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# Oxygen, Carbon and pH Variability in the Indian Ocean

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## Abstract

This chapter provides an overview of the physical and biogeochemical factors that control spatial and temporal patterns of oxygen minimum zones (OMZs), carbon and pH variability in the Indian Ocean. Oxygen concentrations decline to nearly zero in Arabian Sea intermediate water with profound biogeochemical impacts on the nitrogen cycle as a result of denitrification. These impacts can hardly be observed in the Bay of Bengal where oxygen concentrations are poised just above the threshold below which denitrification becomes significant. Hypoxic/anoxic conditions in the open ocean waters of the northern Indian Ocean have not dramatically changed over past decades, but evidence is now emerging that oxygen concentrations are starting to decline, with significant biogeochemical and ecological impacts. The Indian Ocean accounts for ~1/5 of the global oceanic uptake of atmospheric CO<sub>2</sub>, with the Arabian Sea as a source of CO<sub>2</sub> to the atmosphere and the southern subtropical gyre as a CO<sub>2</sub> sink. Net CO<sub>2</sub> flux in the Bay of Bengal is uncertain due to sparse sampling. Surface pH values in the Indian Ocean are anomalously low and projected to decline further with negative impacts on calcifying organisms. Dissolved organic carbon (DOC) concentrations in the Indian Ocean tend to be high in near-surface (sub)tropical waters where autotrophic production of DOC exceeds heterotrophic consumption and vertical stability of the water column favors accumulation. In contrast, the highest particulate organic carbon (POC) concentrations in the Indian Ocean are observed in the northwestern part of the basin and the lowest in the southern subtropical gyre, reflecting primary production patterns. POC export flux patterns in the Indian Ocean are similar to the patterns in POC concentration, though carbon flux is also strongly influenced by lithogenic matter content in river-influenced regions like the Bay of Bengal. Observational and modeling research should target improved understanding of northern Indian Ocean OMZ and carbon system variability as such is needed to predict the impacts of anthropogenic influence and global warming on Indian Ocean biogeochemistry and ecosystems.

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# 1. Introduction

## 1.1 The Northern Indian Ocean oxygen minimum zones (OMZs)

Open ocean Oxygen Minimum Zones (OMZs) occur below the mixed layer of the ocean and are characterized by oxygen concentrations that are low in comparison to surface and deep waters (Dietrich, 1936; Seiwel, 1937; Sverdrup, 1938). Even though mid-water oxygen minimum zones are common features in the ocean, they are typically classified as OMZs only when the lowest minimum oxygen concentrations in the vertical profile drop below certain thresholds (Paulmier et al., 2009). However, due to varying oxygen tolerances of marine organisms, there are multiple definitions for threshold concentrations. Table 1 lays out common definitions, thresholds and impacts of varying definitions of oxygen limitations. A threshold of about 20  $\mu\text{mol/kg}$  is often used to define OMZs because concentrations below this have profound implications for microbes and marine biogeochemical cycles. These oxygen levels are therefore referred to as “microbial hypoxia” (Table 1). The total volume of waters characterized by microbial hypoxia in the global ocean is approximately  $15 \times 10^{15} \text{ m}^3$ , of which 21% ( $3.13 \times 10^{15} \text{ m}^3$ ) is located in the northern Indian Ocean (Acharya et al., 2016; Garcia et al., 2010). Even though the Indian Ocean OMZ constitutes only 0.23% of the ocean’s volume ( $1355 \times 10^{15} \text{ m}^3$ ) its influence on the cycles of nitrogen, carbon and associated elements is globally significant.

The thickest OMZ in the world is found in the Arabian Sea (e.g., Morrison et al., 1999) where functional anoxia (concentrations  $< .05 \mu\text{mol/kg}$ , Table 1) occurs in intermediate water (~200-800 m), with profound impact on the nitrogen cycle as a result of denitrification. Very low oxygen concentrations are also found in intermediate water in the Bay of Bengal, but there are important physical and biogeochemical differences between the Arabian Sea and the Bay of Bengal, which have, so far, prevented the development of persistent functional anoxia in the Bay of Bengal OMZ (Figures 1 and 2). Hence in contrast to the Bay of Bengal, the Arabian Sea OMZ is a globally important zone of denitrification (Naqvi et al., 2005), where  $\text{NO}_3^-$  and  $\text{NO}_2^-$  are converted to  $\text{N}_2\text{O}$  and  $\text{N}_2$  gas, which is then released to the atmosphere. This process of denitrification removes nitrogen containing compounds from the ocean (Figures 1 and 2) and generates  $\text{N}_2\text{O}$ , a prominent greenhouse gas (Ramaswamy et al., 2001, Bange et al. 2001). The Arabian Sea OMZ is most intense in the eastern part of the basin (Figures 1 and 2), with the water column contributing ~10 - 20% of global open-ocean mid-water column denitrification (Codispoti et al., 2001; Rixen et al., 2020; Anju et al., 2022).

Questions remain regarding the relative roles of biological oxygen demand derived from surface organic matter export, versus circulation and ventilation timescales, in maintaining subtle differences in the deep oxygen fields in the Arabian Sea and the Bay of Bengal (Valsala et al., 2009; McCreary et al., 2013; Bopp et al., 2017; Rixen et al., 2020). Recent observational and modeling studies in the Indian Ocean suggest that the OMZs are expanding in response to global warming (Lachkar et al., 2020; Rixen et al., 2020). Perhaps the low nitrogen to phosphorus ratios in the western Bay of Bengal are a harbinger of what is to come (Figure 1). This OMZ expansion is consistent with global modeling studies (Stramma et al., 2008; 2010; Doney, 2010; Figure 3) but uncertainties in global model predictions are large (McCreary et al., 2013; Bopp et

1 al., 2017; Rixen et al., 2020; Kwiatkowski et al., 2020; Schmidt et al., 2020; Schmidt et al.,  
2 2021).

3  
4 Given the importance of the Indian Ocean OMZs in the global carbon and nitrogen cycles,  
5 including the production of radiatively active greenhouse gas, it is essential to understand the  
6 biogeochemical variability associated with these regions, their rates of change and the potential  
7 implications for global warming.

## 8 9 **1.2 Role of the Indian Ocean in the global carbon cycle**

10  
11 The Indian Ocean plays an important role in the global carbon cycle, yet it remains one of the  
12 most poorly sampled ocean regions with respect to inorganic and organic carbon pools and air-  
13 sea carbon fluxes (Figures 4, 5 and 6). Estimates suggest that the Indian Ocean accounts for a  
14 significant fraction of the global oceanic uptake of atmospheric CO<sub>2</sub> (Takahashi et al., 2002;  
15 Sarma et al., 2013; Valsala and Maksyutov, 2013; Sreeush et al., 2018; 2020; de Verneil et al.,  
16 2021). This uptake of anthropogenic CO<sub>2</sub> drives a decrease in pH, and indeed, surface pH values  
17 in the Indian Ocean are among the lowest of the major ocean basins (Feely et al., 2009; Sreeush  
18 et al., 2019). Projected increases in oceanic CO<sub>2</sub> concentrations will lead to further acidification  
19 of the Indian Ocean over the coming decades, with potentially severe negative impacts on coral  
20 reefs and other calcifying organisms (Hoegh-Guldberg et al., 2007; Doney, 2010).

21  
22 Dissolved organic carbon (DOC) is a significant carbon pool in the ocean. As observed  
23 elsewhere in the global ocean, DOC concentrations in the Indian Ocean tend to be high in  
24 stratified near-surface tropical and subtropical waters where DOC is produced and accumulates,  
25 and DOC concentrations are lowest in the deep ocean where heterotrophic consumption of DOC  
26 exceeds autotrophic production (Hansell, 2009; Hansell, et al., 2009; Figures 6 and 7).  
27 Unsurprisingly, the highest concentrations of DOC are found in the near-surface waters of the  
28 central Arabian Sea due to high autotrophic production, and in the northern Bay of Bengal due to  
29 the large inputs of fresh water and associated terrigenous DOC flux from rivers (Hansell, 2009;  
30 Shah et al., 2018; Figure 6; Data from Hansell and Orellana, 2021). In contrast, particulate  
31 organic carbon (POC) concentrations tend to be high in coastal regions of the Indian Ocean  
32 where autotrophic production is high, and lowest in the oligotrophic southern subtropical gyre  
33 and equatorial waters where autotrophic production is low (Gardner et al., 2006; Figure 8). The  
34 highest POC concentrations are found in the western and northern Arabian Sea during the  
35 southwest monsoon (SWM) and northeast monsoon (NEM), respectively (Gardner et al., 2006).  
36 The spatial and temporal variability in POC export flux in the Indian Ocean is similar to the  
37 productivity and POC concentration patterns, consistent with the idea that primary productivity  
38 is the main control on the spatial and temporal variability of organic carbon fluxes (Rixen et al.,  
39 2019a). However, in regions strongly influenced by river inputs, like the Bay of Bengal, the  
40 spatial variability of organic carbon export flux is also strongly influenced by lithogenic matter  
41 content, which provides ballast that increases POC sinking rates (Rixen et al., 2019b).

42  
43 Understanding and predicting these disparate elements of the carbon cycle and its role in basin  
44 acidification in the Indian Ocean is critical for understanding the biogeochemical and ecological  
45 evolution of the Indian Ocean under the impact of human activities.

## 2. Oxygen concentrations and the biogeochemical impacts of the OMZs

The first large ocean-going oceanographic expeditions discovered OMZs in the Indian Ocean between the end of the 19<sup>th</sup> and the first third of the 20<sup>th</sup> century (Sewell, 1934; Sewell and Fage, 1948 and references therein). During one of these cruises, the John Murray expedition of 1933 – 34, Gilson (1933) discovered the secondary nitrite maximum of the Arabian Sea, seen as an accumulation of nitrite within the upper part of the OMZ between 200 and 300m depth (Figure 2). Its existence indicates active denitrification (Naqvi, 1991), which in addition to anammox (Dalsgaard et al., 2003; Kuypers et al., 2003) are the main sinks of fixed nitrogen ( $\text{NH}_4^+$ ,  $\text{NO}_2^-$ ,  $\text{NO}_3^-$ ) in the ocean (Gruber, 2004). Denitrifying microbes use nitrate ( $\text{NO}_3^-$ ) and nitrite ( $\text{NO}_2^-$ ) to fuel heterotrophic respiration in microbially hypoxic waters, in turn reducing these compounds, ultimately, to nitrogen gas ( $\text{N}_2$ ). In contrast, the annamox reaction is carried out by chemoautotrophic microbes that derive energy from the oxidation of  $\text{NH}_4^+$  with  $\text{NO}_2^-$  to produce  $\text{N}_2$  in microbially hypoxic waters.

In contrast to fixed nitrogen,  $\text{N}_2$  is inaccessible to eukaryotic phytoplankton. Even though there are specific bacterial phytoplankton clades capable of fixing or breaking apart and oxidizing  $\text{N}_2$ , the surface ocean is often depleted in fixed nitrogen in comparison to phosphate (Gruber and Sarmiento, 1997). Since phosphate and fixed nitrogen are the primary macro-nutrients required for phytoplankton growth, the intensification of OMZs and the expansion of functional anoxia is assumed to lower marine productivity by favoring denitrification and anammox and thereby the loss of fixed nitrogen in the ocean (Altabet et al., 1995; McElroy, 1983). An interesting feedback is that, over time, reduced productivity and carbon export due to fixed nitrogen loss in turn reduces biological oxygen consumption at depth, such that the expansion of the volume of anoxic waters in the OMZ core can be accompanied by a reduction of the volume of hypoxic waters (Deutsch et al., 2007; Lachkar et al., 2016).

Changes in the coupled marine nitrogen and oxygen cycles also affect the role of the ocean as a sink and/or source of greenhouse gases such as  $\text{CO}_2$  and  $\text{N}_2\text{O}$ . The response of the biologically mediated  $\text{CO}_2$  uptake of the ocean, referred to as the biological carbon pump (Boyd et al., 2019; Volk and Hoffert, 1985), to environmental changes is difficult to predict (Bopp et al., 2013, De La Rocha and Passow, 2014) because of the multiple processes involved in its functioning (Chakraborty et al., 2018; Boyd et al., 2019). In contrast, the  $\text{N}_2\text{O}$  source function of the ocean is assumed to be directly linked to the expansion and intensity of OMZs, since  $\text{N}_2\text{O}$  is an intermediate product formed during denitrification (Bange et al., 2001; Fuhrman and Capone, 1991).

Even though denitrification and anammox occur at microbially hypoxic levels of oxygen (concentrations  $< 20 \mu\text{mol/kg}$ ; Table 1), their impact on the nitrogen cycle becomes significant only with the occurrence of functional anoxia (concentrations  $< .05 \mu\text{mol/kg}$ ; Table 1). At such low oxygen concentrations, the reduction of nitrite to  $\text{N}_2$  outcompetes the re-oxidation of nitrite to nitrate, leading to net loss of fixed nitrogen (Bristow et al., 2017; Gaye et al., 2013). The re-oxidation of nitrite to nitrate prevents, in turn, the loss of fixed nitrogen and the formation of the secondary nitrite maximum at higher oxygen concentrations within the OMZ. In contrast to the Bay of Bengal OMZ, which is on the verge of becoming functionally anoxic (Bristow et al.,

1 2017), functional anoxia is wide spread in the Arabian Sea OMZ as indicated by the extent of the  
2 secondary nitrate maximum (Naqvi, 1991; Rixen et al., 2014; Figure 2).

## 3 4 **2.1 Oxygen distributions, sources and sinks**

5  
6 Low oxygen conditions in the water column are usually associated with slow ventilation and  
7 high biological oxygen demand. OMZs are hence found in productive but poorly ventilated  
8 shadow zones of the subtropical oceans, where subsurface water masses recirculate beneath a  
9 shallow mixed-layer with minimal direct advective or mixing connection to the surface ocean  
10 (Luyten et al., 1983). In these domains, the residence time below the thermocline is at a  
11 maximum and oxygen is slowly supplied mainly through isopycnal (horizontal) mixing  
12 processes (Gnanadesikan et al., 2012; Levy et al., 2021). Moreover, the strong and localized  
13 subsurface biological oxygen consumption in OMZs generates strong spatial gradients in oxygen  
14 at their edges, making mixing central to the oxygen balance. The Arabian Sea OMZ is not a  
15 classical shadow zone due to the northern boundary of the Asian subcontinent, but diapycnal and  
16 isopycnal mixing are likely the main oxygen sources balancing strong biological depletion in the  
17 core of the OMZ, in roughly equal proportions (Resplandy et al., 2012). Model simulations  
18 suggest that without oxygen supplied by eddy mixing, the volume of the Arabian Sea OMZ  
19 would double (Lachkar et al., 2016).

20  
21 In addition to mixing processes in the core of the OMZ, the inflow of oxygen-enriched Indian  
22 Ocean Central Water and Persian Gulf Water are assumed to be the main physical oxygen supply  
23 mechanism to the edges of the OMZ in the Arabian Sea (Lachkar et al., 2019; McCreary et al.,  
24 2013; Resplandy et al., 2012; Sen Gupta and Naqvi, 1984; Swallow, 1984). Indian Ocean Central  
25 Water forms through convective mixing as Subantarctic Mode Water in the southern Indian  
26 Ocean and is advected northward into the OMZs in both the Arabian Sea and Bay of Bengal  
27 (Fine, 1993; McCartney, 1979; Sverdrup et al., 1942). Oxygen-enriched Persian Gulf Water is  
28 introduced into the Arabian Sea OMZ after its outflow from the Persian Gulf (Rixen et al., 2005;  
29 Schmidt et al., 2020; Schmidt et al., 2021; Tchernia, 1980; Lachkar et al. 2019). The negative  
30 water balance of the Persian Gulf drives this localized deep-water formation by increasing the  
31 salinity and hence the density of Persian Gulf Water.

32  
33 The oxygen-poor Arabian Sea intermediate water, in turn, flows into the Bay of Bengal OMZ  
34 (Figure 1), where high freshwater inputs from rivers and monsoon rainfall (Figure 5) reduce the  
35 vertical oxygen supply by increasing stratification in the surface layers (Rixen et al., 2020).  
36 However, in comparison to the Arabian Sea, lower biomass (Figure 8), productivity and a  
37 stronger ballast effect prevent development of functional anoxia in the Bay of Bengal OMZ by  
38 keeping biological oxygen consumption lower (Al Azhar et al., 2017; Rao et al., 1994; Rixen et  
39 al., 2019b). Ballast minerals, which are supplied from land via rivers or as dust, accelerate the  
40 sinking of particles, lowering the residence time of exported organic matter in the water column  
41 and thereby its decomposition in the OMZ (Haake and Ittekkot, 1990; Ramaswamy et al., 1991).  
42 It should be noted, however, that the ballast effect appears to be important primarily in the  
43 northern Bay of Bengal (Ittekkot et al., 1991), which suggests that the primary cause of the higher  
44 oxygen concentrations further south in the Bay of Bengal is the lower biomass and productivity  
45 that results in reduced export and a lower biological oxygen demand. In contrast to the Bay of  
46 Bengal, a weak ballast effect and a higher biological production (Figure 8) sustain a higher

1 biological oxygen consumption at intermediate depths in the Arabian Sea. This enhanced  
2 biological oxygen consumption balances the higher physical oxygen supply in the Arabian Sea  
3 and explains the more intense OMZ (Rixen et al., 2019a).

