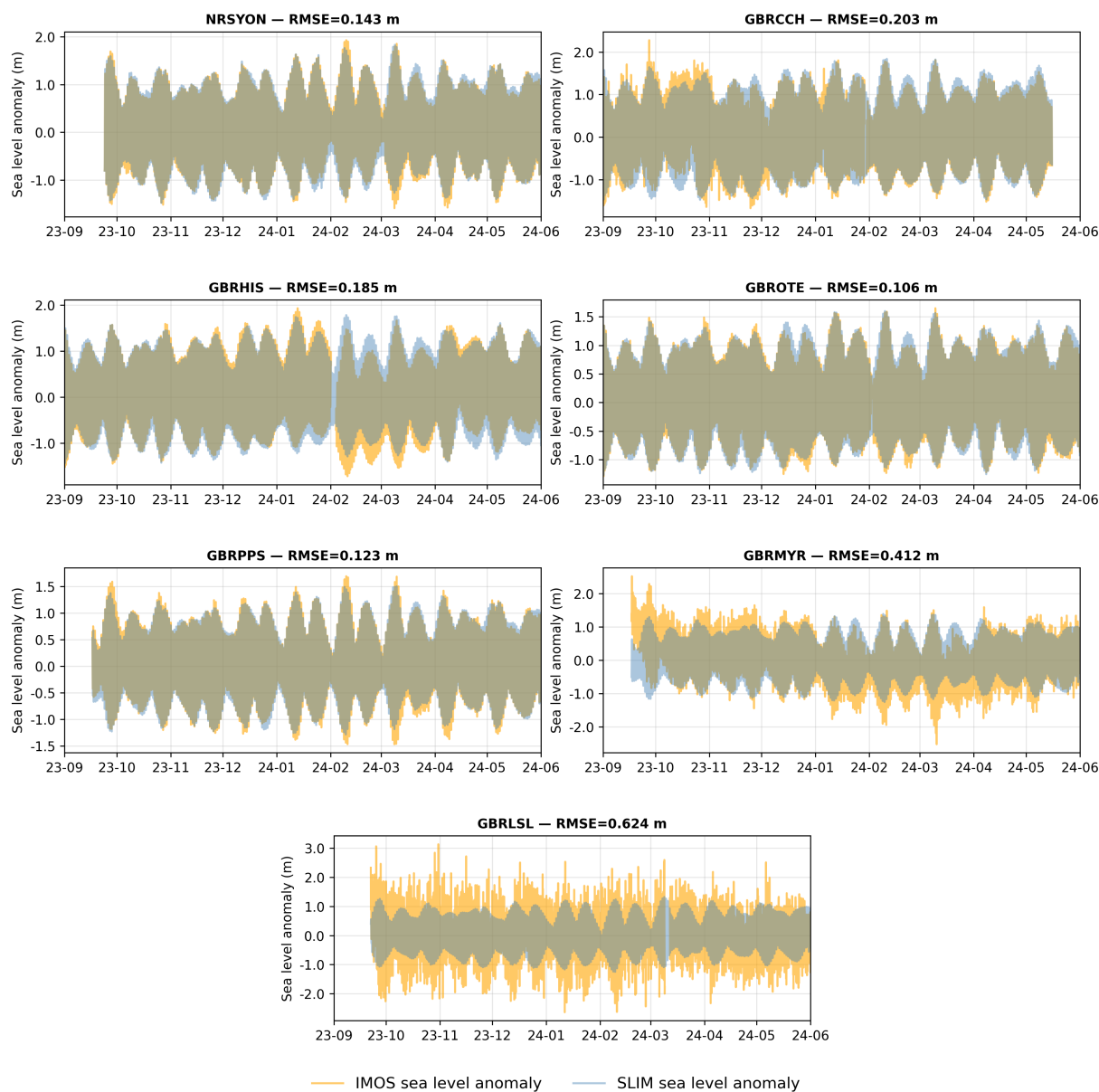


Figure 38: Comparison of sea level anomaly between SLIM and IMOS at the seven validation stations.

Table 6: RMSE and NRMSE values for sea level anomaly validation at the seven IMOS stations.

Station	RMSE (m)	NRMSE (%)
NRSYON	0.143	4.0
GBRCCH	0.203	5.1
GBRHIS	0.185	5.1
GBROTE	0.106	3.6
GBRPPS	0.123	3.9
GBRMYR	0.412	8.2
GBRLSL	0.624	10.8

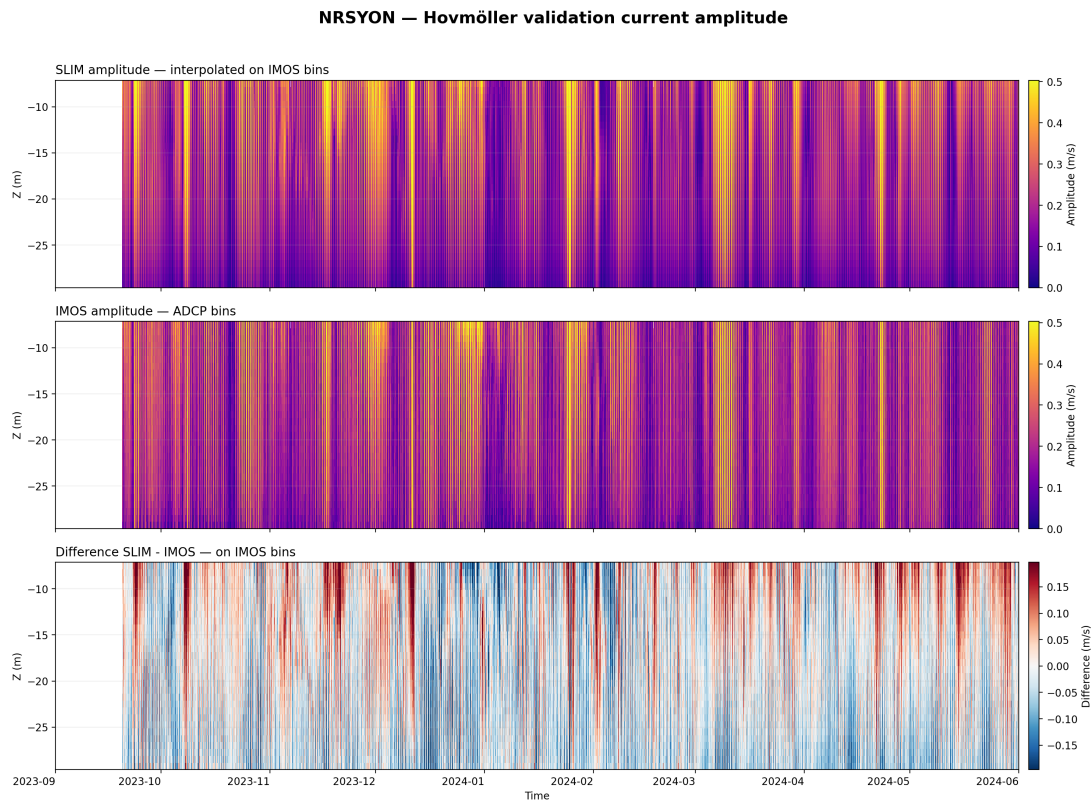


Figure 39: Hovmöller comparison of current speed amplitude between SLIM and IMOS at NRSYON.