4  
5 Contrary to expectations, the Arabian Sea OMZ is more intense in the eastern part of the basin  
6 and less intense in the western parts of the basin where productivity is highest (Antoine and  
7 Morel, 1996; Naqvi, 1991; McCreary et al., 2013). The thickest part of the OMZ is in the  
8 northeastern Arabian Sea (Sarma et al., 2020). Multiple factors are presumed to cause this  
9 asymmetry. The inflow of Indian Ocean Central Water and Persian Gulf Water ventilates the  
10 western Arabian Sea preferentially, and is assumed to cause this eastward displacement  
11 (Resplandy et al., 2012; McCreary et al., 2013; Rixen et al., 2014). Additionally, the seasonal  
12 monsoon-driven reversal of the surface ocean circulation leads to intense vertical eddy mixing in  
13 the western Arabian Sea causing an eastward shift of the upper OMZ relative to the region of  
14 highest productivity (Resplandy et al., 2012; McCreary et al., 2013; Lachkar et al., 2016).  
15 Similarly, the eastward shift of the OMZ has been attributed to weak mixing and high  
16 penetration time of intermediate water masses in the northeastern Arabian Sea, with the  
17 additional influence of organic matter transport from the shelf region that enhances biological  
18 oxygen demand (Sarma et al., 2020).

## 21 **2.2 Biogeochemical impacts of the northern Indian Ocean OMZs**

22  
23 The Arabian Sea and the Bay of Bengal collectively contain ~59% of the Earth's marine  
24 sediments exposed to hypoxia (Helly and Levin, 2004). Denitrification within sediments (benthic  
25 denitrification) underlying functionally anoxic conditions in the water column, is the largest sink  
26 for fixed nitrogen in the ocean (Gruber, 2004; DeVries et al., 2013). However, estimates of the  
27 relative rates of benthic and water column denitrification are still fraught with large uncertainties  
28 on global as well as regional scales. Hence, the role of the Arabian Sea as a sink of fixed  
29 nitrogen is difficult to quantify. Although denitrification in the Arabian Sea has been much more  
30 intensively studied than in the Bay of Bengal, estimates of benthic and water column  
31 denitrification in the Arabian Sea still encompass a wide range with means of  $3.9 \pm 2.9$  Tg N  
32 year<sup>-1</sup> and  $17 \pm 16$  Tg N year<sup>-1</sup>, respectively (Bange et al., 2000; Bristow et al., 2017; Deuser et  
33 al., 1978; Gaye et al., 2013; Howell et al., 1997; Naqvi et al., 1982; Somasundar et al., 1990).  
34 Considering global mean estimates of benthic and water-column denitrification of  $183 \pm 118$  Tg  
35 N year<sup>-1</sup> and  $155 \pm 116$  Tg N year<sup>-1</sup> (Eugster and Gruber, 2012; Gruber, 2004; Somes et al.,  
36 2013), on average the Arabian Sea comprises approximately 2% and 11% of the global mean  
37 benthic and water column denitrification, respectively. It should be noted, however, that more  
38 recent data indicate that benthic denitrification at the Pakistan continental margin alone could be  
39 up to  $10.5$  Tg N year<sup>-1</sup> (Schwartz et al., 2009; Somes et al., 2013), greatly exceeding the  
40 previously estimated benthic denitrification of  $3.9 \pm 2.9$  Tg N year<sup>-1</sup> for the entire Arabian Sea.  
41 Moreover, the most recent observations suggest the northeastern Arabian Sea has a water column  
42 denitrification rate of  $25.3 \pm 7.0$  Tg N yr<sup>-1</sup> (Anju et al., 2022). These estimates suggest that the  
43 Arabian Sea comprises at least 6% and 16% of the global mean benthic and water column  
44 denitrification, respectively.

1 The volume of hypoxic waters and rate of denitrification in the Arabian Sea strongly increase in  
2 response to enhanced monsoon winds (Lachkar et al., 2018). Stronger winds intensify the  
3 upwelling, increasing biological productivity and respiration in excess of the increase in wind-  
4 driven ventilation and are also responsible for the deepening of the OMZ. However, as discussed  
5 above, increases in denitrification have the potential to reduce biological productivity through  
6 nitrogen removal, and hence the efficiency of the biological pump of carbon, at the basin scale  
7 (and beyond) on timescales of decades to centuries (Lachkar et al., 2018; Canfield et al., 2019;  
8 McElroy, 1983). Therefore, an expansion of Indian Ocean OMZs can affect the large-scale  
9 biogeochemical cycles of nitrogen and carbon and contribute to climate variations over long  
10 timescales (Altabet et al., 2002; Gaye et al., 2018). The extent of this control depends on the  
11 magnitude of the nitrogen removal and the importance of concurrent stabilizing negative  
12 feedbacks, including the feedback of reduced productivity on biological oxygen demand and the  
13 tight coupling between denitrification and N<sub>2</sub> fixation downstream of OMZs (Deutsch et al.,  
14 2007, Gruber and Galloway, 2008). If the excess in phosphorus relative to nitrogen supports  
15 local nitrogen fixation within the Indian Ocean basin, the timescales for feedbacks with  
16 denitrification could be much shorter than the timescale of the global ocean circulation.

### 17 18 **2.3 Recent (decadal) changes in oxygen concentrations**

19  
20 Early reports on the occurrence of coastal hydrogen sulfide in the north-eastern Arabian Sea and  
21 off Oman at Ras-al-Hadd (Carruthers et al., 1959; Ivanenkov and Rozanov, 1966) indicate that  
22 the Arabian Sea OMZ was more intense in the past because the emergence of hydrogen sulfide  
23 indicates the transition from functional anoxia (oxygen < .05 μmol/kg) towards anoxic (zero  
24 oxygen) conditions (Table 1). These are the only reports of the occurrence of hydrogen sulfide in  
25 the Arabian Sea except for Naqvi et al. (2000), who discovered an anoxic event that developed  
26 along the western Indian coast off Mumbai in the late summer of 1999. Such strong events do  
27 not develop every year (Gupta et al., 2016; Sudheesh et al., 2016), but their appearance shows  
28 that the spreading of oxygen-depleted zones in coastal regions is a global phenomenon that does  
29 not spare the Indian shelf (Altieri et al., 2017; Diaz and Rosenberg, 2008; Diaz et al., 2019).  
30 However, contrary to this prevailing understanding, Gupta et al. (2021) argue that the hypoxic-  
31 anoxic zone along the west coast of India is formed through a natural process, i.e., upwelling of  
32 deoxygenated waters during the summer monsoon, and that the persistence and extent of this  
33 coastal oxygen deficiency depend on the degree of deoxygenation of source waters for the  
34 upwelling. Moreover, the volume of anoxic waters is strongly modulated by the Indian Ocean  
35 Dipole (IOD), with positive IOD events preventing anoxia due to wind-forced downwelling  
36 coastal Kelvin waves propagating along the west coast of India (Vallivattathillam et al., 2017).

37  
38 In contrast to these shelf processes, there is only a weak decadal decline in dissolved oxygen  
39 concentrations in the OMZs of the Arabian Sea and the Bay of Bengal in comparison to OMZs  
40 of the South Atlantic Ocean and the Pacific Ocean (Ito et al., 2017; Naqvi, 2019; Schmidtko et  
41 al., 2017; Stramma et al., 2008). The analysis of all oxygen data available from the Arabian Sea  
42 between 1959 and 2004 by Banse et al. (2014) ascribes this modest change to opposing regional  
43 trends within the Arabian Sea, where oxygen concentrations increased in the southern part of the  
44 Arabian Sea and declined in the central Arabian Sea. In contrast, Sarma et al. (2018) ascribe the  
45 modest change in oxygen levels in the Bay of Bengal to the influence of anticyclonic eddies in  
46 that precluded the OMZ from intensifying. Follow-up studies report decreasing oxygen



1 concentrations in the western and northern Arabian Sea (Piontkovski and Al-Oufi, 2015; Queste  
2 et al., 2018). In the northern Arabian Sea, dissolved oxygen concentrations in the surface mixed  
3 layer largely reflect the decreasing trend seen in the OMZ, as indicated by a compilation of  
4 dissolved oxygen data covering the period from the 1960s to 2010 (Gomes et al., 2014). In  
5 response to this deoxygenation, the secondary nitrite maximum expanded southward and  
6 westward in the early 1990s (Rixen et al., 2014). Perhaps in response to changing oxygen  
7 concentrations, the planktonic community structure has changed as shown by the development  
8 and now regular occurrence of large *Noctiluca* winter blooms since the 2000s in the Arabian Sea  
9 (Hood et al, 2022 this volume; Gomes et al, 2014; Goes and Gomes, 2016; Goes et al., 2020). It  
10 should be noted, however, that it has been shown that *Noctiluca* blooms occur in oxic conditions,  
11 and that both natural and anthropogenic processes appear to have contributed to the development  
12 of the massive blooms in the northeastern Arabian Sea (Sarma et al., 2018; Sridevi and Sarma,  
13 2022). Numerical model results suggest that reduced ventilation caused by a reduced inflow of  
14 Persian Gulf Water in response to warming, combined with reduced solubility of oxygen in  
15 surface water of the Persian Gulf, may have contributed to these developments (Lachkar et al.,  
16 2019). In addition to changes in the Gulf outflow, recent numerical model results also attribute  
17 decreasing oxygen concentrations in the northern Arabian Sea to reduced local ventilation due to  
18 global warming-induced increases in stratification and weakening of winter convective mixing  
19 (Lachkar et al., 2020).

20  
21 Nonetheless, widespread and or more frequent outbreaks of hydrogen sulfide as seen in the  
22 upwelling systems off Peru (Schunck et al., 2013) and Namibia (Weeks et al., 2002) have not so  
23 far been reported in the northern Indian Ocean during the last 50 years. This finding implies that  
24 the interplay between physical oxygen supply and the biological oxygen consumption has  
25 prevented the development of persistent anoxia (zero oxygen) in the Arabian Sea and functional  
26 anoxia (oxygen < .05  $\mu\text{M}$ ) in the Bay of Bengal OMZ (Rixen et al., 2020).

## 27 28 **2.4 Future changes in oxygen concentrations**

29  
30 There is high uncertainty in the future evolution of the Arabian Sea OMZ, and of all major  
31 OMZs, projected by Earth System models by year 2100 (Bopp et al., 2013, Cabré et al. 2015,  
32 Kwiatkowski et al., 2020; Rixen et al., 2020). The uncertainty largely arises from differences  
33 among models in the magnitude and timing of changes in ventilation and biological oxygen  
34 demand and how strongly they offset each other (Resplandy, 2018). Both processes are strongly  
35 influenced by model resolution and are thus challenging to simulate at the centennial scale. It is  
36 projected that OMZs may either shrink or expand under projected climate change, depending on  
37 the efficiency of mixing (Duteil and Oschlies, 2011, Bahl et al., 2019, Lévy et al., 2021). As  
38 discussed above, this uncertainty is particularly large in the northern Indian Ocean where global  
39 forecast models generally fail to reproduce the current oxygen concentrations and distributions  
40 (McCreary et al., 2013; Rixen et al., 2020; Schimdt et al., 2021). This failure, and the  
41 uncertainty in future projections, may also be related to the lack of consideration of the role of  
42 cross-shelf transport of organic matter in Earth System models, which is a process that is known  
43 to contribute to the development of the OMZs in both the Arabian Sea and Bay of Bengal (Sarma  
44 et al., 2020; Udaya Bhaskar et al., 2021).

## 45 46 **3. Carbon concentrations and fluxes**

1  
2 The earliest CO<sub>2</sub>, pH and POC measurements in the Indian Ocean date back to the International  
3 Indian Ocean Expedition in the early 1960s (Figure 4; e.g., Newell, 1969; Berhman, 1981).  
4 Carbonate system measurements were also made in the Indian Ocean under the Geochemical  
5 Ocean Section Study (GEOSECS) in the 1970s (Moore, 1984), the Joint Global Ocean Flux  
6 Study (JGOFS; Fasham, 2003; see also  
7 [https://en.wikipedia.org/wiki/Joint\\_Global\\_Ocean\\_Flux\\_Study](https://en.wikipedia.org/wiki/Joint_Global_Ocean_Flux_Study)) and the World Ocean  
8 Circulation Experiment (WOCE) in the 1990s (Woods, 1985). Subsequently in the first decade  
9 of the 21<sup>st</sup> century CLIVAR Repeat Hydrography project started (Gould et al., 2013), and  
10 carbonate system measurements continue through the present under the ongoing GO-SHIP  
11 program (Talley et al, 2017; see also <https://www.go-ship.org/About.html>) and the Second  
12 International Indian Ocean Expedition (IIOE-2; Hood et al., 2015). In addition to these major  
13 international efforts, there have been numerous national expeditions and programs that have  
14 contributed to the current inventory of carbonate system measurements in the Indian Ocean. For  
15 example, time-series observations of pH and pCO<sub>2</sub> are being made in the coastal waters of India  
16 revealing recent trends (Sarma et al., 2021). There is also now a mooring in the central Bay of  
17 Bengal (Figure 4) that has been collecting continuous CO<sub>2</sub> and pH measurements since  
18 November, 2013 (see <https://www.pmel.noxxaa.gov/co2/story/BOBOA>).  
19

20 In contrast, the history of DOC measurements in the Indian Ocean dates back only to the early  
21 1990s when it was discovered (thanks to advances in measurement methods that achieved ~ 1  
22 µmol/kg precision sufficient to reveal DOC variability over space and time) that DOC  
23 concentrations are more dynamic than previously thought (Hansell, 2009). The timing of this  
24 discovery coincided with the onset of the JGOFS Arabian Sea Process Study when some of the  
25 earliest “modern” Indian Ocean DOC measurements were made. Subsequently, DOC  
26 concentration measurements were made in the Indian Ocean under the US CLIVAR Repeat  
27 Hydrography project and they continue under the ongoing GO-SHIP program and through  
28 national expeditions. DOC isotopes measurements are also now being made in the northern  
29 Indian Ocean to provide insight into DOC sources (Rao and Sarma, 2022).  
30

### 31 **3.1 Inorganic carbon distributions and fluxes**

32

33 The Indian Ocean accounts for ~20% of the global oceanic uptake of atmospheric CO<sub>2</sub>  
34 (Takahashi et al., 2002). The Arabian Sea is a source of CO<sub>2</sub> to the atmosphere due to elevated  
35 pCO<sub>2</sub> within the Southwest Monsoon (SWM)-driven upwelling (Figure 4; see also Takahashi et  
36 al., 2009; 2014; Valsala and Maksyutov, 2010; 2013; de Verneil et al., 2021). North of 14°S the  
37 Indian Ocean loses CO<sub>2</sub> at a rate of 0.12 – 0.16 PgC/yr (Takahashi et al., 2002; 2009; 2014; de  
38 Verneil et al., 2021). The most intense air-sea CO<sub>2</sub> exchange occurs during the SWM where  
39 outgassing rates reach ~6 molC/m<sup>2</sup>/yr in the upwelling regions off Oman and Somalia, but the  
40 entire Arabian Sea contributes CO<sub>2</sub> to the atmosphere (Valsala and Murtugudde, 2015; Sreesh  
41 et al., 2018; 2020; de Verneil et al., 2021).  
42

43 Time series measurements have shown that the rates of increase in DIC and pCO<sub>2</sub> per decade are  
44 consistent with global trends in the southwestern coastal Bay of Bengal, whereas rates in the  
45 northwestern coastal Bay of Bengal have been observed to be 3 to 5 times higher than the global  
46 trends (Sarma et al., 2015a). Thus, the northwestern Bay of Bengal, which was previously

1 considered to be a significant sink for atmospheric CO<sub>2</sub>, now seems to have become a source of  
2 CO<sub>2</sub> to the atmosphere. Variability in CO<sub>2</sub> fluxes in the northwestern shelf region of the Bay of  
3 Bengal depends on the river discharge characteristics and the East India Coastal Current that  
4 distributes this water along the coast (Sarma et al., 2012). Nonetheless, it is still uncertain  
5 whether the entire Bay of Bengal is a net CO<sub>2</sub> source or sink due to the high levels of variability  
6 combined with sparse spatial and temporal sampling (Figure 4; Bates et al., 2006). South of 14°S  
7 the Indian Ocean appears to be a strong net CO<sub>2</sub> sink (-0.44 PgC/yr in the band 14°S-50°S;  
8 Figure 4). The solubility pump (CO<sub>2</sub> dissolution and its physical mixing and transport) and the  
9 biological pump (biologically mediated processes that export carbon) contribute equally to the  
10 CO<sub>2</sub> sink in the south Indian Ocean region (Valsala et al., 2012). Cold temperatures increase CO<sub>2</sub>  
11 solubility at higher latitudes and subduction in the subtropical front can transport this CO<sub>2</sub> into  
12 the ocean interior, but there is also evidence that chemical and biological factors are important,  
13 e.g., potential iron fertilization that facilitates particulate carbon export in the southern  
14 hemisphere of the Indian Ocean (Piketh et al., 2000).

15  
16 In addition, the IOD leads to a substantial sea-to-air CO<sub>2</sub> flux variability in the southeastern  
17 tropical Indian Ocean over a broad region (70–105°E, 0–20°S), with the most intense effects  
18 manifesting near the coast of Java-Sumatra due to the impacts on upwelling dynamics and  
19 associated westward propagating anomalies. The sea-to-air CO<sub>2</sub> fluxes, surface ocean partial  
20 pressure of CO<sub>2</sub> (pCO<sub>2</sub>), the concentration of dissolved inorganic carbon (DIC), and ocean  
21 alkalinity (ALK) range as much as ±1.0 mole m<sup>-2</sup> yr<sup>-1</sup>, ±20 µatm, ±35 µmole kg<sup>-1</sup>, and ±22  
22 µmole kg<sup>-1</sup>, respectively, within 80–105°E, 0–10°S due to the IOD. The DIC and ALK are  
23 significant drivers of pCO<sub>2</sub> variability associated with IOD (Valsala et al., 2020).