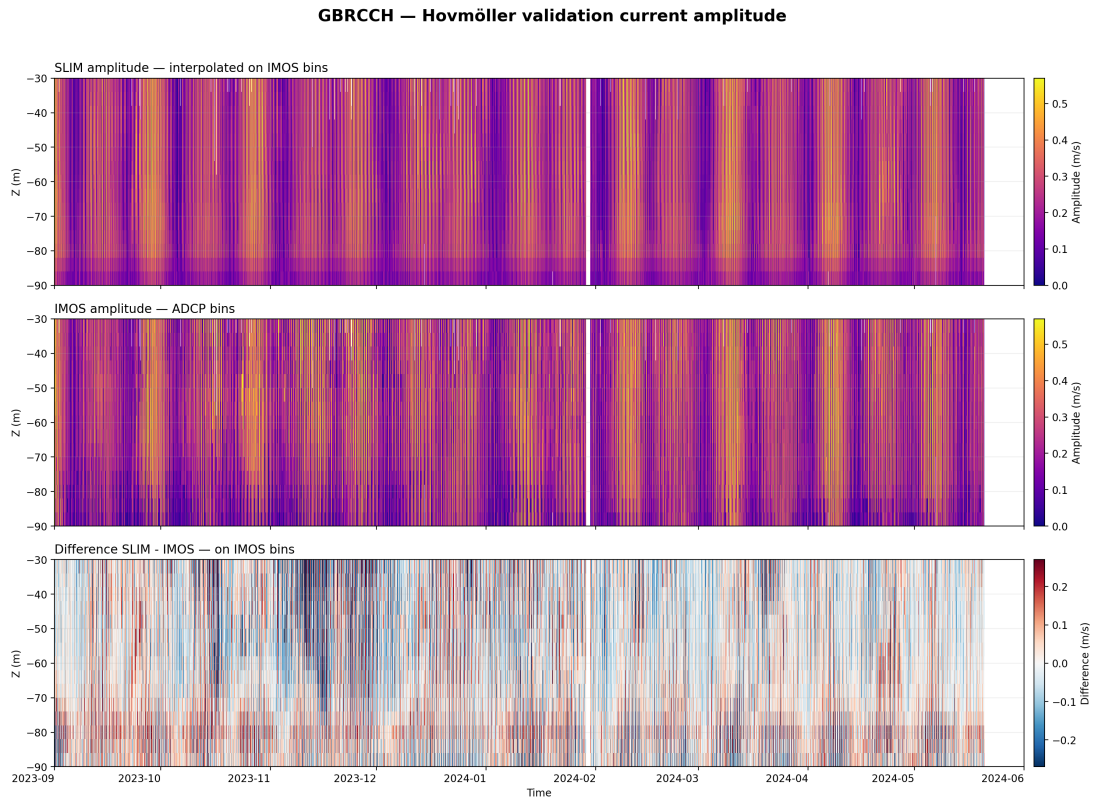


Figure 40: Hovmöller comparison of current speed amplitude between SLIM and IMOS at GBRCCH.

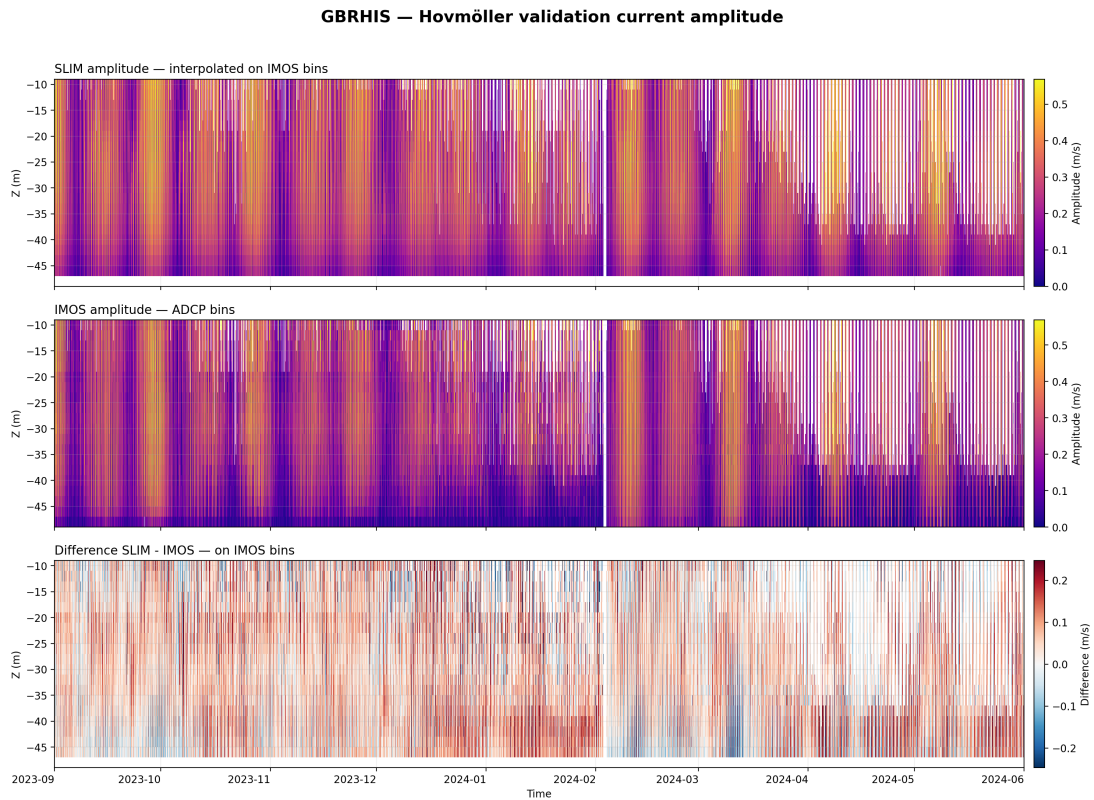


Figure 41: Hovmöller comparison of current speed amplitude between SLIM and IMOS at GBRHIS.

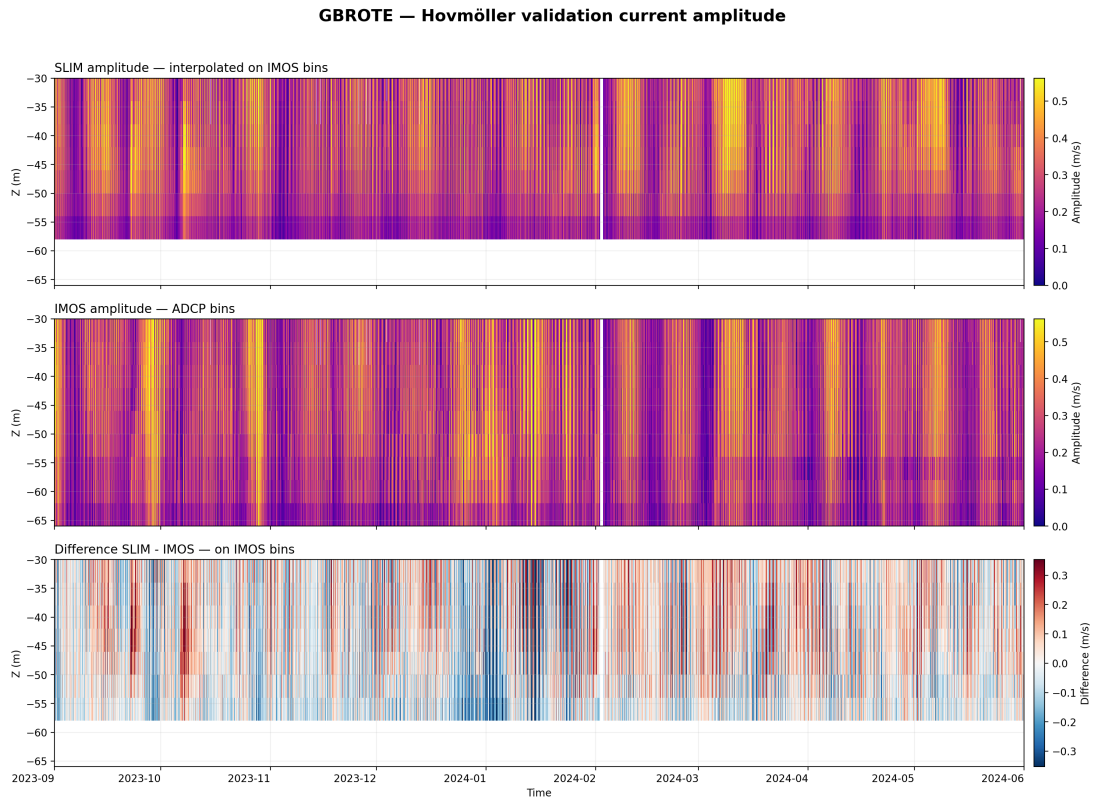


Figure 42: Hovmöller comparison of current speed amplitude between SLIM and IMOS at GBROTE.

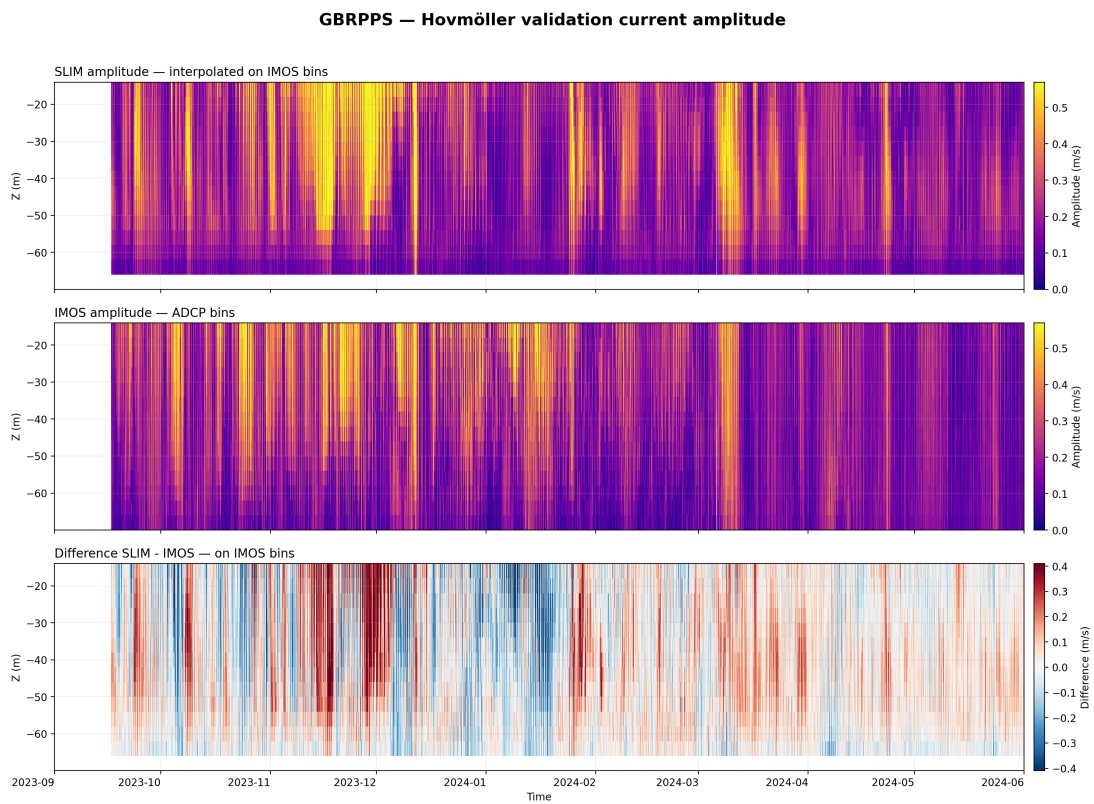


Figure 43: Hovmöller comparison of current speed amplitude between SLIM and IMOS at GBRPPS.

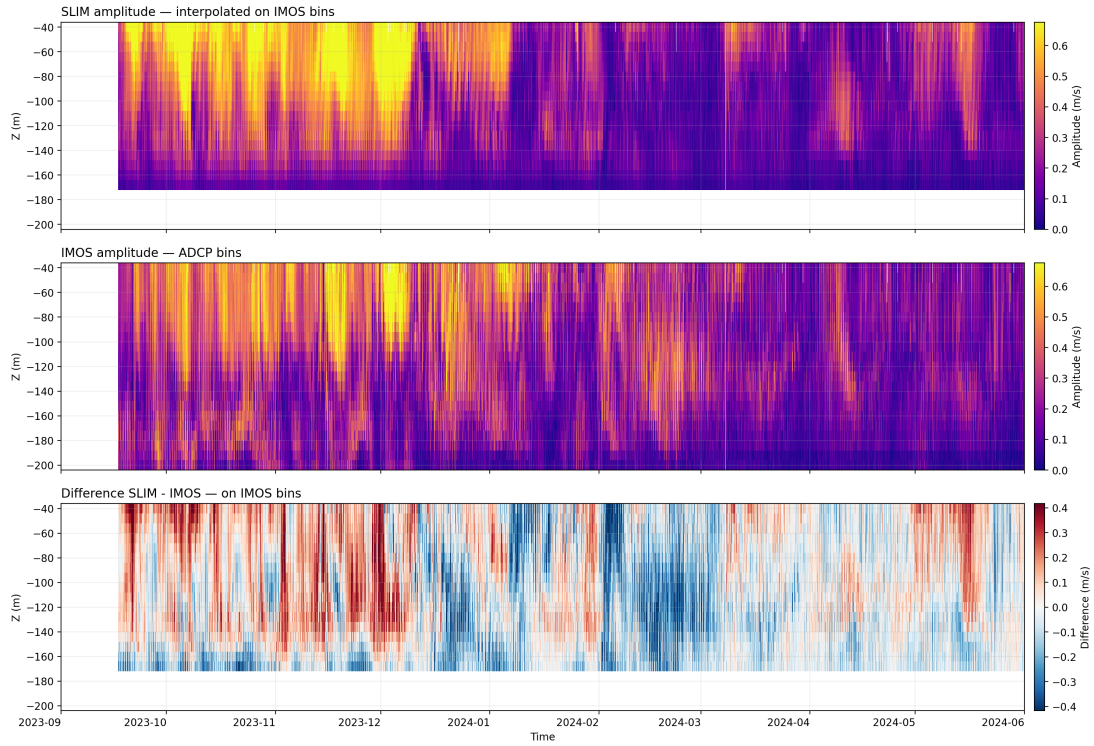
GBRMYR — Hovmöller validation current amplitude

Figure 44: Hovmöller comparison of current speed amplitude between SLIM and IMOS at GBRMYR.