24  
25 Synthesis of the seasonal, annual and interannual air-sea CO<sub>2</sub> fluxes based on both models  
26 (ocean, atmospheric inversions) and observations (Takahashi et al., 2009) reveals that the net  
27 sea-air CO<sub>2</sub> uptake estimated from observations (-0.24 PgC/yr) is consistent with uptake derived  
28 from models (-0.37 PgC/yr), given the uncertainties (Sarma et al., 2013). However, some models  
29 overestimate flux in the Bay of Bengal and underestimate flux in the Arabian Sea / northwestern  
30 Indian Ocean (Sarma et al., 2013). These offsets are likely due to the models being inadequately  
31 constrained by CO<sub>2</sub> observations. There are fewer observations in the Indian Ocean (especially  
32 north of 20°S) compared to other oceans in recent years (Figure 4; Bakker et al., 2014, see also  
33 [www.socat.info](http://www.socat.info)). More observations of ocean DIC concentrations and carbon flux are needed in  
34 the Indian Ocean to constrain the models and reduce uncertainties in the fluxes. Toward this  
35 goal, Valsala et al. (2021) has recently done observing system simulation experiments for Indian  
36 Ocean pCO<sub>2</sub> arrays and recommended deployment of additional moorings at suitable locations to  
37 monitor and better constrain the Indian Ocean air-sea CO<sub>2</sub> fluxes. A few key ship-of-opportunity  
38 routes for underway pCO<sub>2</sub> sampling in the Indian Ocean are also recommended in their study.

### 39 40 **3.2 Spatial and temporal variability in pH**

41  
42 Uptake of anthropogenic CO<sub>2</sub> by the ocean results in fundamental changes in seawater chemistry  
43 that can have significant impacts on ocean ecosystems (Doney et al., 2010; Gattuso and Hansson,  
44 2011). Intergovernmental Panel on Climate Change (IPCC) business-as-usual emission scenarios  
45 indicate that atmospheric CO<sub>2</sub> levels will reach 800 ppm near the end of this century (Feely et  
46 al., 2009). The associated increase in oceanic CO<sub>2</sub> concentrations will lead to acidification (lower

1 pH) of the Indian Ocean over the coming decades, with potential severe negative impacts on  
2 coral reefs and other calcifying organisms (Doney, 2010; IPCC, 2019). In 1995, the surface pH  
3 values for the northern (20°E-120°E, 0°-24.5°N) and southern (20°E-120°E, 0°-40°S) Indian  
4 Oceans were 8.068 +/- 0.03 and 8.092 +/- 0.03, respectively, which is the lowest of the major  
5 ocean basins (Feely et al., 2009). It is not entirely clear why surface pH values are so low in the  
6 Indian Ocean (Takahashi et al., 2014). Increases in sulphate and nitrogen aerosol loadings over  
7 the Bay of Bengal from the Indo-Gangetic Plain and Southeast Asia may be mainly responsible  
8 for the increased acidity in the northwestern Bay of Bengal in recent years (Sarma et al., 2015a).  
9 Reduced Godavari River discharge together with a positive IOD event have also been shown to  
10 contribute to enhanced acidification and pCO<sub>2</sub> levels in the coastal waters in the western Bay of  
11 Bengal (Sarma et al., 2015b). A study from the eastern Bay of Bengal indicates a decline in pH  
12 of 0.2 from 1994 to 2012 (Rashid et al., 2013), which is considerably faster than the global  
13 decline of 0.1 over the last century (IPCC 2007). Upwelling of low pH subsurface waters in the  
14 Arabian Sea results in low surface pH (<7.9) during the SWM (Takahashi et al., 2014). Sreesh  
15 et al. (2019) estimated that in addition to the anthropogenic causes, ocean warming exacerbates  
16 acidification in the western Arabian Sea by an additional 16%. Chakraborty et al. (2021) studied  
17 the seasonal drivers of surface ocean pH in the Arabian Sea and Bay of Bengal and found that  
18 DIC and SST make complementary contributions to the seasonal cycle in pH. Tarique et al.  
19 (2021) reported the first time series of Arabian Sea pH from proxy records of Boron Isotopes  
20 from 1990 to 2013. Their investigation reveals that physical oceanographic processes, for  
21 example, upwelling, downwelling and convective mixing modulated by El Niño-Southern  
22 Oscillation (ENSO), largely control surface pH variability and mask expected long-term ocean  
23 acidification trends resulting from anthropogenic CO<sub>2</sub> rise. An increase in pH has been observed  
24 in the eastern and southern Bay of Bengal during all seasons associated with warming and a  
25 decrease in salinity (Sridevi and Sarma, 2021). In contrast, a decrease in pH (-0.001 yr<sup>-1</sup>) and a  
26 pCO<sub>2</sub> increase (+0.1 to +0.7 μatm yr<sup>-1</sup>) has been observed in the western and northern Bay of  
27 Bengal during winter and spring seasons due to deposition of atmospheric pollutants (Sridevi and  
28 Sarma, 2021). These studies suggests that increases in freshwater input due to melting of  
29 Himalayan ice cover and deposition of atmospheric pollutants are dominant controlling factors  
30 on surface ocean pH and pCO<sub>2</sub> in the Bay of Bengal between 1998 and 2015, and this region is  
31 acting as a stronger sink for the atmospheric CO<sub>2</sub> in the present than in the past two decades  
32 (Sarma et al., 2021; Sridevi and Sarma, 2021).

33  
34 The susceptibility of the Indian Ocean coral communities to warming has been revealed by the  
35 large-scale coral bleaching events of 1998, 2005, 2011 as well as between 2014 and 2017 caused  
36 by high SST (McClanahan et al., 2007; Moore et al., 2012, Obura et al. 2017, Cerutti et al.  
37 2020). Ocean acidification has the potential to exacerbate these negative impacts on coral reef  
38 ecosystems. For example, the 1998 bleaching event altered the age distribution of commercially  
39 harvested fish (Graham et al. 2007). Because of the combined effects of acidification, human  
40 development and global warming, coral reef ecosystems may be at greater risk than previously  
41 thought (Hoegh-Guldberg et al., 2007; IPCC, 2019). Moreover, some commercially fished  
42 species (e.g., shelled mollusks) are directly vulnerable to ocean acidification (Hoegh-Guldberg et  
43 al., 2014). A study off Somalia suggests that human-induced ocean acidification reduced the rate  
44 at which foraminifera calcify, resulting in lighter shells (de Moel et al., 2009). Due to the effects  
45 of acidification on calcifying pteropods, which are preyed on by many higher trophic level

1 organisms, the Southern Ocean sector of the Indian Ocean could experience major disruptions in  
2 pelagic food webs (Bednarsek et al., 2012).

3  
4 In addition to the direct impacts of acidification, increasing CO<sub>2</sub> in the upper ocean could lead to  
5 increased primary productivity for some species (e.g., diazotrophic cyanobacteria; Hutchins et  
6 al., 2007), altering rates of nitrogen fixation and therefore the biogeochemistry of particulate  
7 organic matter formation and remineralization. Declines in pH also shift the chemical  
8 equilibrium from ammonia (NH<sub>3</sub>) to ammonium (NH<sub>4</sub><sup>+</sup>), which could alter key biogeochemical  
9 processes such as nitrogen assimilation by phytoplankton and microbial nitrification (Gattuso  
10 and Hansson, 2011).

### 11 **3.3 Large scale DOC and POC distribution and fluxes**

12  
13  
14 Near-surface (< 50 meters) DOC concentrations in the Indian Ocean vary between ~50 and 80  
15 μmolC/kg with the lowest values occurring south of 40° S (Hansell, 2009; Hansell et al., 2009;  
16 Figures 6, 7). In general, near-surface concentrations of DOC are positively correlated with  
17 temperature and, to a lesser extent, they are negatively correlated with nutrient concentrations.  
18 The positive correlation with temperature happens because warm, low-latitude waters tend to be  
19 more strongly stratified, and there is reduced vertical mixing. As a result, DOC concentrations  
20 increase in these waters even though there are low rates of net DOC production because the DOC  
21 is not mixed away (Hansell, 2009). As a result, near-surface DOC concentrations in the Indian  
22 Ocean tend to be highest (> 70 μmolC/kg) in tropical and subtropical waters (Hansell, 2009;  
23 Hansell et al., 2009; Figure 7). These patterns are also found in the Atlantic and Pacific  
24 (Hansell, 2009; Hansell et al., 2009).

25  
26 Near-surface DOC concentrations tend to be inversely correlated with nitrate, silicate and  
27 phosphate concentrations because, when these nutrients are consumed, DOC is produced  
28 (Hansell, 2009). Thus, a negative correlation is also observed between DOC and nitrate and  
29 phosphate concentrations in the deep waters (> 500 meters) of the Indian Ocean where DOC  
30 concentrations decline to < 50 μmolC/kg (Figure 7) and nitrate and phosphate concentrations are  
31 substantially enriched. However, these correlations are not observed between DOC and silicate  
32 concentrations because in deep waters of the Indian Ocean there are large gradients in silicate  
33 that are associated with very little change in DOC (Hansell, 2009).

34  
35 As in the Pacific and Atlantic, the lowest DOC concentrations in the Indian Ocean (40 - 42  
36 μmol/kg at 3000 meters) are found below 1000 meters depth (Figure 7; Figure 2 in Hansell, et  
37 al., 2009). Global observations show that DOC concentrations below 1000 meters depth in the  
38 Indian Ocean are comparable to those in the Atlantic and significantly higher than deep DOC  
39 concentrations in the Pacific (< 40 μmolC/kg at 3000 meters; Hansell et al., 2009). DOC  
40 concentrations decline below the euphotic zone because heterotrophic consumption exceeds  
41 autotrophic production (Hansell et al., 2009). Presumably, a significant fraction of the DOC  
42 below 1000 meters in the Indian Ocean is very old (> 4000 years) and refractory (resistant to  
43 bacterial consumption) as has been shown to be the case in the Atlantic and Pacific (Bercovici et  
44 al., 2018).

1 As observed elsewhere in the global ocean, satellite-estimated near-surface POC concentrations  
2 are elevated in coastal regions of the Indian Ocean with values often exceeding  $120 \text{ mgC/m}^3$  (=  $10 \text{ } \mu\text{molC/kg}$ ; Figure 8; Gardner et al., 2006). The highest POC concentrations in the Indian  
3 Ocean are observed in the northwestern part of the basin in the Arabian Sea and off of the coast  
4 of Somalia with values estimated to be greater than  $96 \text{ mgC/m}^3$  (=  $8 \text{ } \mu\text{molC/kg}$ ) extending well  
5 into the open ocean during the NEM (Figure 8). These high POC values are associated with high  
6 chlorophyll concentrations, although chlorophyll itself only accounts for  $> .6\%$  of the POC and  
7 presumably the remainder is cellular and detrital material (Gardner et al., 2006). Note that C:Chl  
8 ratios range from  $\sim 15$ - $158$  by weight (Sathyendrenath et al, 2009). Elevated POC concentrations  
9 ( $> 72 \text{ mgC/m}^3 = 6 \text{ } \mu\text{molC/kg}$ ) are also observed in the Indonesian Throughflow region  
10 (particularly during the Southeast Monsoon, SEM) and off of southern Africa over the Agulhas  
11 Bank, and in the Agulhas Current and its retroflexion, again associated with high chlorophyll  
12 concentrations ( $> .4\%$  of POC; Figure 8, Gardner et al., 2006). The lowest POC concentrations  
13 in the Indian Ocean are observed between  $40^\circ \text{ S}$  and  $10^\circ \text{ N}$  in the oligotrophic southern  
14 subtropical gyre and equatorial waters where primary production is very low (Gardner et al.,  
15 2006; Figure 8; see below).  
16

17  
18 Unsurprisingly, the spatial and temporal variability in POC export flux in the Indian Ocean is  
19 similar to the above mentioned patterns in POC and chlorophyll concentration, consistent with  
20 primary production as the main control on the spatial and temporal variability of organic carbon  
21 fluxes (Rixen et al., 2019a). However, as discussed above, in river-influenced regions like the  
22 Bay of Bengal, the spatial variability of organic carbon flux is also strongly influenced by  
23 lithogenic matter content. Rixen et al. (2019b) estimate that lithogenic matter content enhances  
24 organic carbon flux rates on average by 45% and by up to 62% in river-influenced regions of the  
25 Indian Ocean. This strong ballast effect explains why organic carbon fluxes are lower in the  
26 highly productive western Arabian Sea than they are in the relatively unproductive southern Java  
27 Sea. This explanation appears to be broadly consistent with the earlier global estimates of  
28 organic carbon burial and sediment accumulation rates calculated by Janke (1996; Figure 9),  
29 indicating that both are elevated in the coastal and northern regions of the Arabian Sea where  
30 rates of primary production are very high, but river influence is small. In contrast, both organic  
31 carbon burial and sediment accumulation rates are elevated over a much larger area in the Bay of  
32 Bengal where primary production is lower, but river influence is large (Janke, 1996; Figure 9).  
33

### 34 **3.4 Regional DOC and POC distributions and fluxes**

#### 35 **3.4.1 Arabian Sea**

36  
37  
38 DOC in the surface ocean is a short-term reservoir for carbon, expanding and contracting  
39 seasonally. In the Arabian Sea the highest near-surface ( $< 50$  meters) DOC concentrations ( $80$ -  
40  $100 \text{ } \mu\text{molC/kg}$ ) are observed near the coast when upwelling is not active. During the SWM,  
41 upwelling results in lowered near-surface DOC concentrations along the western side of the  
42 basin, i.e., DOC concentrations increase seaward from  $< 70 \text{ } \mu\text{molC/kg}$  near the coast to  $> 80$   
43  $\text{ } \mu\text{molC/kg}$  in the central Arabian Sea. In the open ocean, the highest surface DOC concentrations  
44 ( $80$ - $95 \text{ } \mu\text{molC/kg}$ ) are observed during the Northeast Monsoon (NEM) and they remain high  
45 through mid SWM. The lowest open ocean near-surface DOC concentrations ( $65$ - $75 \text{ } \mu\text{molC/kg}$ )

1 are observed during late SWM and during the Fall Intermonsoon (Hansell and Peltzer, 1998;  
2 Hansell, 2009).

3  
4 The seasonal accumulation of DOC north of 15°N in the Arabian Sea happens mostly during the  
5 NEM, and it has been estimated to be equivalent to 6-8% of annual primary production and 80%  
6 of net community production. In contrast, net DOC production is very small during the SWM  
7 (Hansell and Peltzer, 1998).

8  
9 In the vertical, DOC concentrations decrease to values of <55  $\mu\text{molC/kg}$  at the top of the OMZ  
10 (~100 meters). Vertical layering is also observed in DOC concentrations in the Arabian Sea and  
11 is associated with vertical layering of the water masses, with Persian Gulf Water producing a  
12 clear signal of elevated DOC. DOC concentrations below 500 meters are low (<50  $\mu\text{molC/kg}$ )  
13 and relatively uniform across the basin and there appears to be little impact of the OMZ on DOC  
14 concentrations (Hansell and Peltzer, 1998; Hansell, 2009).

15  
16 Consistent with satellite estimates (Gardner et al., 2006), in situ measurements of near surface (0  
17 - 150 meters) POC concentrations in the Arabian Sea are generally high near the coast of Oman  
18 and decrease offshore throughout the year with concentrations often exceeding 12.5  $\mu\text{molC/kg}$   
19 (Gunderson et al., 1998; Figure 10). The spatial distributions are always patchy due to the  
20 influence of mesoscale eddies and jets. Seasonal variations in POC concentrations in the upper  
21 150 m can be the same order of magnitude as basin-wide spatial variations and these  
22 concentrations are strongly influenced by nutrient availability and mixing (Gunderson et al.,  
23 1998). POC and chlorophyll concentrations are clearly correlated (Figures 10 and 11), indicating  
24 that variability in phytoplankton production drives much of the observed POC variations  
25 (Gunderson et al., 1998; Gardner et al., 2006). Subsurface POC and chlorophyll maxima are also  
26 observed, particularly offshore and during the intermonsoon periods with the latter generally  
27 deeper and more pronounced, as observed elsewhere (e.g., Fennel and Boss, 2003; see also  
28 review by Cullen, 2015). Interestingly, when POC and chlorophyll concentrations are vertically  
29 integrated the values can be higher during the spring intermonsoon than during the SWM  
30 because of the strong subsurface POC and chlorophyll maxima that develop during the spring  
31 intermonsoon (Gunderson, et al., 1998).

32  
33 In the Arabian Sea, POC export normalized to 2000 meters ranges from < 6 to > 22  $\text{mgC m}^{-2} \text{d}^{-1}$   
34 (equal to < 0.5 to > 1.8  $\text{mmolC m}^{-2} \text{d}^{-1}$ ) depending upon location and season (Rixen et al., 2019;  
35 Honjo et al., 1999). In contrast, POC export from the base of the euphotic zone in the western  
36 Arabian Sea estimated from the satellite-derived net primary production data and export fluxes  
37 calculated with a model give 258.3 and 335.7  $\text{mgC m}^{-2} \text{d}^{-1}$  (equal to 21.5 and 27.9  $\text{mmolC m}^{-2} \text{d}^{-1}$   
38 <sup>1</sup>), respectively (Sreeush et al., 2018), suggesting dramatic declines in export at depth compared  
39 to the surface. The POC fluxes are generally highest in the western Arabian Sea during the SWM  
40 (>20  $\text{mgC m}^{-2} \text{d}^{-1}$  equal to > 1.7  $\text{mmolC m}^{-2} \text{d}^{-1}$ ) where large seasonal variations in flux (7-24  
41  $\text{mgC m}^{-2} \text{d}^{-1} = 0.6\text{-}2 \text{ mmolC m}^{-2} \text{d}^{-1}$ ) are also observed (Honjo et al., 1999; Rixen et al., 2019b).  
42 In contrast, POC fluxes at 2000 meters are generally much lower in the central Arabian Sea (4-  
43 12  $\text{mgC m}^{-2} \text{d}^{-1} = 0.25\text{-}0.5 \text{ mmolC m}^{-2} \text{d}^{-1}$ ) with the highest fluxes (> 10  $\text{mgC m}^{-2} \text{d}^{-1}$  equal to >  
44 0.8  $\text{mmolC m}^{-2} \text{d}^{-1}$ ) also occurring during SWM, though with much less pronounced seasonal  
45 variations (Rixen et al., 2019; Honjo et al., 1999). These patterns of POC flux in the Arabian  
46 Sea are broadly consistent with organic carbon burial and sediment accumulation rates calculated

1 by Janke (1996), though with the latter an order of magnitude lower (Figure 9;  $<5$  to  $>50$   $\mu\text{molC}$   
2  $\text{m}^{-2} \text{yr}^{-1}$  which is equal to  $<0.01$  to  $>.14$   $\mu\text{molC} \text{m}^{-2} \text{d}^{-1}$ ). Major flux events are also observed in  
3 the western Arabian Sea during both the SWM and NEM in association with passing eddies and  
4 wind-curl events; these events can dominate the annual mass flux (Honjo et al., 1999). <sup>234</sup>Th-  
5 estimated POC export fluxes in the Arabian Sea reveal similar spatial and temporal patterns with  
6 POC export efficiencies varying from  $<2$  to  $5\%$  (Subha Anand et al, 2018a,b).