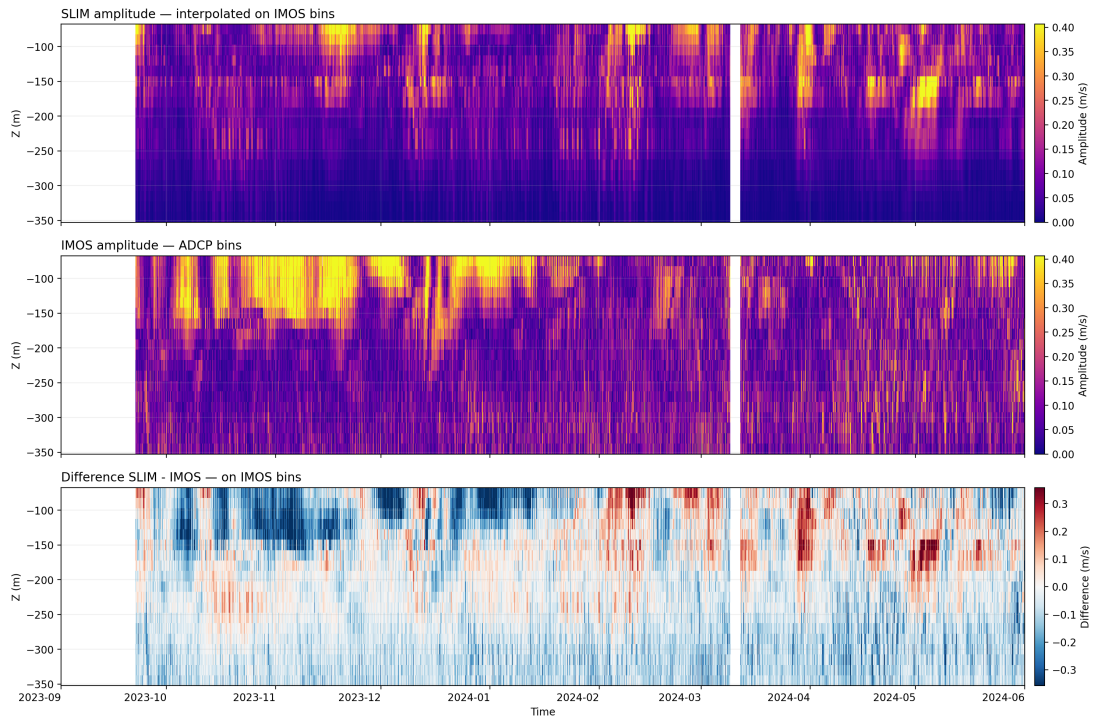
GBRSL — Hovmöller validation current amplitude
RMSE=0.1316 m/s | NRMSE=15.5% | N=341905


Figure 45: Hovmöller comparison of current speed amplitude between SLIM and IMOS at GBRSL.

Table 7: RMSE and NRMSE of current speed amplitude at the three validation depths for the seven IMOS stations.

Metric	Depth	NRSYON	GBRCCH	GBRHIS	GBROTE	GBRPPS	GBRMYR	GBRLSL
3*RMSE (m/s)	z_{min}	0.117	0.134	0.158	0.155	0.213	0.192	0.230
	z_{half}	0.086	0.122	0.109	0.154	0.171	0.191	0.129
	$z_{station}$	0.081	0.123	0.096	0.151	–	–	0.122
3*NRMSE (%)	z_{min}	8.5	14.3	9.1	15.0	22.0	14.1	25.0
	z_{half}	6.9	15.3	9.0	14.3	19.9	20.5	21.0
	$z_{station}$	8.1	15.3	17.0	16.2	–	–	23.1

Current direction roses — $z_{\min}$

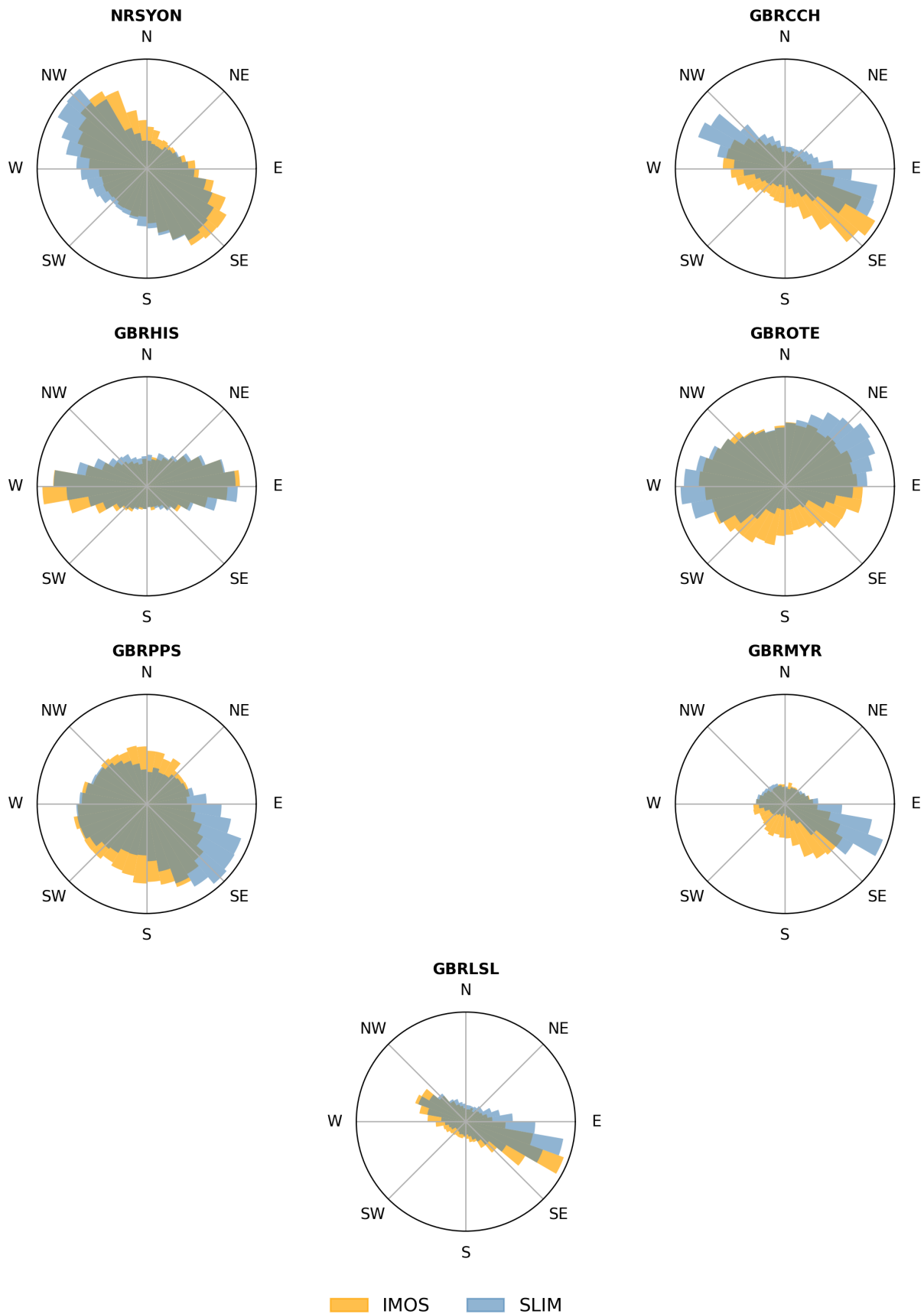


Figure 46: Current direction distributions from SLIM and IMOS at $z_{\min}$ for the seven validation stations.

Current direction roses — z_{half}

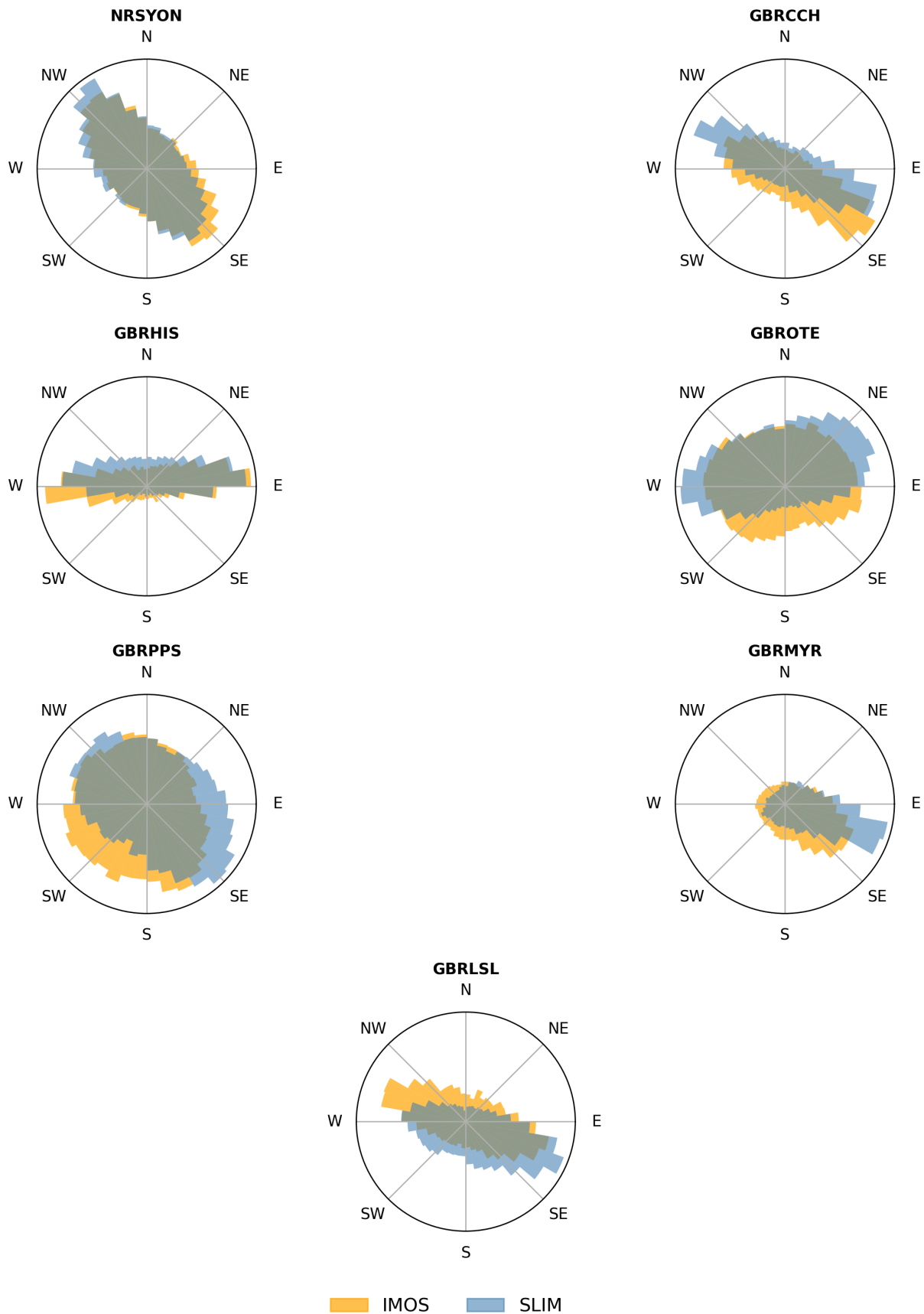


Figure 47: Current direction distributions from SLIM and IMOS at z_{half} for the seven validation stations.