### 8 **3.4.2 Bay of Bengal**

9  
10 Some of the highest concentrations of DOC in the Bay of Bengal ( $75$ - $100$   $\mu\text{molC/kg}$ ) are  
11 observed in the near-surface ( $< 50$  meters) waters of the northern Bay of Bengal, primarily due to  
12 the large inputs of fresh water and terrigenous DOC from rivers (Hansell, 2009; Shah et al.,  
13 2018; Figure 7). In addition, this fresh water enhances near-surface stratification which helps to  
14 maintain the elevated DOC concentrations because it inhibits vertical mixing, as discussed  
15 above. The concentration of DOC in the upper ocean in the Bay of Bengal also exhibits a  
16 significant relationship with Chlorophyll-a, POC and DOC exudation rates, suggesting possible  
17 sources through in situ biological processes (Rao and Sarma, 2022). Elevated DOC  
18 concentrations are also observed to  $> 200$  meters depth in the Bay of Bengal due to the  
19 remineralization of sinking POM from the surface waters (Shah et al., 2018; Figure 7). Near-  
20 surface DOC concentrations tend to decline at lower latitudes ( $<14^\circ\text{N}$ , Hansell, 2009; Shah et al.,  
21 2018). Rao et al. (2021) report that about half of the primary production in the Bay of Bengal is  
22 released as DOC during the summer monsoon due to existence of oligotrophic conditions, warm  
23 waters and dominance of picophytoplankton biomass and Shah et al. (2018) estimate that DOC  
24 remineralization fuels  $\sim 18\%$  of the apparent oxygen utilization. As observed in the Arabian Sea,  
25 DOC concentrations below  $500$  meters in the Bay of Bengal are low ( $<50$   $\mu\text{molC/kg}$ ) and  
26 relatively uniform across the basin, and there appears to be little impact of the OMZ on DOC  
27 concentrations (Figure 7).

28  
29 In the Bay of Bengal, the near-surface concentrations of POC can vary from  $< 4$  to  $> 10$   $\mu\text{mol/kg}$   
30 depending on location and season (Fernandes et al., 2009; Figure 12). Spatial and temporal  
31 variations are smaller compared to the Arabian Sea with POC concentrations varying from  $4.3$  to  
32  $11.1$   $\mu\text{molC/kg}$ ,  $3.1$  to  $10.9$   $\mu\text{molC/kg}$ , and  $4.3$  to  $9.0$   $\mu\text{molC/kg}$  during SWM, fall intermonsoon,  
33 and spring intermonsoon, respectively, in one study (Fernandes et al. 2009). In contrast to the  
34 Arabian Sea, POC and chlorophyll concentrations can be higher offshore compared to coastal  
35 stations in the Bay of Bengal, especially during SWM (Fernandes et al., 2009; Vinayachandran,  
36 2009; Thushara et al., 2019). These elevated offshore POC and chlorophyll concentrations are  
37 not, however, revealed by remote sensing. Rather, the satellite estimates generally indicate that  
38 the highest concentrations occur near the coast (Gardner et al., 2006; Figure 8). As in the  
39 Arabian Sea, POC and chlorophyll concentrations are correlated in the Bay of Bengal with  
40 subsurface chlorophyll maxima generally deeper and more pronounced, as discussed above (e.g.,  
41 Fernandes et al., 2009). In general, POC concentrations and percent POC contribution to the  
42 total suspended particulate matter tends to decrease with increasing water column depth  
43 indicating heterotrophic remineralization of sinking particles (Fernandes et al., 2009). High  
44 chlorophyll concentrations and high Chl $a$ /POC ratios observed during the SWM and fall  
45 intermonsoon indicate the presence of relatively fresh POM in the Bay Bengal during these two  
46 seasons (Fernandes et al., 2009). POC concentration measured during the spring intermonsoon



1 were lowest in the southern Bay of Bengal (0.51–0.65  $\mu\text{molC/kg}$ , Subha Anand et al., 2017). In  
2 general, seasonal and spatial differences in river influence, combined with physical forces such  
3 as eddies and mixing/entrainment that pump nutrients into the euphotic zone, drive seasonal and  
4 spatial variations in the quantity and quality of POM in the Bay of Bengal (Fernandes et al.,  
5 2009; Vinayachandran, 2009; Subha Anand et al., 2017; Thushara et al., 2019).

6  
7 POC export normalized to 2000 meters in the Bay of Bengal is generally lower and less strongly  
8 seasonal than in the Arabian Sea, with fluxes varying between 2 and 14  $\text{mgC m}^{-2} \text{d}^{-1}$  (= 0.17 –  
9 1.2  $\text{mmolC m}^{-2} \text{d}^{-1}$ ; Rixen et al., 2019b). In contrast, POC export from the euphotic zone in the  
10 Sri Lanka Dome region estimated from the satellite-derived net primary production data and  
11 export fluxes calculated with a model give 118.5 and 427.4  $\text{mgC m}^{-2} \text{d}^{-1}$  (equal to 9.8 and 35.6  
12  $\text{mmolC m}^{-2} \text{d}^{-1}$ ), respectively (Sreeush et al., 2018), again suggesting dramatic declines in export  
13 at depth compared to the surface. In general, riverine freshwater and nutrient inputs to the Bay  
14 of Bengal increase POC export near the coast. This pattern is revealed by  $^{234}\text{Th}$ -estimated POC  
15 export flux in the Bay of Bengal during the NEM, which varies from 0.1 to 1.6  $\text{mmolC m}^{-2} \text{d}^{-1}$   
16 with the highest flux observed in the east Indian coastal zone (Subha Anand et al., 2017). These  
17 patterns of POC flux are also broadly consistent with organic carbon burial and sediment  
18 accumulation rates calculated by Janke (1996), where the former are highest along the east coast  
19 of India (Figure 9). Though, as in the Arabian Sea, the organic carbon burial rates estimated by  
20 Janke (1996) are an order of magnitude lower than the POC fluxes normalized to 2000 meters  
21 from Rixen et al. (2019).

### 22 23 **3.4.3 Equatorial waters, Indonesian Throughflow, and the Leeuwin Current**

24  
25 A modeling study reported by Hansell (2009) and Hansell et al. (2009) suggests a zonal decline  
26 in near-surface (30 meter) DOC concentrations along the equator in the Indian Ocean with values  
27 dropping from  $\sim 75 \mu\text{molC/kg}$  in the east off of northern Sumatra to  $< 70 \mu\text{molC/kg}$  in the west  
28 off of central east Africa. This pattern appears to be linked to the aforementioned declines in  
29 near-surface DOC concentrations to  $< 70 \mu\text{molC/kg}$  that are associated with upwelling along the  
30 western side of the basin (Hansell and Peltzer, 1998; Hansell, 2009). The modeling study  
31 reported by Hansell (2009) and Hansell et al. (2009) also suggests that near-surface DOC  
32 concentrations are elevated to  $> 75 \mu\text{molC/kg}$  in an open ocean region between  $10^\circ$  and  $20^\circ\text{S}$  and  
33  $80^\circ$  and  $100^\circ\text{E}$  coincident with shallow OMZ (Figure 1), minimum sedimentation rate (Figure 9),  
34 and lower salinity (Figure 5), and that concentrations decline eastward to  $< 70 \mu\text{molC/kg}$  in the  
35 the Indonesian Throughflow region and in the Leeuwin Current, but these features are not readily  
36 apparent in meridional or zonal sections (Figure 7). DOC concentrations below 500 meters in  
37 these waters are also low ( $< 50 \mu\text{molC/kg}$ ) and uniform (Figure 7).

38  
39 Satellite-estimated near-surface POC concentrations in open ocean equatorial waters of the  
40 Indian Ocean are generally low ( $< 60 \text{mgC m}^{-3}$  which is equal to  $< 5 \mu\text{molC/kg}$ ) with  
41 concentrations increasing during the SWM, especially in the west due to the aforementioned  
42 influence of coastal upwelling (Gardner et al., 2006; Figure 8). Elevated POC concentrations ( $>$   
43  $72 \text{mgC/m}^3$  which is equal to  $> 6 \mu\text{molC/kg}$ ) are also observed in the eastern equatorial Indian  
44 Ocean along the coast off Sumatra and Java and in the Indonesian Throughflow region,  
45 particularly during the SEM. All of these regions of elevated POC are associated with elevated  
46 chlorophyll concentrations ( $> 0.4\%$  of POC; Figure 8, Gardner et al., 2006). Unsurprisingly,

1 POC and chlorophyll concentrations are generally low along the coast of western Australia  
2 where the downwelling Leeuwin Current flows southward (Figure 8). Estimated organic carbon  
3 burial rates show similar patterns, i.e., low rates ( $< 5 \text{ mmolC m}^{-2} \text{ yr}^{-1}$ ) in open ocean equatorial  
4 waters and higher rates in the west and east in association with upwelling and the influence of  
5 the Indonesian Throughflow (Janke, 1996; Figure 9). Surprisingly, estimated organic carbon  
6 burial rates are distinctly elevated ( $5\text{-}10 \text{ mmolC m}^{-2} \text{ yr}^{-1}$ ) along the coast of Western Australia  
7 (Janke, 1996; Figure 9), likely due to coastal mixing processes, intermittent localized upwelling  
8 and eddy generation in the otherwise oligotrophic downwelling Leeuwin Current (Hood et al.,  
9 2017).

10  
11  $^{234}\text{Th}$ -estimates of POC export flux have revealed elevated values in eastern-central equatorial  
12 waters of the Indian Ocean during spring intermonsoon (up to  $7.7 \text{ mmolC m}^{-2} \text{ d}^{-1}$ ) in association  
13 with elevated rates of primary production (Subha Anand et al., 2017). These elevated rates are  
14 somewhat surprising given the downwelling circulation that is associated with the eastward-  
15 flowing equatorial currents (Strutton et al., 2015), but they are consistent with eastward increases  
16 in organic carbon burial and sedimentation rates along the equator (Janke, 1996; Figure 9). They  
17 are also consistent with eastward increases POC export from the euphotic zone estimated from  
18 the satellite-derived net primary production data and export fluxes calculated with a model that  
19 give  $149.4$  and  $590.5 \text{ mgC m}^{-2} \text{ d}^{-1}$  (equal to  $12.54$  and  $49.2 \text{ mmolC m}^{-2} \text{ d}^{-1}$ ), respectively, off  
20 Sumatra, compared to similarly estimated values of  $110.8$  and  $220.1 \text{ mgC m}^{-2} \text{ d}^{-1}$  (equal to  $9.2$   
21 and  $18.3 \text{ mmolC m}^{-2} \text{ d}^{-1}$ ), respectively, in the Seychelles-Chagos Thermocline Ridge region  
22 (Sreeush et al., 2018).

23  
24 There are few direct measurements of POC concentrations and export flux off of western  
25 Australia. Waite et al. (2016) has shown that mesoscale eddies generated by the Leeuwin Current  
26 can strongly impact carbon export fluxes. Specifically, the subsurface distribution of particles in  
27 these eddies funnel into a wineglass shape down to  $1000 \text{ m}$  (Figure 13), leading to a sevenfold  
28 increase of vertical carbon flux in the eddy center versus the eddy flanks. This is consistent with  
29 the aforementioned idea that eddy generation in the Leeuwin Current gives rise to elevated  
30 organic carbon burial rates along the coast of Western Australia (Janke, 1996; Figure 9).

### 31 32 **3.4.4 Southwestern Indian Ocean**

33  
34 Although the modeling study reported by Hansell (2009) and Hansell et al. (2009) suggests that  
35 near-surface ( $30 \text{ meter}$ ) DOC concentrations are lower ( $< 65 \text{ }\mu\text{molC/kg}$ ) in the southwestern  
36 Indian Ocean and in the Mozambique Channel, these declines are not apparent in zonal sections  
37 (e.g., Figure 7). Rather, a meridional section along  $\sim 30^\circ \text{ E}$  (IO6N from Arctica to western South  
38 Africa) shows a very clear increase in near-surface ( $< 50 \text{ meter}$ ) DOC concentrations north of  
39  $45^\circ \text{ S}$  from  $< 50 \text{ }\mu\text{molC/kg}$  to  $> 70 \text{ }\mu\text{molC/kg}$  (Hansel, 2009; Figure 7). Moreover, these elevated  
40 concentrations extend downward to  $> 300 \text{ meters}$  depth (Figure 7). These data suggest that the  
41 Agulhas Current advects relatively high tropical and subtropical DOC concentrations  
42 southwestward along the coast of South Africa and also eastward in the Agulhas Retroflexion,  
43 and that these elevated concentrations are being mixed downward in the current. This  
44 speculation is consistent with the fact that the Agulhas Current extends to  $> 1000 \text{ meters}$  depth  
45 and is derived from oligotrophic tropical and subtropical sources waters (Hood et al., 2017) that

1 have relatively high DOC concentrations (Figure 7). DOC concentrations below 1000 meters in  
2 these waters are low ( $<50 \mu\text{molC/kg}$ ) and relatively uniform (Figure 7).

3  
4 Satellite estimates suggest that POC concentrations are generally low ( $< 72 \text{ mgC m}^{-3}$  which is  
5 equal to  $< 6 \mu\text{molC/kg}$ ) in the southwestern Indian Ocean and in the Agulhas Current during both  
6 the austral summer and winter seasons (Gardner et al., 2006; Figure 8). But some elevated POC  
7 concentrations ( $> 72 \text{ mgC/m}^3$ ,  $> 6 \mu\text{molC/kg}$ ) are observed in the austral winter in the  
8 Mozambique Channel, along the southwestern coast of South Africa, over the Agulhas Bank and  
9 in the Agulhas retroflection (Gardner et al., 2006; Figure 8). All of these regions are associated  
10 with elevated chlorophyll concentrations ( $> .4\%$  of POC; Figure 8, Gardner et al., 2006). In the  
11 Mozambique Channel the elevated concentrations are associated with entrainment and offshore  
12 advection of coastal POM in eddies (Kolasinsky et al., 2012), whereas the elevated POM  
13 concentrations along the southwestern coast of South Africa and over the Agulhas Bank are  
14 associated with wind-induced and topographically controlled coastal upwelling (Hood et al.,  
15 2017; Hood et al., 2021, this volume). The tongue of elevated POC and chlorophyll  
16 concentration that is associated with the Agulhas retroflection is particularly striking (Figure 8).  
17 This is a region where strong eastward flows and persistent eddies are observed (Pazan and  
18 Niiler, 2004). It is possible that these eddies mix or upwell nutrients into the euphotic zone,  
19 fueling a low level of primary production. This region of elevated POC and chlorophyll extends  
20 all the way to the Kerguelen Islands during the austral summer (Figure 8; Gardner et al., 2006).

21  
22 Estimated organic carbon burial rates in the southwestern Indian Ocean range from  $< 5 - \sim 10$   
23  $\text{mmolC m}^{-2} \text{ yr}^{-1}$  with the highest rates observed along the coastal zone of southeastern Africa  
24 (Janke, 1996; Figure 9). Interestingly, these estimates also clearly show elevated organic carbon  
25 burial (and sedimentation) rates in the same region where there is a tongue of elevated near-  
26 surface POC and chlorophyll concentration associated with the Agulhas retroflection (Gardner et  
27 al., 2006; compare Figures 8 and 9). These elevated carbon burial rates also extend all the way  
28 to the Kerguelen Islands. This is in contrast to the low carbon burial rates ( $< 5 \text{ mmolC m}^{-2} \text{ yr}^{-1}$ ;  
29 Janke, 1996; Figure 9) and low  $^{234}\text{Th}$ -estimated POC export fluxes (from 0.10 to 2.53  $\text{mmolC m}^{-2}$   
30  $\text{d}^{-1}$  below 100m) that are observed slightly further south (Coppola et al., 2005).

### 31 32 **3.4.5 Southern subtropical gyre**

33  
34 Near surface ( $< 50$  meters) DOC concentrations in the southern subtropical gyre of the Indian  
35 Ocean are significantly elevated ( $> 75 \mu\text{molC/kg}$ ) between  $10^\circ$  and  $35^\circ$  S (Figure 7), consistent  
36 with the positive correlation between DOC concentrations and temperature, i.e., the southern  
37 subtropical gyre is warm and stratified and DOC concentrations become elevated because there  
38 is reduced vertical mixing. Near-surface DOC concentrations decline precipitously to  $< 50$   
39  $\mu\text{molC/kg}$  south of  $40^\circ$  S due to increased vertical mixing in the higher latitude waters (Figure  
40 7). There is also evidence of subduction and northward transport of DOC in the subtropical  
41 front between  $10^\circ$  and  $35^\circ$  S where elevated DOC concentrations can be seen extending to  $> 250$   
42 meters depth (Figure 7). The source of this subducted water has DOC concentrations between  
43 65 and  $70 \mu\text{molC/kg}$ , and as this water moves downward and equatorward the DOC is  
44 remineralized, dropping to  $\sim 55 \mu\text{molC/kg}$  at 250 meters depth (Hansell, 2009; Figure 7). DOC  
45 concentrations below 500 meters in the southern subtropical Indian Ocean are also low ( $< 50$   
46  $\mu\text{molC/kg}$ ) and relatively uniform (Figure 7).