Current direction roses — zstation

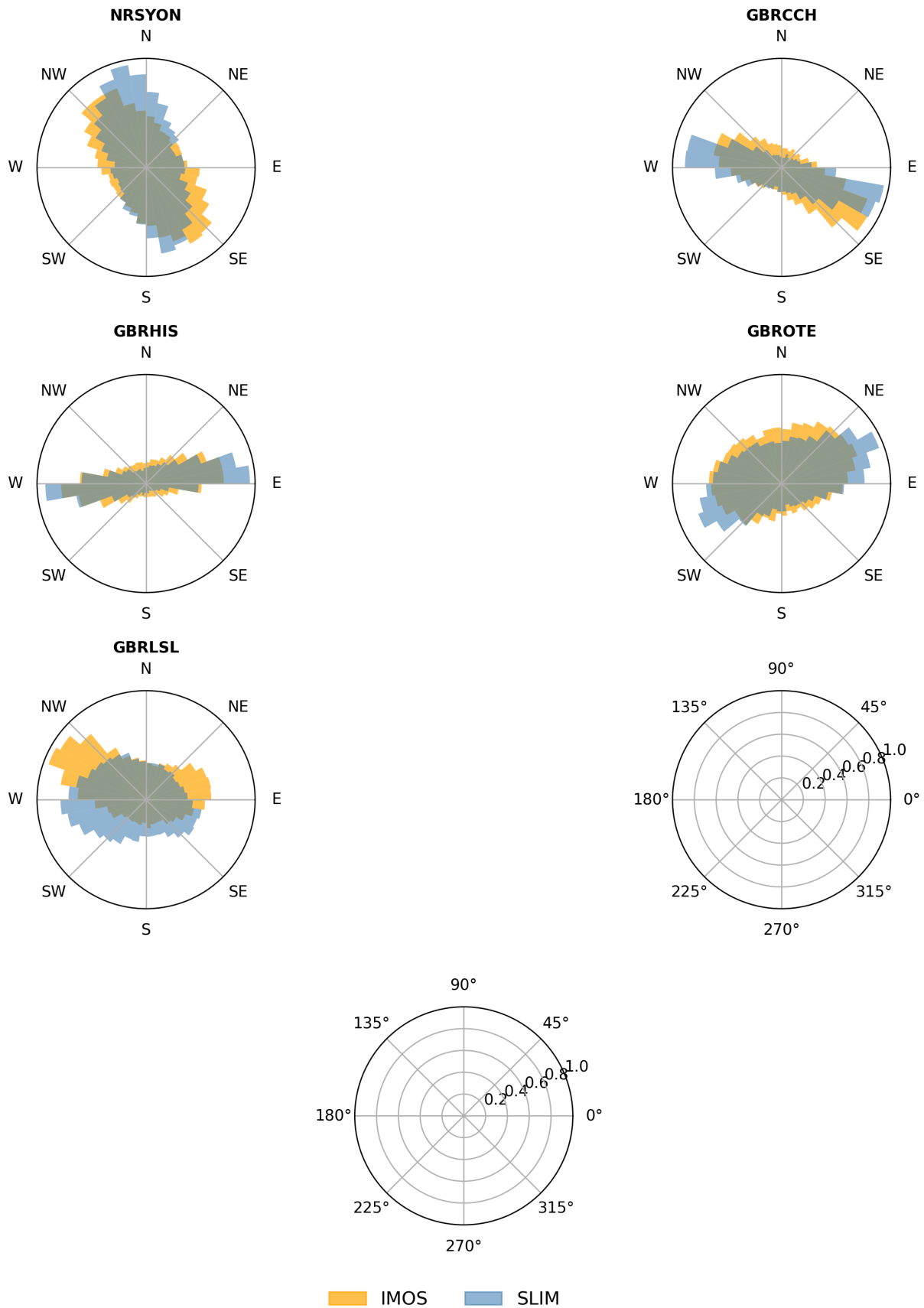


Figure 48: Current direction distributions from SLIM and IMOS at z_{station} for the available validation stations.

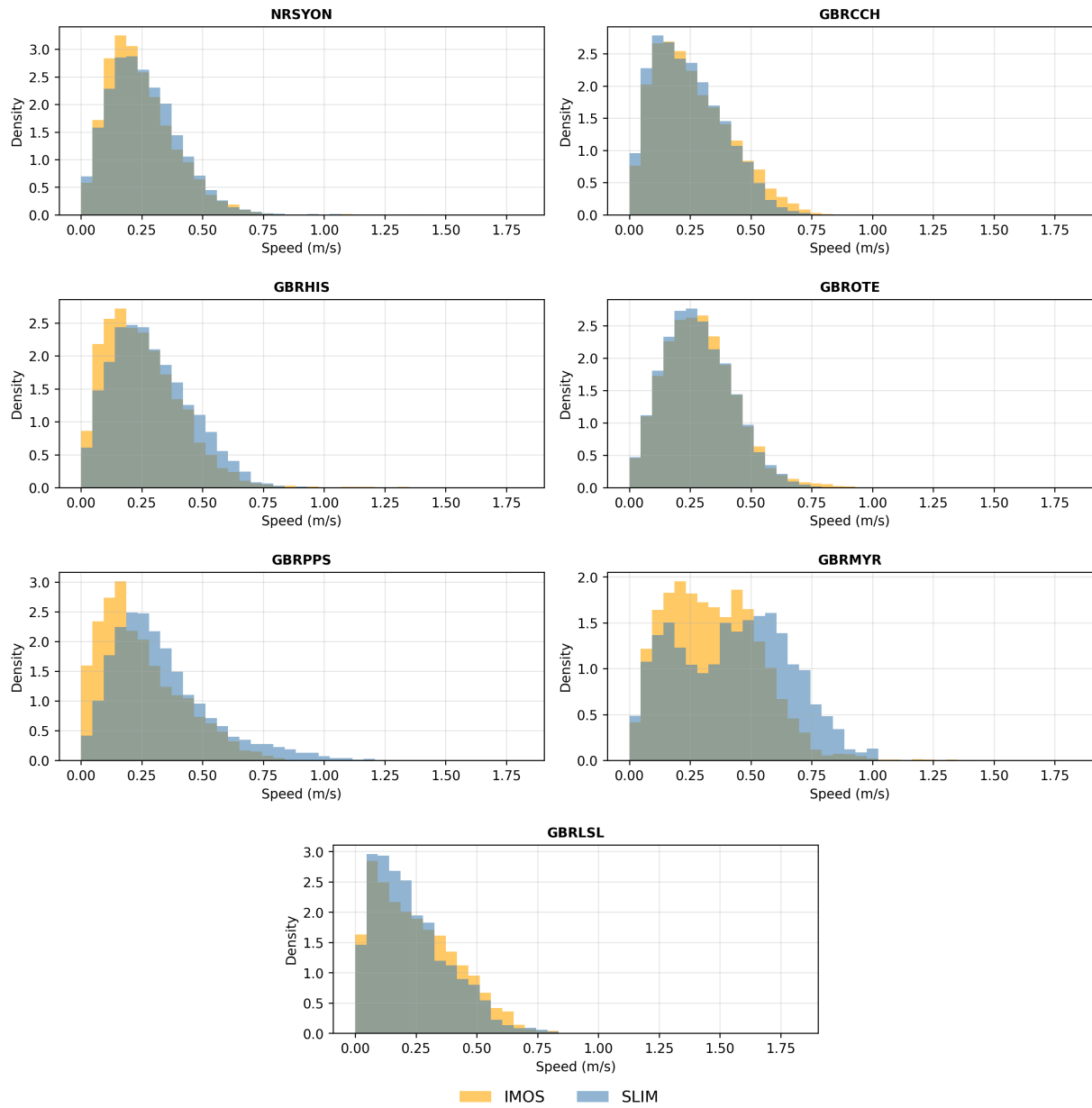
Current speed distributions — $z_{\min}$ 

Figure 49: Current speed distributions from SLIM and IMOS at $z_{\min}$ for the seven validation stations.

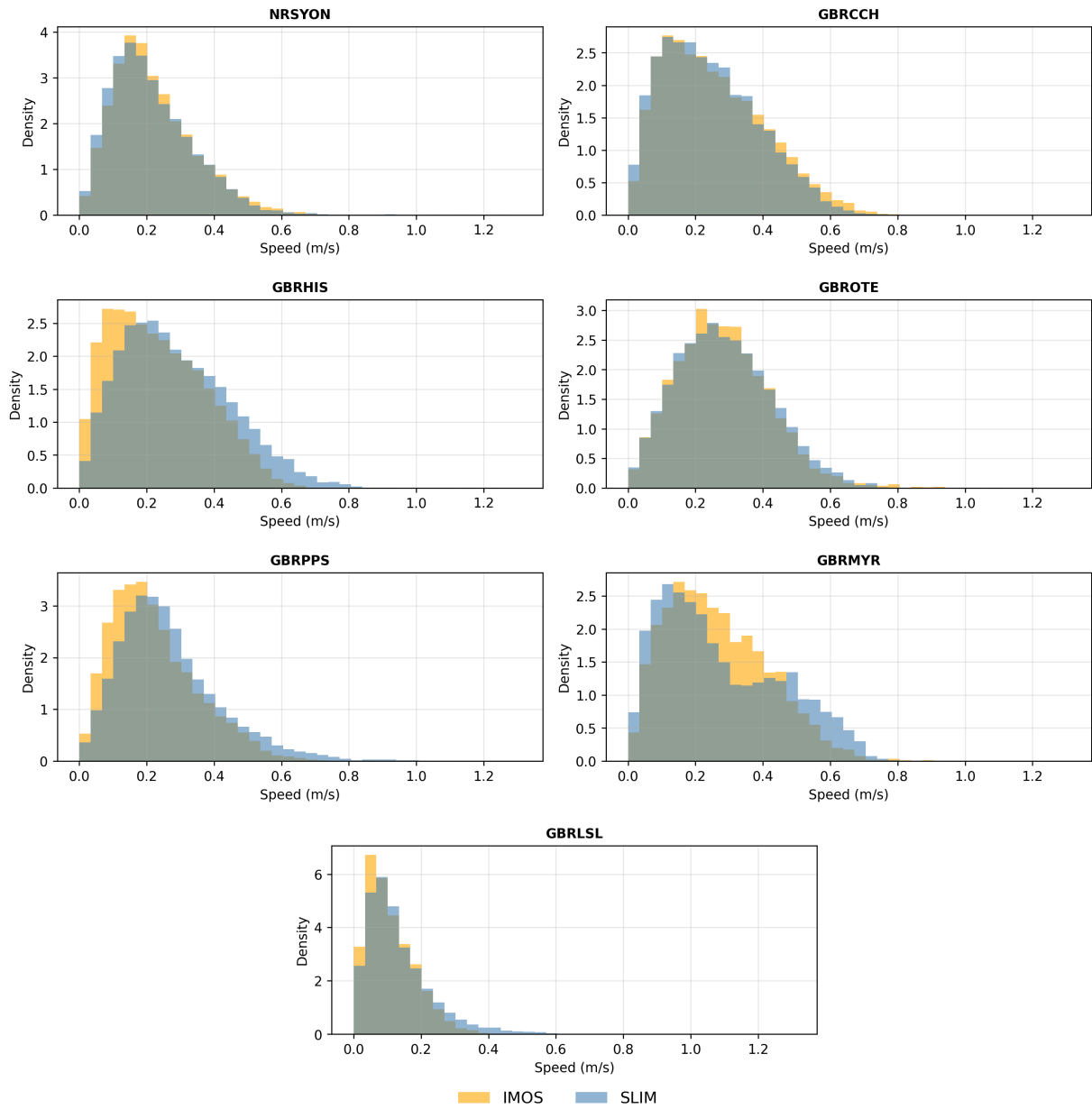
Current speed distributions — z_{half} 

Figure 50: Current speed distributions from SLIM and IMOS at z_{half} for the seven validation stations.

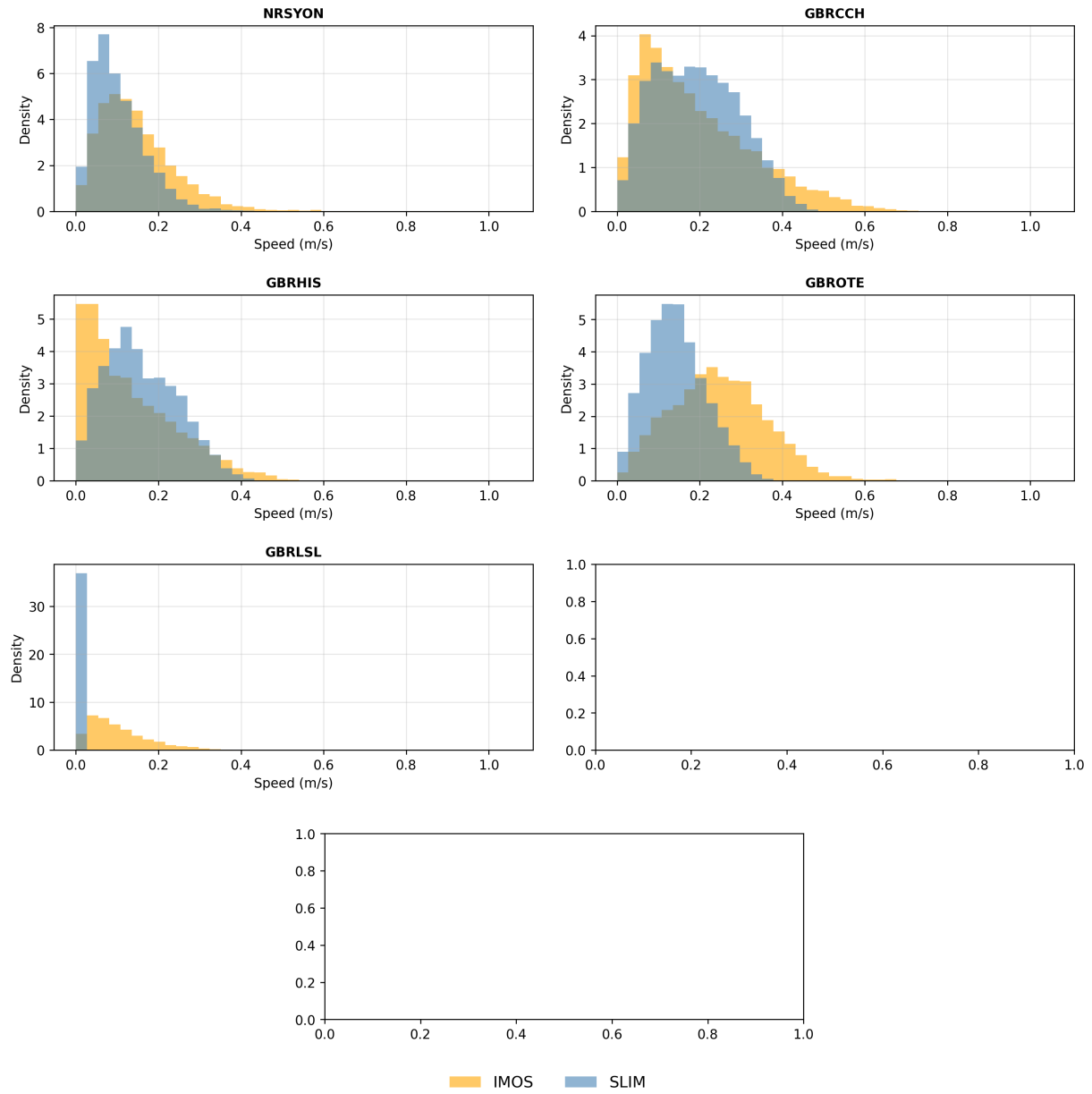
Current speed distributions — zstation

Figure 51: Current speed distributions from SLIM and IMOS at z_{station} for the available validation stations.