1  
2 The lowest POC concentrations ( $< 48 \text{ mgC/m}^3$  equal to  $< 4 \text{ umolC/kg}$ ) in the Indian Ocean are  
3 observed in the oligotrophic southern subtropical gyre. These low POC values are associated  
4 with low chlorophyll concentrations ( $< .3\%$  of POC; Figure 8, Gardner et al., 2006). These low  
5 POC and low percent chlorophyll waters dramatically increase in area during the Southern  
6 Hemisphere summer due to increases in summer stratification and lower nutrient availability  
7 (Gardner et al., 2006; Figure 8). Nonetheless, subsurface chlorophyll maxima exist below these  
8 oligotrophic regions (see Hood et al., 2022, this volume). It is likely that the subsurface  
9 chlorophyll maxima are associated with somewhat shallower subsurface POC maxima, with the  
10 latter weaker as discussed above.

11  
12 Sediment trap-measured POC fluxes in the southern subtropical gyre of the Indian Ocean are  
13 some of the lowest recorded worldwide ( $\sim 0.50 \text{ mgC m}^{-2} \text{ day}^{-1} = 0.04 \text{ mmolC m}^{-2} \text{ d}^{-1}$ , measured at  
14 500–600 and 2600–3500 meters; Harms et al., 2021). Low POC fluxes are consistent with the  
15 extremely low organic carbon burial ( $< 2 \text{ mmolC m}^{-2} \text{ yr}^{-1}$ ) and sediment accumulation rates  
16 calculated by Janke (1996; Figure 9). These low POC fluxes are the result of the strongly  
17 stratified, nutrient-depleted and low productivity near-surface waters in the gyre (Harms et al.,  
18 2021; Hood et al., 2022; this volume). These continuously oligotrophic conditions result in an  
19 almost constant POC fluxes in space and time. The lack of seasonality in the POC fluxes can also  
20 be attributed to intense organic matter degradation in the water column (Harms et al., 2021). The  
21 small amount of variability that is observed is related to variations wind-induced physical mixing  
22 events and the passage of eddies. Preliminary estimates indicate that the average POC export  
23 efficiency is extremely low ( $\sim 0.03\%$ ) in these waters (Harms et al., 2021).

## 24 25 **5. Summary and conclusions**

26  
27 The thickest OMZ in the world is found in the northern Indian Ocean in the Arabian Sea where  
28 intermediate water ( $\sim 200\text{-}800 \text{ m}$ ) oxygen concentrations decline to nearly zero, with consequent  
29 impacts on nitrogen cycling. These impacts include denitrification-driven reductions in deep  
30  $\text{NO}_3^-$  concentrations, appearance of  $\text{NO}_2^-$  maxima, generation of greenhouse gases ( $\text{N}_2\text{O}$ ) and  
31 globally-significant losses of fixed nitrogen from the ocean. In contrast, these biogeochemical  
32 impacts are not observed in the Bay of Bengal where intermediate water dissolved oxygen  
33 concentrations are poised just above the threshold below which denitrification becomes  
34 significant. The low oxygen conditions in the water column in both the Arabian Sea and the Bay  
35 of Bengal are associated with slow ventilation and high biological oxygen demand. It appears  
36 that a weaker ballast effect and higher biological productivity sustain higher biological oxygen  
37 consumption in the Arabian Sea compared to the Bay of Bengal, which balances the higher  
38 physical oxygen supply in the Arabian Sea and explains its more intense OMZ.

39  
40 The volume of hypoxic waters and denitrification in the Arabian Sea strongly increase in  
41 response to increases in monsoon winds. Such increases in Arabian Sea denitrification are  
42 expected to cause increases in  $\text{N}_2$  and  $\text{N}_2\text{O}$  production. However, global syntheses of long term  
43 OMZ variability reveal only a weak decrease of dissolved oxygen concentrations in the OMZs of  
44 the Arabian Sea and the Bay of Bengal in comparison to OMZs of the South Atlantic Ocean and  
45 the Pacific Ocean. Moreover, outbreaks of hydrogen sulfide have so far not been reported in the  
46 northern Indian Ocean during the last 50 years, other than in bottom waters on the Indian shelf.

1 The absence of H<sub>2</sub>S implies that the interplay between physical oxygen supply and the biological  
2 oxygen consumption has largely maintained the current hypoxic/anoxic conditions in the  
3 Arabian Sea and the Bay of Bengal OMZs. However, recent observational and modeling studies  
4 indicate that oxygen concentrations are now decreasing in the Arabian Sea and that these  
5 decreases are having significant biogeochemical and ecological impacts. Future evolution of the  
6 northern Indian Ocean OMZs projected by Earth System models is highly uncertain.

7  
8 The Indian Ocean remains one of the most poorly sampled ocean regions with respect to  
9 inorganic and organic carbon pools and air-sea carbon fluxes. The system accounts for ~20% of  
10 the global oceanic uptake of atmospheric CO<sub>2</sub>. The Arabian Sea is a source of CO<sub>2</sub> to the  
11 atmosphere due to elevated pCO<sub>2</sub> within the SWM-driven upwelling whereas it is still uncertain  
12 whether the Bay of Bengal is a CO<sub>2</sub> source or sink due to the sparse spatial and temporal  
13 sampling. South of 14°S the Indian Ocean appears to be a strong net CO<sub>2</sub> sink due to the  
14 combined effects of both the solubility pump and the biological pump. Surface pH values in the  
15 Indian Ocean are among the lowest of the major ocean basins, and the reasons for this are poorly  
16 understood. Increases in sulphate and nitrogen aerosol loadings over the Bay of Bengal may be  
17 mainly responsible for the increased acidity in the northwestern Bay of Bengal in recent years  
18 and reduced river discharge together with a positive IOD event have also been shown to  
19 contribute to enhanced acidification and pCO<sub>2</sub> levels in the coastal waters in the western Bay of  
20 Bengal. Projected increases in oceanic CO<sub>2</sub> concentrations will lead to further acidification of the  
21 Indian Ocean over the coming decades, with potentially severe negative impacts on coral reefs  
22 and other calcifying organisms.

23  
24 DOC concentrations in the Indian Ocean vary between ~40 and 80 μmolC/kg. DOC  
25 concentrations in the Indian Ocean tend to be high in stratified near-surface tropical and  
26 subtropical waters where DOC is produced and accumulates, and DOC concentrations are lowest  
27 in the deep ocean where heterotrophic consumption of DOC exceeds autotrophic production. As  
28 observed elsewhere in the global ocean, satellite-estimated near-surface POC concentrations are  
29 elevated in coastal regions of the Indian Ocean with values often exceeding 120 mgC/m<sup>3</sup> (= 10  
30 μmolC/kg). The highest POC concentrations in the Indian Ocean are observed in the  
31 northwestern part of the basin in the Arabian Sea and off the coast of Somalia with values  
32 estimated to be greater than 96 mgC/m<sup>3</sup> (= 8 μmolC/kg). These high POC values are associated  
33 with high chlorophyll concentrations. The lowest POC concentrations (< 48 mgC/m<sup>3</sup> = 4  
34 μmolC/kg) are observed in the oligotrophic southern subtropical gyre. These low POC values  
35 are associated with low chlorophyll concentrations.

36  
37 The spatial and temporal variability in POC export flux in the Indian Ocean is similar to the  
38 above mentioned patterns in POC and chlorophyll concentration, consistent with primary  
39 production as the main control on the spatial and temporal variability of organic carbon fluxes.  
40 However, in river-influence regions, like the Bay of Bengal, the spatial variability of organic  
41 carbon flux is also strongly influenced by lithogenic matter content. In the Arabian Sea POC  
42 export at 2000 meters ranges from < 6 to > 22 mgC m<sup>-2</sup> d<sup>-1</sup> (< 0.5 to > 1.8 mmolC m<sup>-2</sup> d<sup>-1</sup>) with  
43 the highest fluxes observed during the SWM. In contrast, POC export at 2000 meters in the Bay  
44 of Bengal is generally lower and less strongly seasonal than in the Arabian Sea with fluxes  
45 varying between 2 and 14 mgC m<sup>-2</sup> d<sup>-1</sup> (0.17 to 1.1 mmolC m<sup>-2</sup> d<sup>-1</sup>). <sup>234</sup>Th-estimates of POC  
46 export flux have revealed elevated values in eastern-central equatorial waters of the Indian

1 Ocean (up to  $7.7 \text{ mmolC m}^{-2} \text{ d}^{-1}$ ) in association with elevated rates of primary production. These  
2 elevated rates are surprising given the downwelling equatorial circulation, but they are consistent  
3 with zonal variations in satellite and model-based estimates of near surface POC export flux and  
4 organic carbon burial rates. There are few direct measurements of POC concentrations and  
5 export flux off of Western Australia where estimated organic carbon burial rates are distinctly  
6 elevated, perhaps due to the influence of Leeuwin Current eddies. Estimated organic carbon  
7 burial rates in the southwestern Indian Ocean range from  $< 5 - \sim 10 \text{ mmolC m}^{-2} \text{ yr}^{-1}$ , with the  
8 highest rates observed along the coastal zone of southeastern Africa and in the Agulhas  
9 retroflection. POC export fluxes in the southern subtropical gyre of the Indian Ocean are some  
10 of the lowest recorded worldwide ( $\sim 0.50 \text{ mgC m}^{-2} \text{ d}^{-1} = 0.04 \text{ mmolC m}^{-2} \text{ d}^{-1}$ ).

11  
12 As in the other ocean basins, it is clear that there is a strong connection in the Indian Ocean  
13 between the physics that drives (or suppresses) nutrient delivery to the photic zone and responses  
14 of oxygen,  $\text{CO}_2$  flux, pH, DOC, POC and POC export. Monsoonal wind forcing is a major driver  
15 of biogeochemical variability throughout the northern Indian Ocean and in equatorial waters. In  
16 addition, there are regionally specific processes that significantly modulate oxygen,  $\text{CO}_2$  flux,  
17 pH, DOC, POC and POC export. For example, upwelling and strong advective impacts in the  
18 Arabian Sea; freshwater and stratification in the Bay of Bengal; the influence of ITF, poleward  
19 transport, downwelling and eddies in the southeastern Indian Ocean; and poleward transport of  
20 tropical waters, combined with localized upwelling in the southwestern Indian Ocean. The  
21 southern subtropical gyre is extremely oligotrophic.

22  
23 Observational and modeling research should be aimed at improving understanding of northern  
24 Indian Ocean OMZ variability at seasonal and decadal time scales. Further, uncertainty in future  
25 projections needs to be reduced. Additional observations are also needed to better constrain air-  
26 sea  $\text{CO}_2$  fluxes and the carbon cycle in general. Measurements of  $\text{CO}_2$  and DOC concentrations,  
27 capturing spatial and temporal variability, are particularly sparse. This improved understanding  
28 is needed to predict the impacts of anthropogenic influence and global warming on Indian Ocean  
29 biogeochemistry and ecosystems and also for understanding the role of the Indian Ocean in  
30 global ocean biogeochemical cycles both now and in the future.

## 31 32 **6. Educational Resources**

33  
34 -Ocean Data View, free software for plotting oceanographic data. Available at:

35 <https://odv.awi.de>

36  
37 -World Ocean Atlas, a collection of objectively analyzed, quality controlled temperature,  
38 salinity, oxygen, phosphate, silicate, and nitrate means based on profile data from the World  
39 Ocean Database. Available at: <https://www.ncei.noaa.gov/products/world-ocean-atlas>

40  
41 -Surface Ocean  $\text{CO}_2$  Atlas (SOCAT) is a synthesis of quality-controlled, surface ocean  $f\text{CO}_2$   
42 (fugacity of carbon dioxide) observations by the international marine carbon research  
43 community. Available at: <https://www.socat.info>

44  
45 -Satellite ocean color data. Available at: <https://oceancolor.gsfc.nasa.gov>

46

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## 13 **Author contributions**

14  
15 All authors contributed to writing of the text, discussion of content, and structure  
16 of the chapter.

## 18 **Index Terms**

19  
20 Indian Ocean, Arabian Sea, Bay of Bengal, Indonesian Throughflow, Leeuwin  
21 Current, Mozambique Channel, Agulhas Current, Indian Ocean southern  
22 subtropical gyre, oxygen, oxygen minimum zone, salinity, stratification,  
23 denitrification, N<sub>2</sub>O, CO<sub>2</sub>, pH, dissolved organic carbon (DOC), particulate organic  
24 carbon (POC), carbon export flux.

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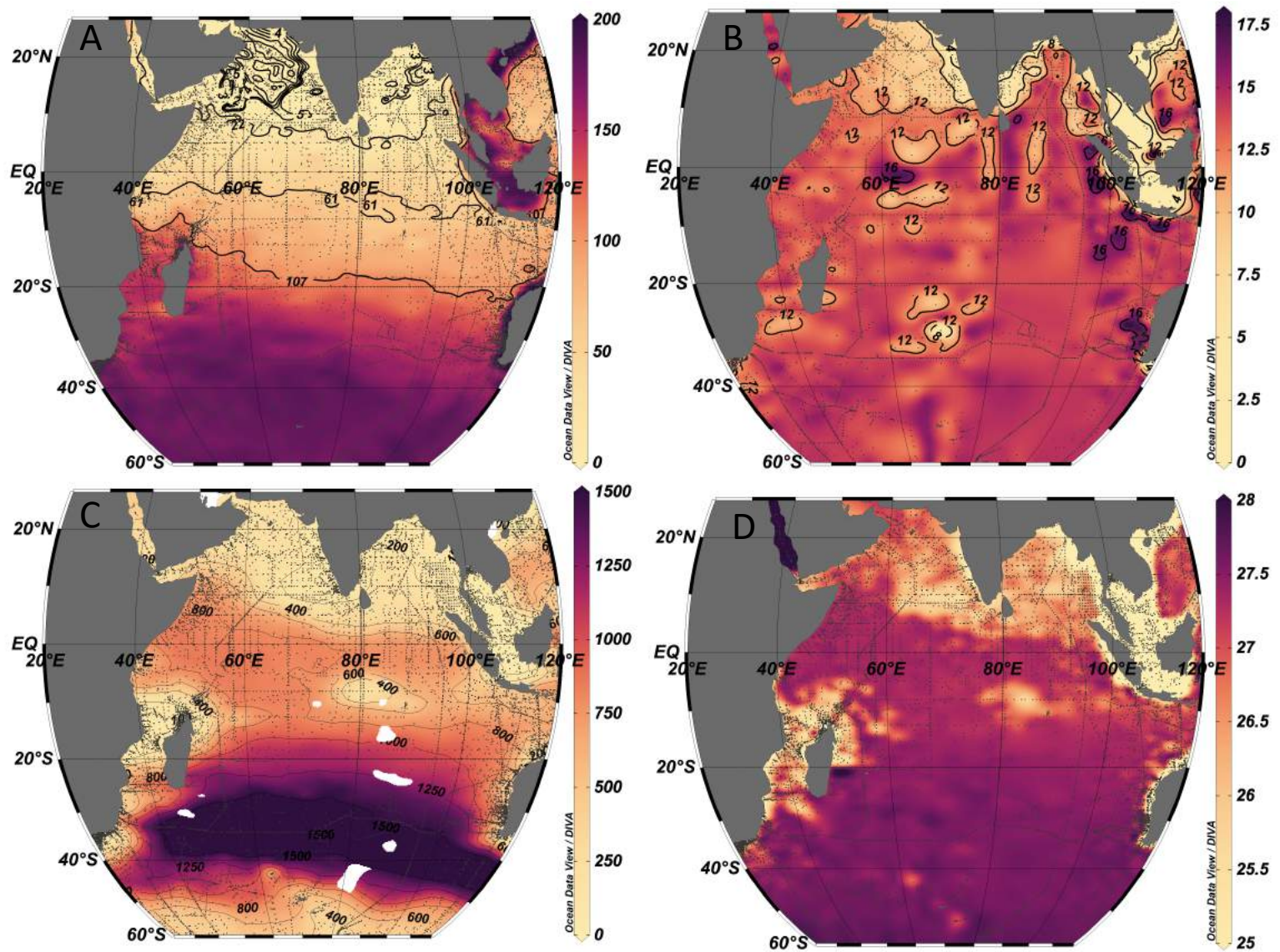
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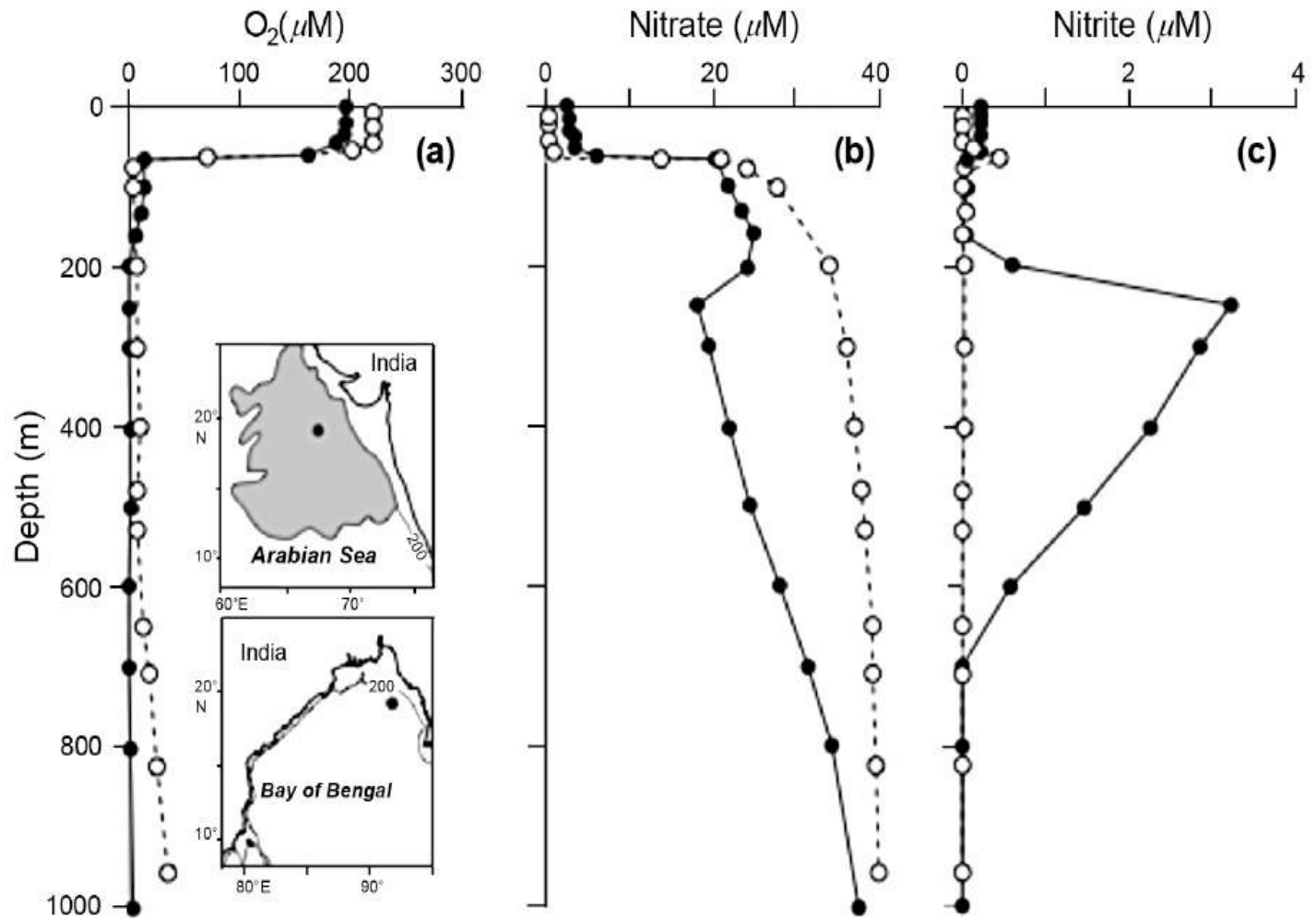
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**Table 1: Definitions, thresholds, and impacts of different hypoxia thresholds from Hofmann et al 2011.**