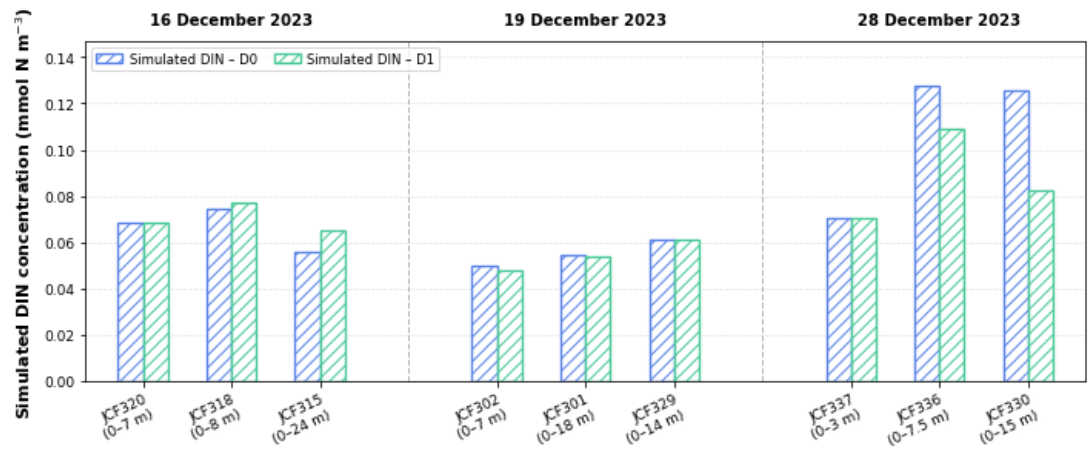


Figure 52: Zoomed illustration of the extracted values used for the JCU DIN validation.

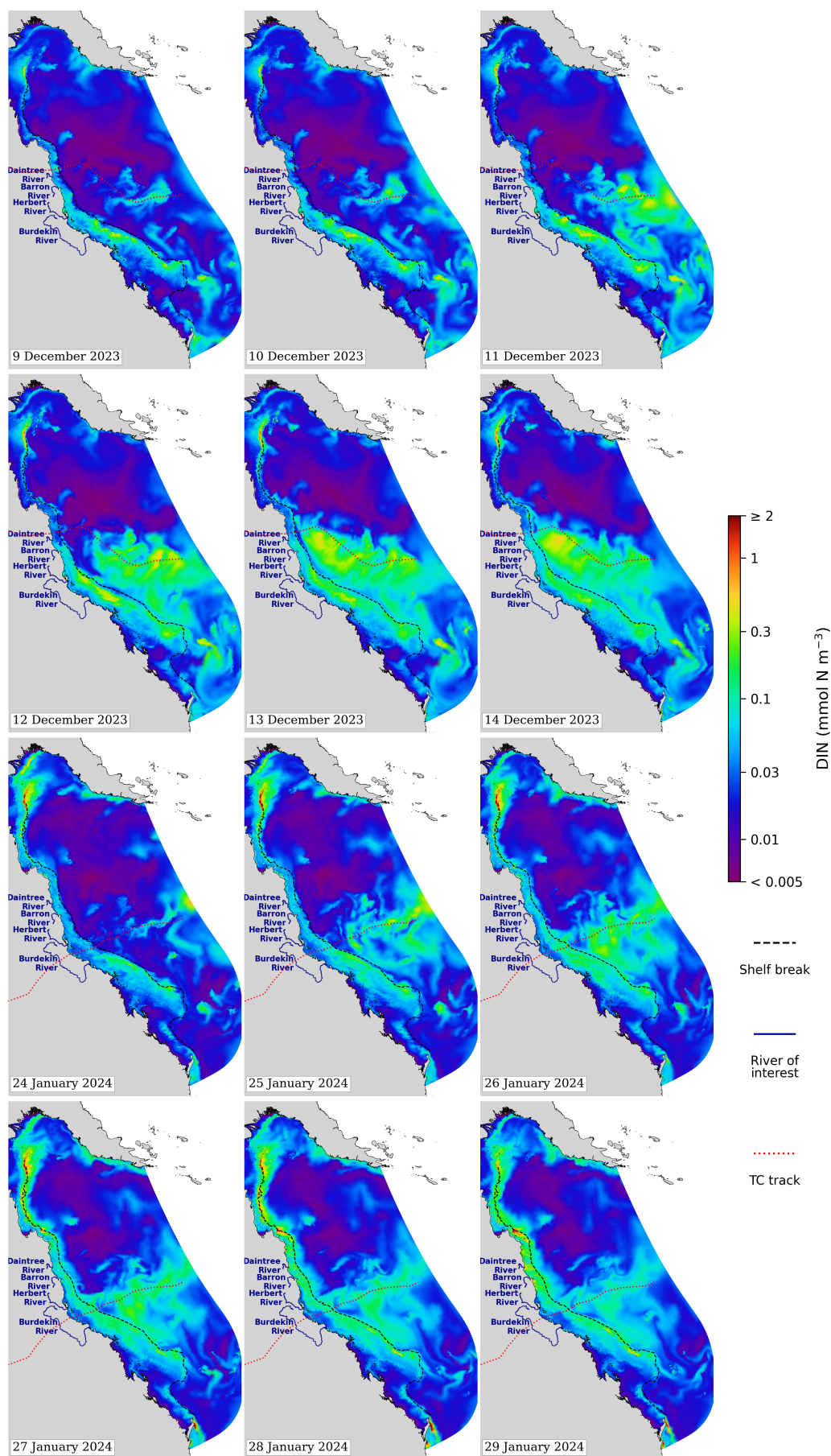


Figure 53: Temporal evolution of simulated surface DIN concentrations during the passages of TC Jasper (9–14 December 2023) and TC Kirrily (24–29 January 2024). The dotted red lines indicate the cyclone tracks.

Seamless water quality modelling under extreme weather conditions: Linking land nitrogen outputs to coastal ecosystem health

Charles Lavry

Abstract

Tropical cyclones can generate major hydrological and biogeochemical disturbances in coastal ecosystems through extreme rainfall, river discharge, and associated nutrient inputs. These processes are particularly relevant for the Great Barrier Reef (GBR), where coral reefs and seagrass meadows are exposed to variations in coastal water quality. This study investigates how TC-associated nitrogen inputs and hydrodynamic processes shaped phytoplankton dynamics during and following TC Jasper and TC Kirrily in the 2023–24 wet season, and the resulting potential ecosystem exposure. A three-dimensional hydrodynamic model of the GBR region was developed using SLIM3D and coupled to a simplified NPZD biogeochemical model incorporating dissolved inorganic nitrogen inputs from four rivers affected by the two cyclones. Model performance was evaluated against hydrodynamic observations, satellite-derived chlorophyll-a, and in situ DIN measurements. Simulated phytoplankton concentrations were then used to quantify cumulative exposure of coral reefs and seagrass meadows, complemented by a River–No-river sensitivity experiment. The hydrodynamic model reproduced the main regional circulation and hydrographic structures, whereas the biogeochemical model underestimated the observed post-Jasper coastal DIN enrichment and phytoplankton response. Nevertheless, contrasting exposure patterns emerged between the two cyclone periods. Exposure increased strongly in the northern GBR following Kirrily, with mean values in Cape York South reaching $123.3 \text{ mmol N m}^{-2} \text{ d}$ for coral reefs and $103.7 \text{ mmol N m}^{-2} \text{ d}$ for seagrass meadows, compared with 47.7 and $21.4 \text{ mmol N m}^{-2} \text{ d}$, respectively, following Jasper. This intensification coincided with an episodic intrusion of DIN-rich subsurface waters and hydrodynamic conditions potentially favouring local retention. Overall, cyclone-driven phytoplankton exposure in the GBR appears to result from the interaction of riverine nitrogen inputs, the pre-existing biogeochemical state, and hydrodynamic processes controlling nutrient supply, transport, and retention.

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