Terminology	Threshold for impacts to occur maybe in different units	Impacts	Indian Ocean Regions with oxygen minimums below threshold	References
Mild Hypoxia	107 $\mu\text{mol/kg}$ , 109 $\mu\text{M}$ , 3.5 $\text{mgO}_2/\text{l}$ , ~53 % saturation,	Sensitive species show avoidance reactions	Eastern and Central Indian Ocean south of 20S, Western Indian Ocean south of 10S	Cenr 2010
Hypoxia	61 $\mu\text{mol/kg}$ , 63 $\mu\text{M}$ , 2 $\text{mgO}_2/\text{l}$ , ~30 % saturation,	Fishes and the majority of higher organisms suffer from oxygen deficiency, ecosystem adapted to low oxygen conditions	Arabian Sea, Bay of Bengal	Ekau et al., 2010; Vaquer-Sunyer and Duarte 2008
Microbial hypoxia, suboxic, Severe hypoxia, typical OMZ definition.	22 $\mu\text{mol/kg}$ , 22 $\mu\text{M}$ , 0.71 $\text{mgO}_2/\text{l}$ , ~11 % saturation	Microbes start to experience the toxic effects of oxygen on anaerobic processes, only highly specialized species survive	Northern tropical Indian Ocean	Rixen et al., 2020, Diaz and Rosenberg (2008)
Functional anoxia	0.05 $\mu\text{M}$	Anaerobic microbial processes dominate, nitrite accumulation	Central and Eastern Arabian Sea.	Rixen et al., 2020, Morrison et al., 1999, Ulloa et al, 2012

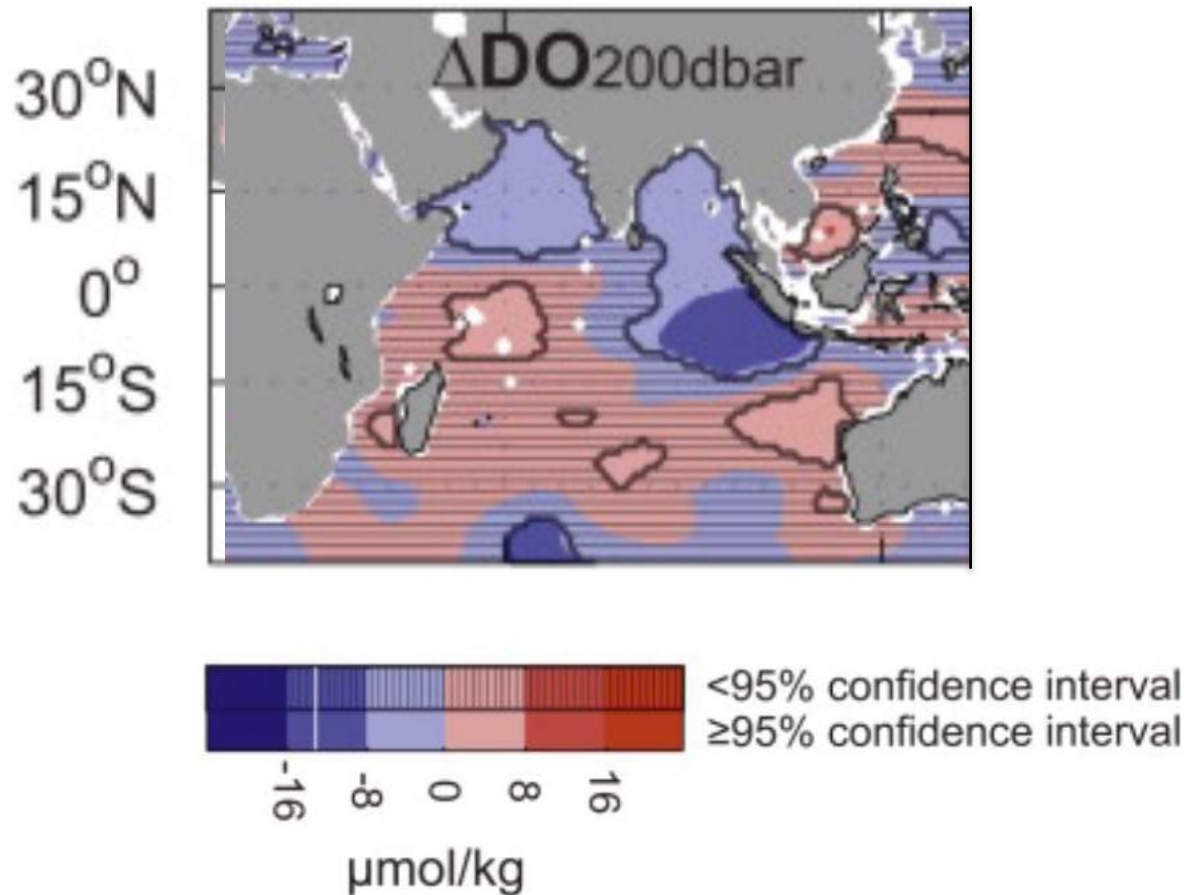


**Figure 1:** Spatial maps of the A) oxygen concentration (umol/kg) at the depth of the oxygen minimum, B) nitrate to phosphorus ratio at the depth of the oxygen minimum, C) the depth (m) of the oxygen minimum, D) the potential density anomaly (kg/m<sup>3</sup>) at the depth of the oxygen minimum.

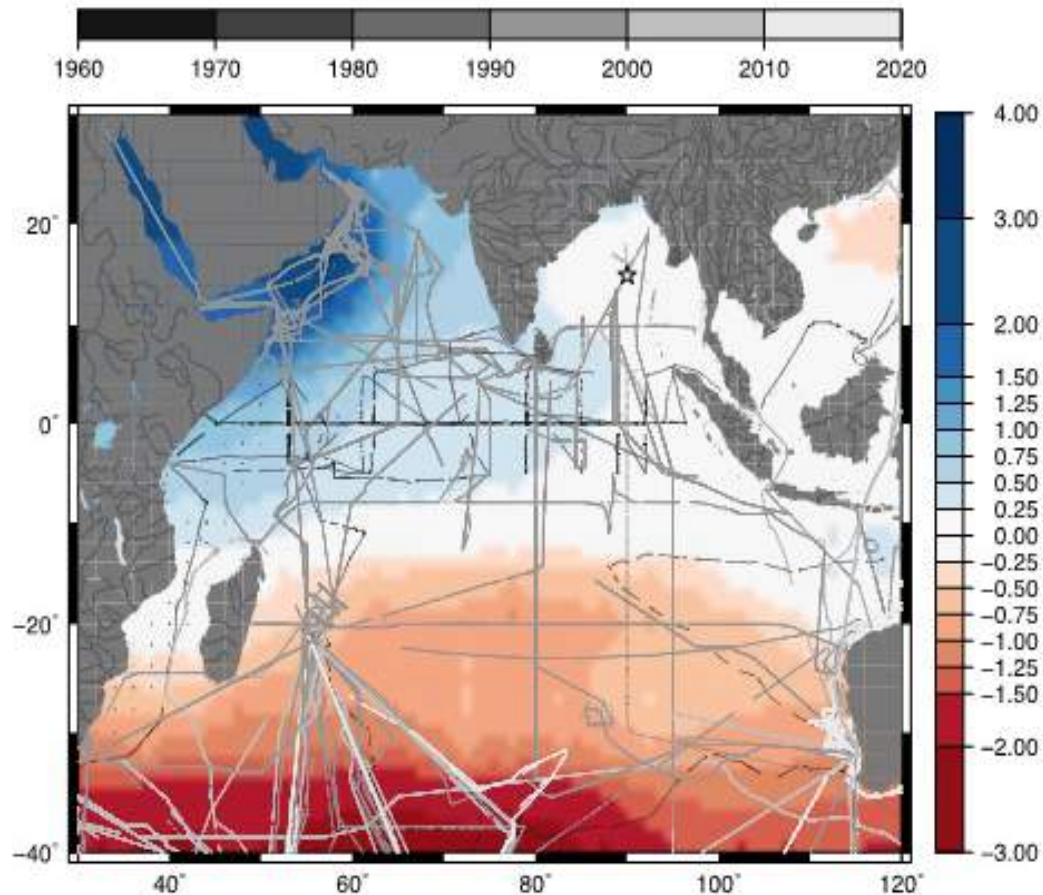


**Figure 2:** Comparison of vertical profiles of (a) oxygen, (b) nitrate and (c) nitrite in the Arabian Sea (filled circles) and Bay of Bengal (open circles). Station locations are shown in insets. The Arabian Sea inset also shows limit of the denitrification zone to the eastern-central basin. Figure reproduced from Naqvi et al. (2006).

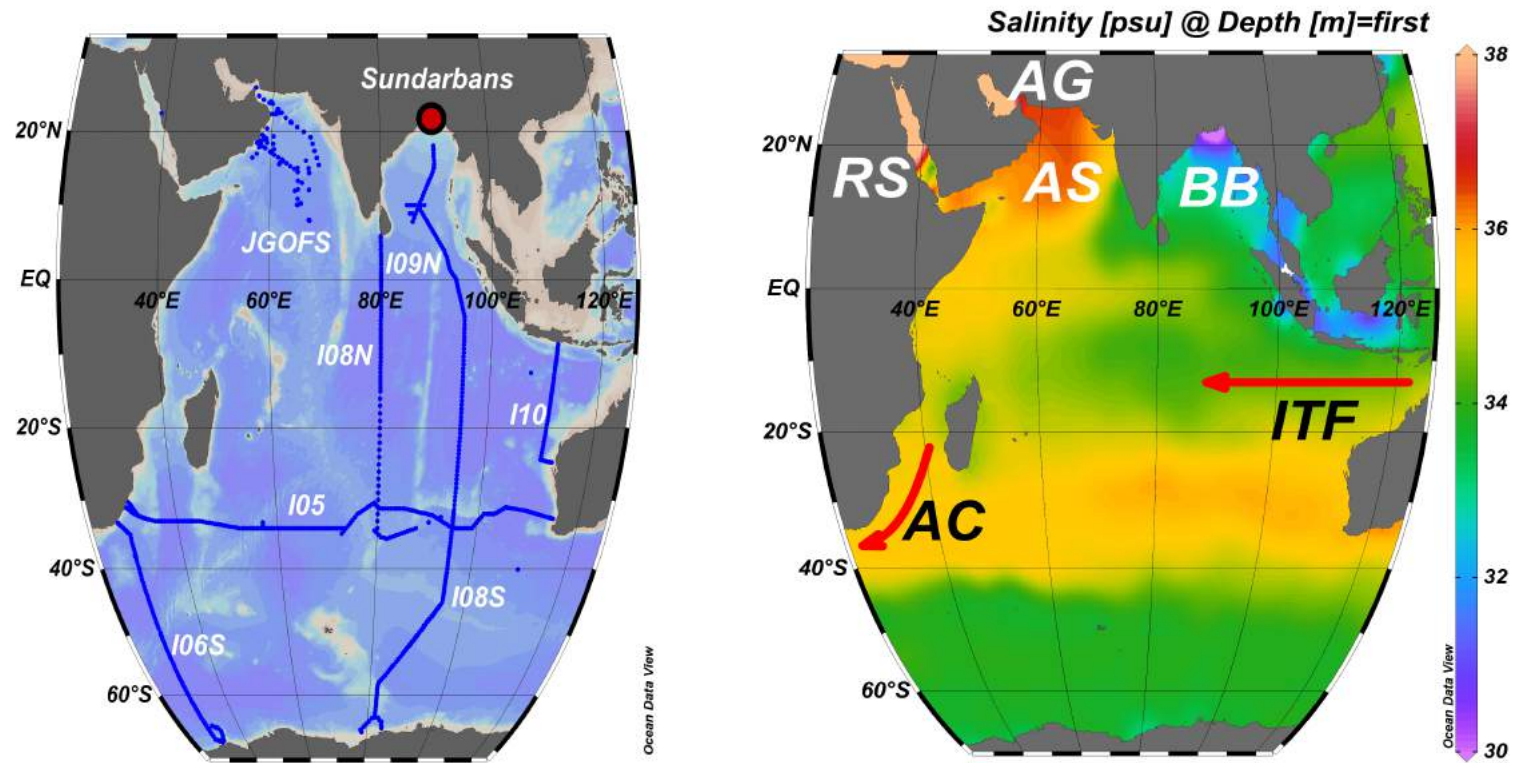




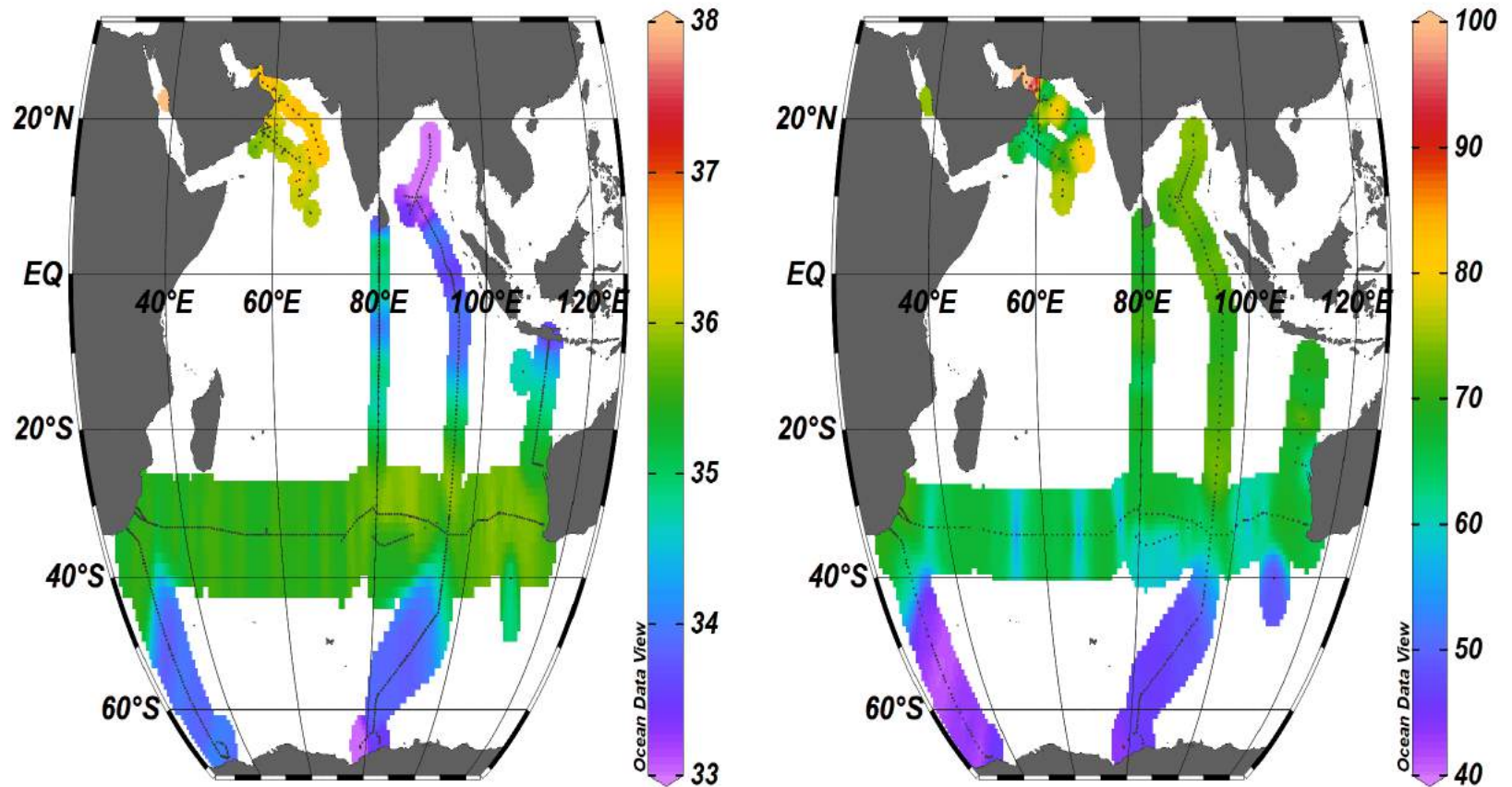
**Figure 3:** Changes in dissolved oxygen ( DO) ( $\mu mol kg^{-1}$  ) at 200 dbar between 1960–1974 and 1990–2008 in the Indian Ocean. Increases (decreases) in DO are indicated in red (blue). Areas with differences below the 95% confidence interval are shaded by black horizontal lines. Figure modified from Stramma et al. (2010).



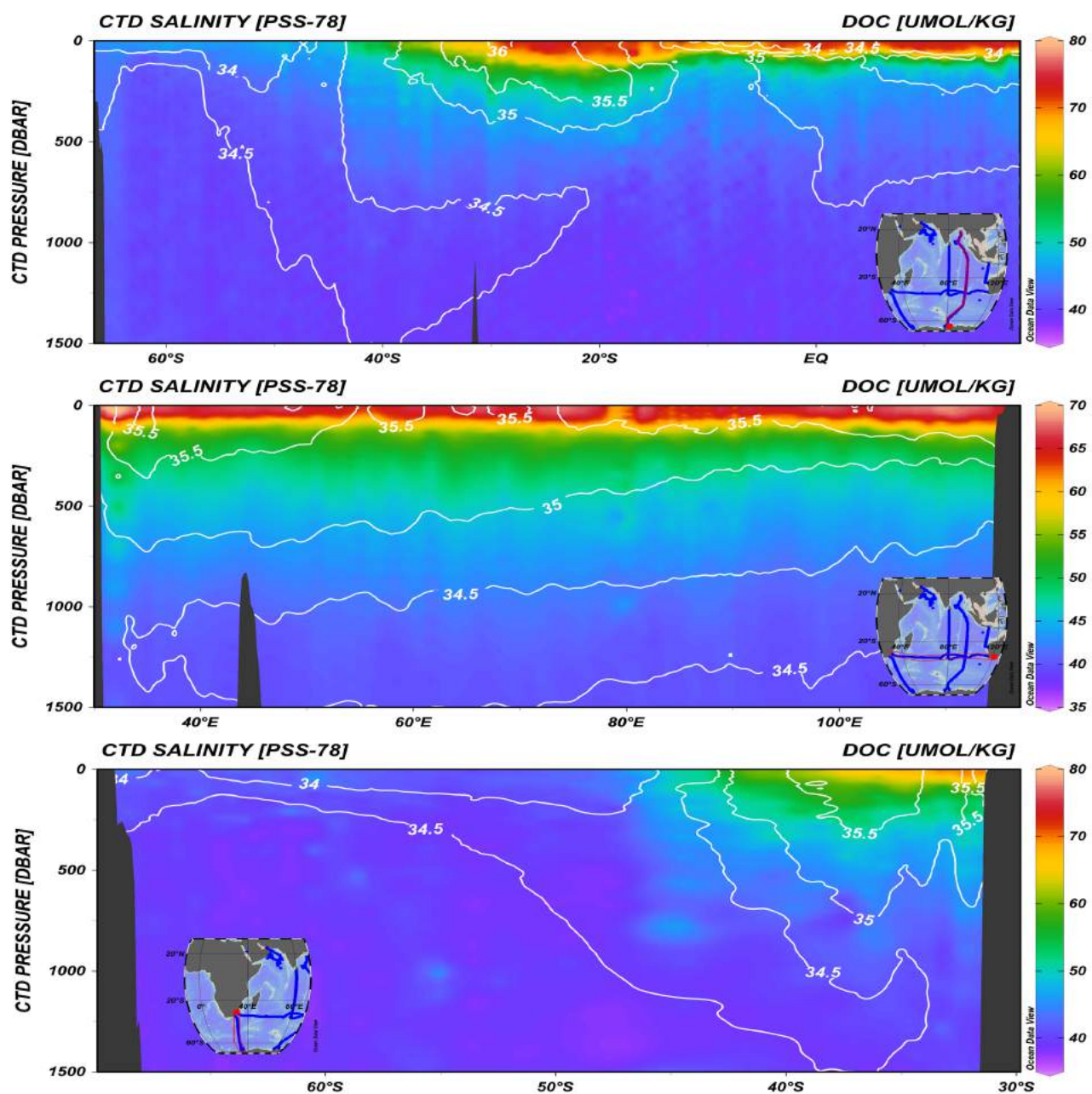
**Figure 4:** Annual CO<sub>2</sub> flux (mol C m<sup>-2</sup> yr<sup>-1</sup>) referenced to the year 2000 (Takahashi et al., 2009) over the Indian Ocean. Data points colored by year of collection (Takahashi & Sutherland, 2016) are overlaid as points. Major rivers are also delineated over the continents. The star in the central Bay of Bengal shows the location of a mooring that has been collecting continuous CO<sub>2</sub> and pH measurements since November, 2013. Figure and caption from Hood et al. (2019).



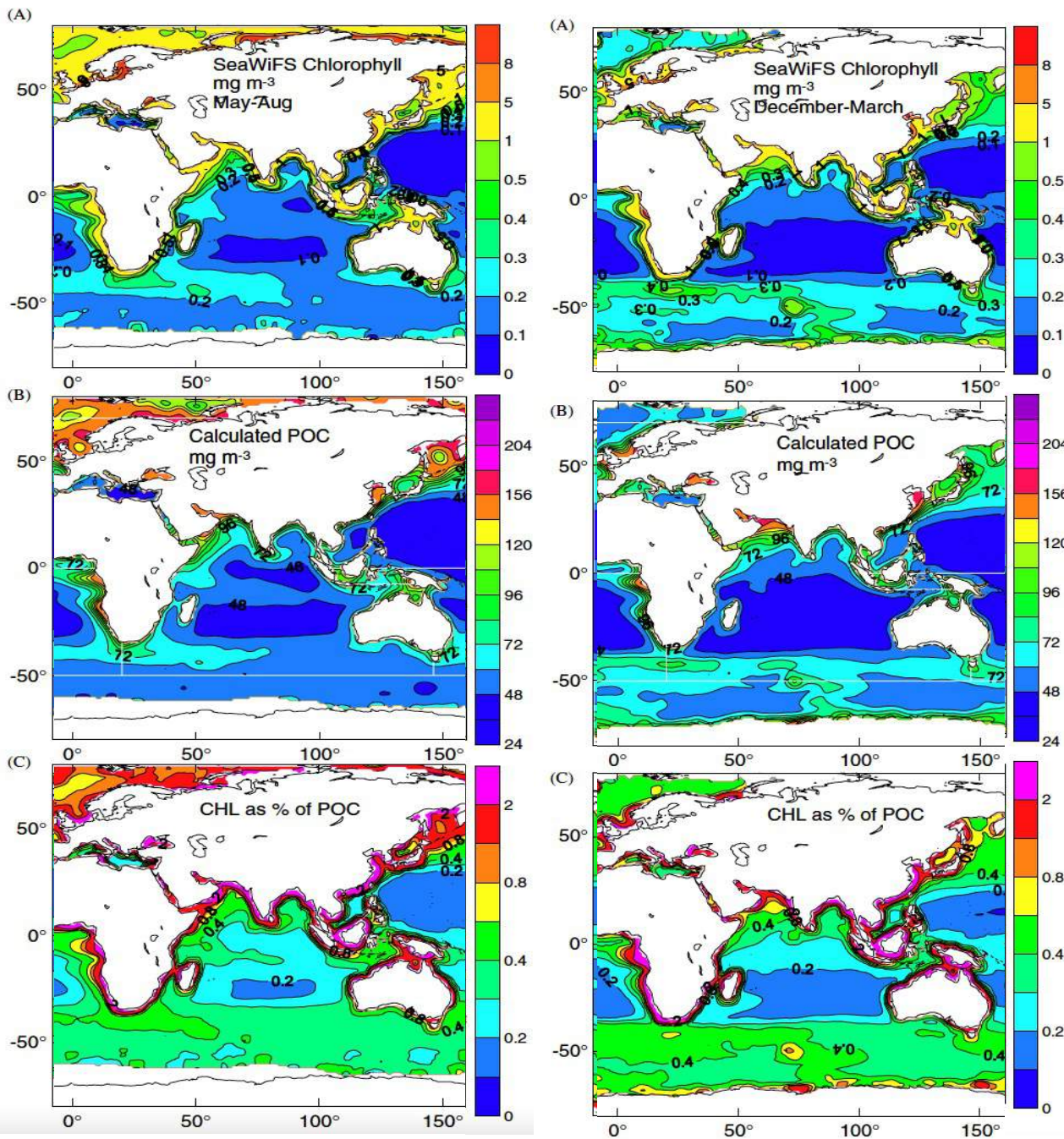
**Figure 5:** Map of the Indian Ocean showing oceanographic sections discussed (left) and gridded surface salinity (right) during austral summer (WOA), with transports associated with the Indonesian Throughflow (ITF) and Agulhas Current (AC) indicated. Relevant basins are labeled: Arabian Sea (AS); Bay of Bengal (BB); Red Sea (RS); Arabian Gulf (AG), also known as the Persian Gulf.



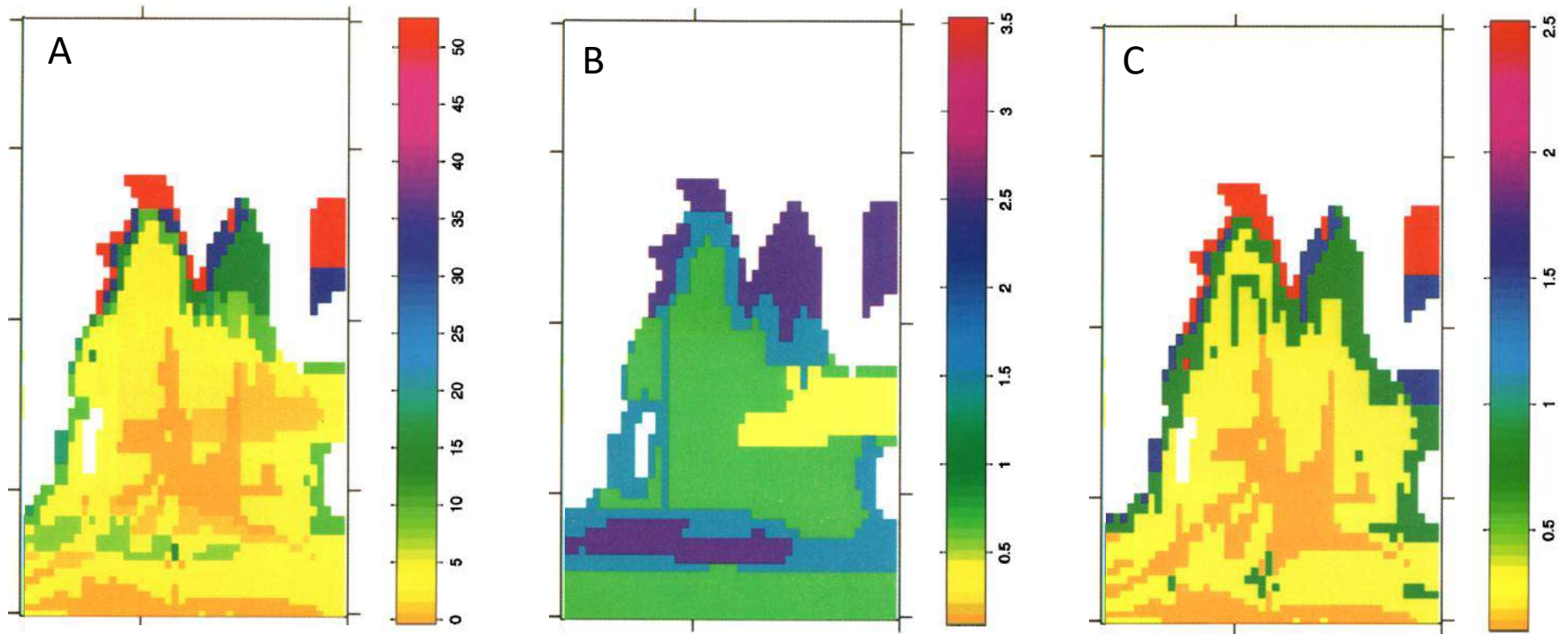
**Figure 6:** Surface distributions of salinity (left) and DOC ( $\mu\text{molC/kg}$ ; right) observed during cruises. Data from Hansell et al. (2021).



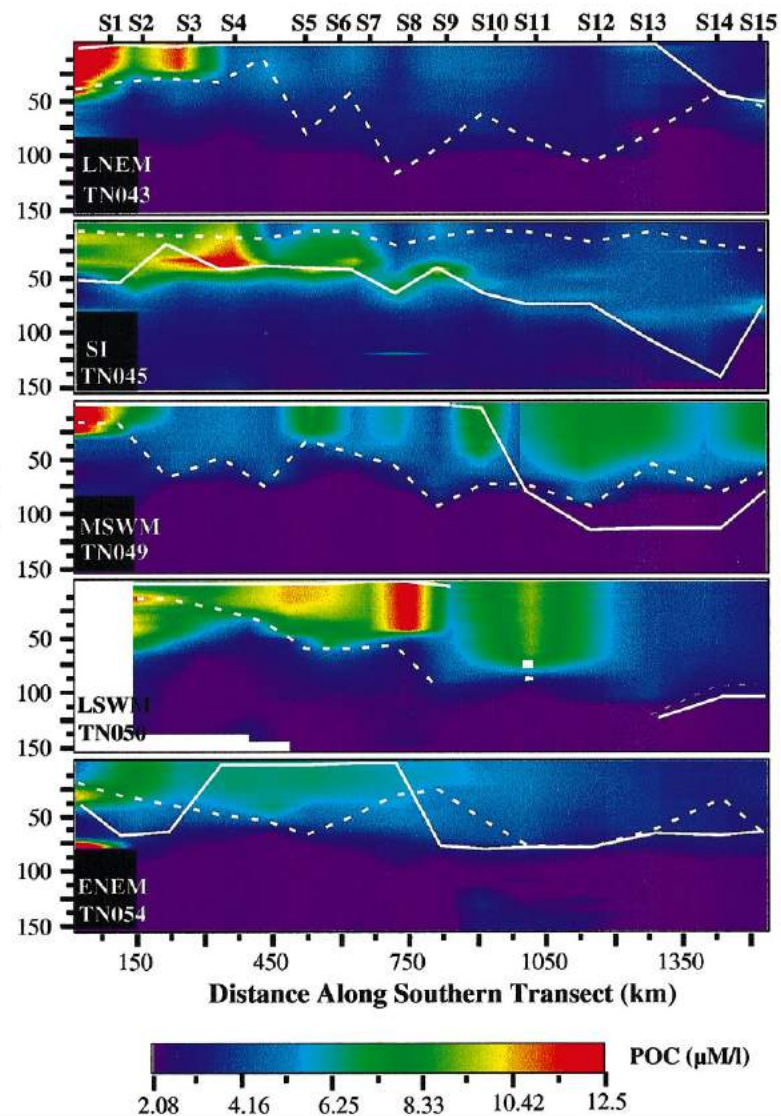
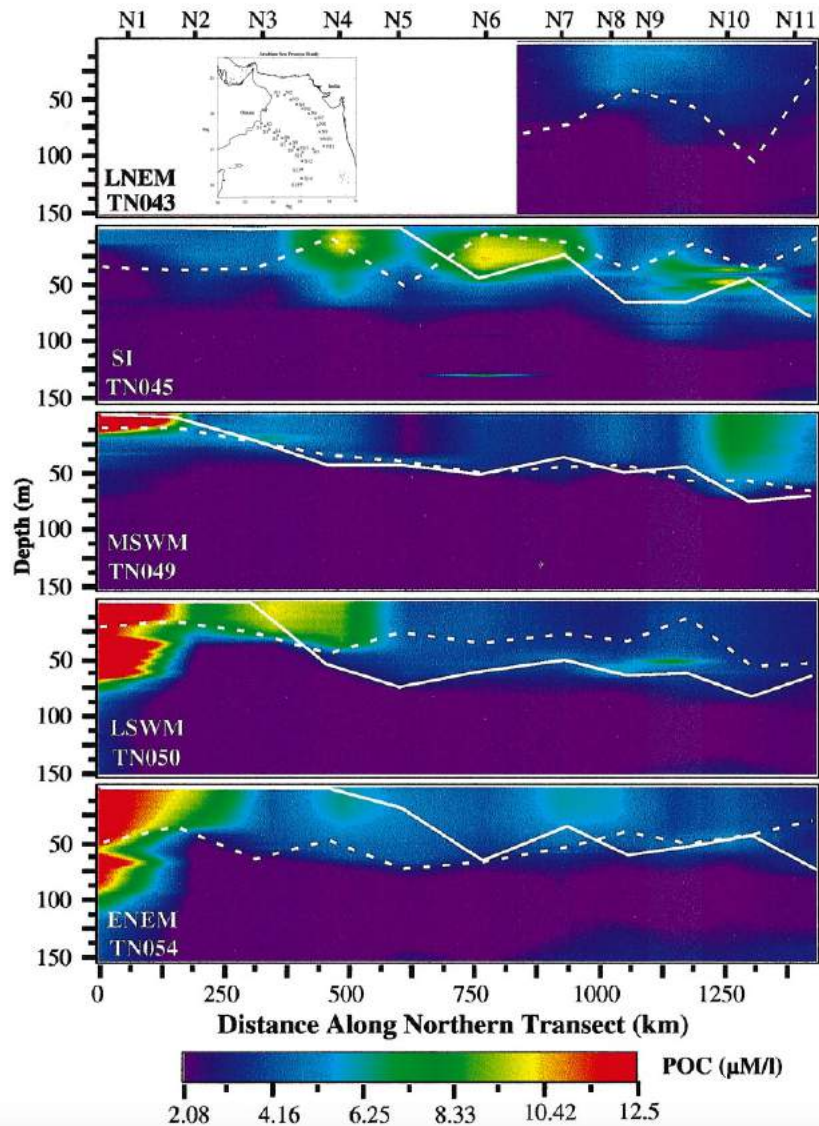
**Figure 7:** Upper 1500 m section plot of DOC (color;  $\mu\text{molC/kg}$ ) and salinity (contours) along I09N/I08S (top panel, occupied in 2016), I05 (middle panel, occupied in 2009) and I06S (bottom panel, occupied in 2019). Data from Hansell et al. (2021).



**Figure 8:** Distribution of: (A) SeaWiFS CHL ( $\text{mg m}^{-3}$ , level 3, reprocessing 4 data); (B) average POC ( $\text{mgC m}^{-3}$ ) over one attenuation depth calculated from  $K_{490}:c_p:\text{POC}$ ; (C) CHL as a % of POC for: (Left Panels) summer season (1997–2002, May–August, 20 months); (Right Panels) winter season (1997–2002, December–March, 20 months). White lines in (B) mark boundaries separating ocean basins as used by Behrenfeld and Falkowski (1997). Figure and caption modified from Gardner et al. (2006).

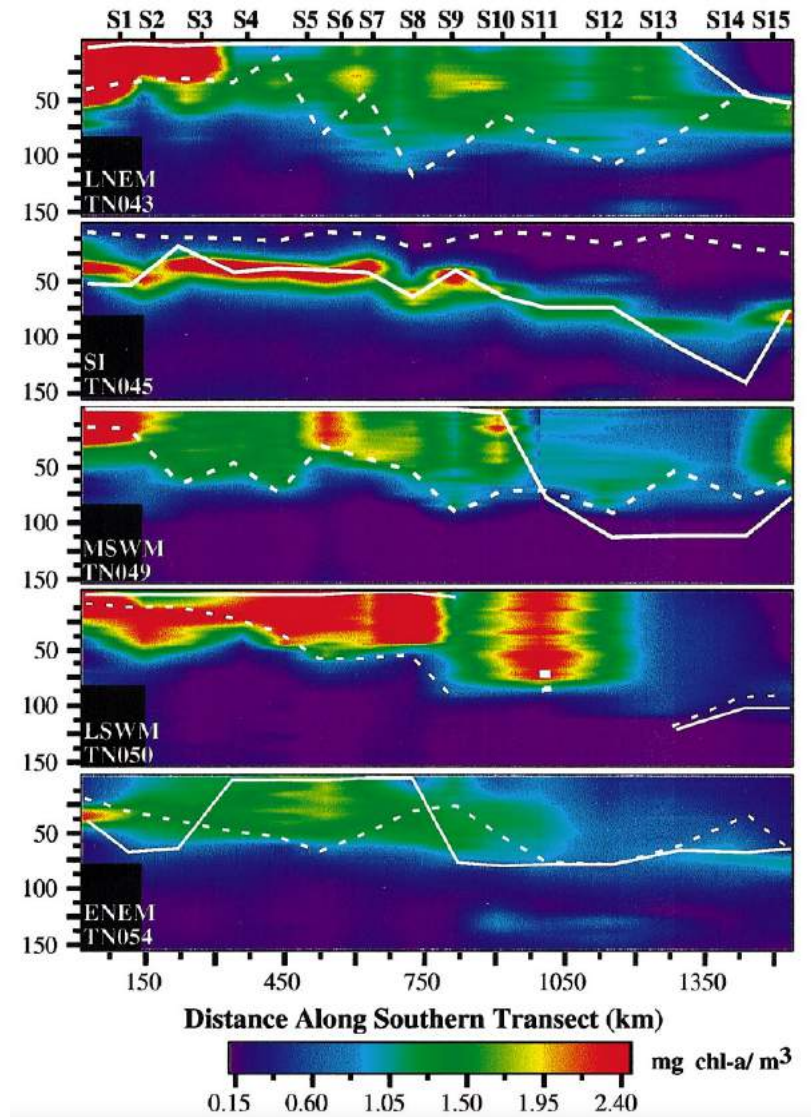
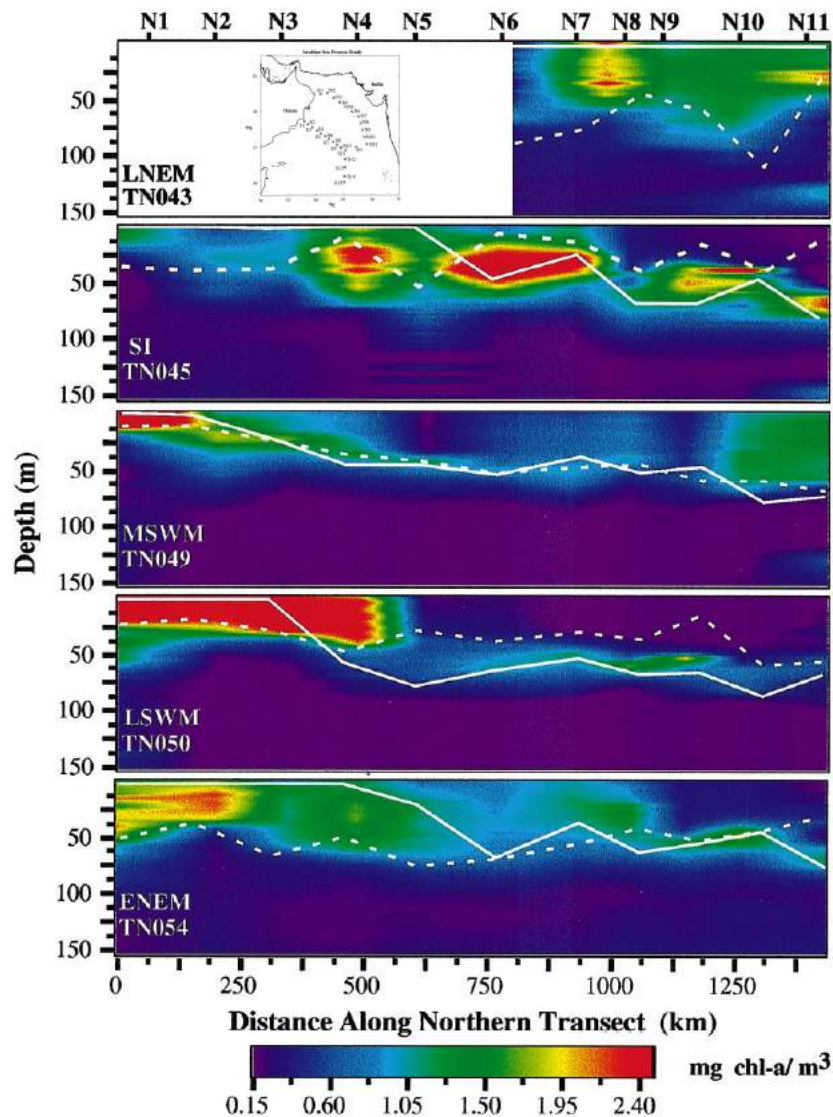


**Figure 9:** Estimated A) organic carbon burial rate ( $\text{mmolC m}^{-2} \text{yr}^{-1}$ ); B) sediment accumulation rate (grams per square centimeter per thousand years); and C) seafloor distribution of organic carbon (weight percent). Figures and caption modified from Jahnke (1996).

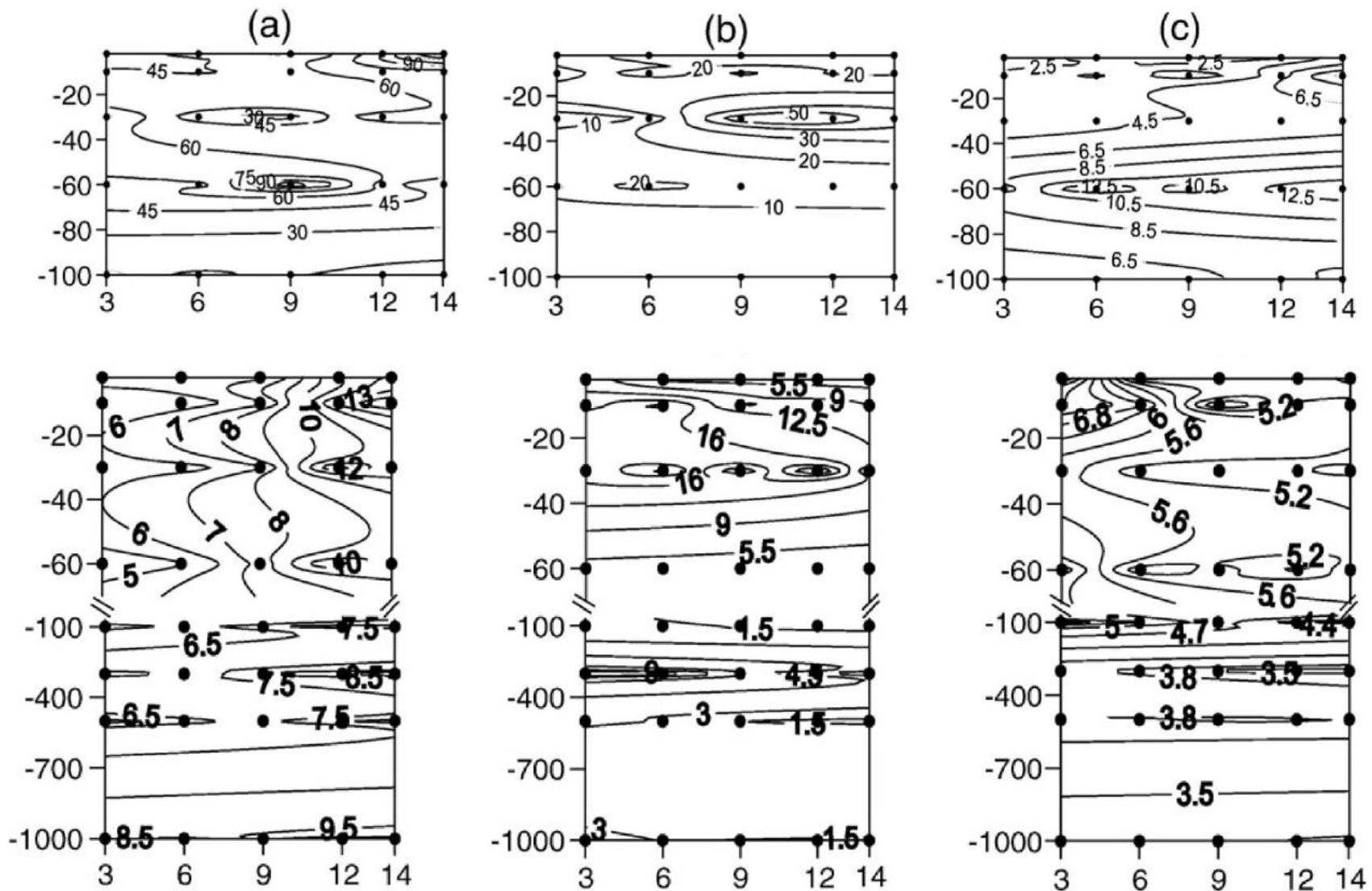


**Figure 10:** POC sections of the northern transect (left panels) and southern transect (right panels) for each cruise. Mean MLD [ $\Delta \sigma = 0.03$ ] (dotted line) and  $0.5 \mu\text{M/l}$  nitrate isopleth (solid line) are shown for each cruise. The station locations are shown in the map inset in the upper left panel. Figures and caption modified from Gunderson et al. (1998).

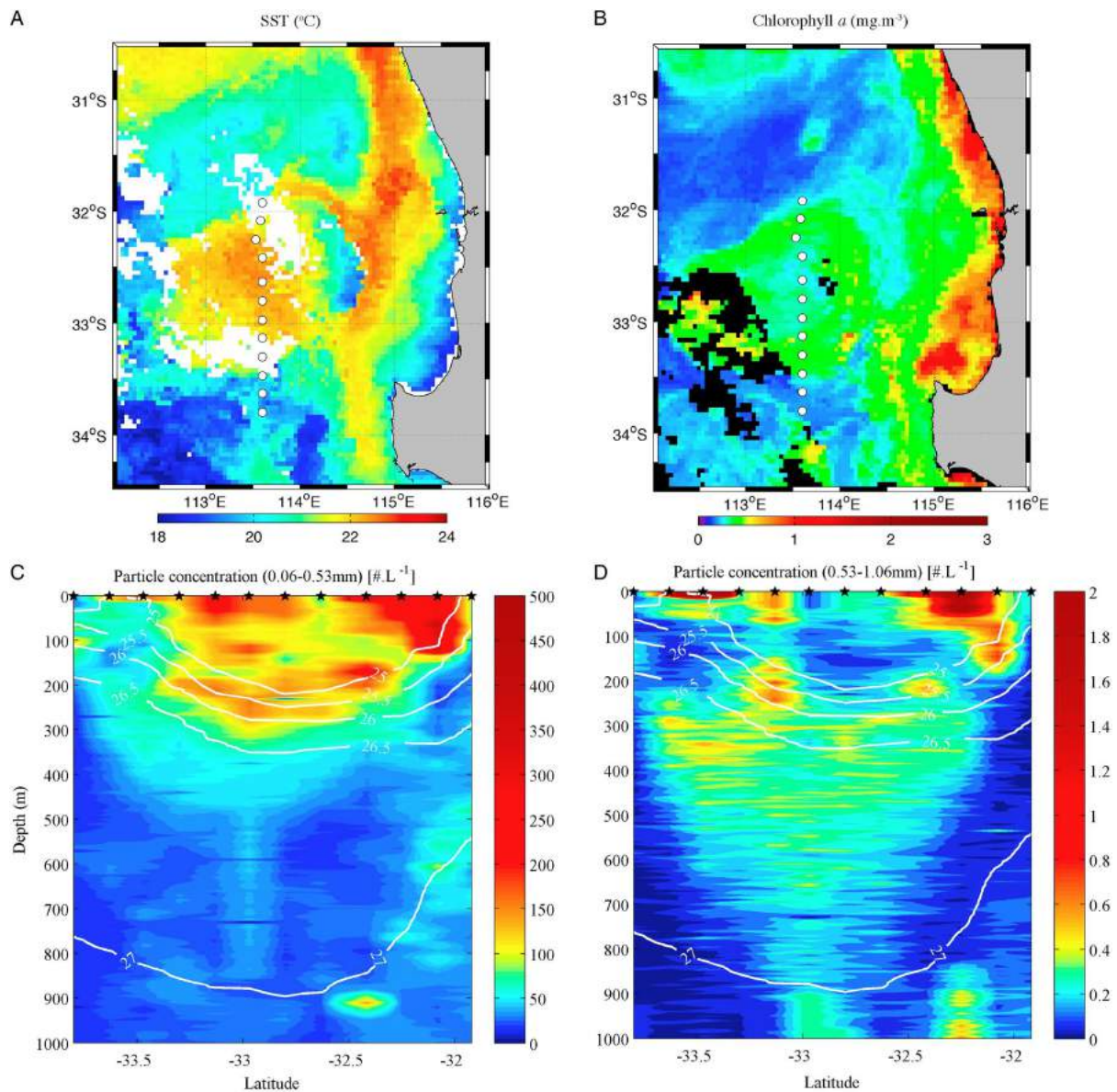




**Figure 11:** Chlorophyll sections of the northern transect (left panels) and southern transect (right panels) for each cruise. Mean MLD [ $\Delta \sigma = 0.03$ ] (dotted line) and 0.5  $\mu\text{M/l}$  nitrate isopleth (solid line) are shown for each cruise. The station locations are shown in the map inset in the upper left panel. Figures and caption modified from Gunderson et al. (1998).



**Figure 12:** Distributions of chlorophyll a concentration (Chl a ng/l, top panels) and particulate organic carbon (POC  $\mu\text{M C}$ , bottom panels) in the central Bay of Bengal during (a) SWM, (b) FIM and (c) SPIM. Figure and caption modified from Fernandes et al. (2009).



**Figure 13:** (a) Sea-surface temperature (°C) and (b) chlorophyll *a* (mg m<sup>-3</sup>) derived from Moderate Resolution Imaging Spectroradiometer (MODIS) satellite, showing the mesoscale eddy in the Leeuwin Current (LC) off Australia. White circles indicate stations sampled. Spatial distributions of particles across the eddy: (c) Small (0.06–0.5 mm) and (d) large (0.5–1 mm) particles shown in color, with isopycnals contoured in white. Figures and caption modified from Waite et al. (2016).