

Introduction to the Indian Ocean[★]

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1 Introduction

The Indian Ocean is a remarkable place. Unlike the Pacific and Atlantic, the Indian Ocean has a low-latitude land boundary to the north and the Indian subcontinent partitions the northern basin (Fig. 1). The partitioning of the northern basin effectively creates two subbasins consisting of the Arabian Sea and the Bay of Bengal, where the differences in evaporation, precipitation, and river runoff give rise to pronounced differences in salinity and stratification between subbasins. As a result of the proximity of the high mountainous terrain of the Eurasian land mass and the heating and cooling of air masses over it, the tropical Indian Ocean is subject to strong monsoonal wind forcing that reverses seasonally (for reviews, see Schott & McCreary, 2001; Ummenhofer et al., 2024a). In the northern hemisphere, these are referred to as the Southwest Monsoon winds (blowing from the southwest toward the northeast during the boreal summer) and the Northeast Monsoon winds (blowing from the northeast toward the southwest in the boreal winter). The Coriolis effect causes the winds to change their direction at the equator, and in the Southern Hemisphere, they become the Southeast Monsoon winds (blowing from the southeast toward the northwest in the austral winter) and the Northwest Monsoon winds (blowing from the northwest toward the southeast in the austral summer), respectively (Fig. 1). These winds, combined with the large differences in salinity and stratification, profoundly impact marine physical, biogeochemical, and ecosystem dynamics in the Arabian Sea and the Bay of Bengal that extend into the southern tropical Indian Ocean (Brewin et al., 2012; Hood et al., 2009, 2017, 2024a, 2024b; Marsac et al., 2024; Phillips et al., 2021; Schott & McCreary, 2001; Sprintall et al., 2024).

The winds along the equator in the Indian Ocean are stronger during the westerlies of the Southwest Monsoon and, therefore, downwelling favorable on average (Schott et al., 2009; Wang & McPhaden, 2017). As a result, the near-surface waters along the equator tend to be oligotrophic, i.e., typically undetectable nitrate, phosphate, and silicate concentrations down to ~100m (see Hood et al., 2024a). Upwelling centers in the Indian Ocean are generally found in off-equatorial regions. Coastal upwelling occurs in the Arabian Sea and the Bay of Bengal in response to the Southwest Monsoon winds. Local Southwest Monsoon-induced wind stress curl drives upwelling in the southern tropical Indian Ocean between ~5–15°S and ~50–80°E at the Seychelles-Chagos Thermocline Ridge (SCTR) (Hermes & Reason, 2008; McPhaden & Nagura, 2014; Nyadjro et al., 2017; Xie et al., 2002; Yokoi et al., 2008) and the east of Sri Lanka between ~6–10°N and ~82–87°E in Sri Lanka Dome (Burns et al., 2017; Schott & McCreary, 2001; Shankar et al., 2002; Vinayachandran & Yamagata, 1998). In addition, upwelling occurs along the coasts of Sumatra, Java, and Bali in Indonesia in response to the Southeast Monsoon winds (Hood et al., 2017, 2024a; Sprintall et al., 1999; Susanto et al., 2001).

In the southeastern tropical Indian Ocean, the Indonesian Throughflow (ITF) connects the Pacific and Indian Ocean basins. The ITF influences water mass properties of the Indian Ocean through exchanges of heat and freshwater (Schott & McCreary, 2001) and biogeochemical properties through exchanges of nutrients (Ayers et al., 2014; Hood et al., 2024a; Sprintall et al., 2024; Talley & Sprintall, 2005) and presumably also planktonic organisms (Hood et al., 2017).

The southern Indian Ocean subtropical gyre is an oligotrophic ocean habitat. To the north, the South Equatorial Current transports warm, nutrient-enriched freshwater from the ITF across the basin (Schott & McCreary, 2001; Sprintall et al., 2024). To the west, the Agulhas Current is an unusually large, poleward-flowing, upwelling-favorable western boundary

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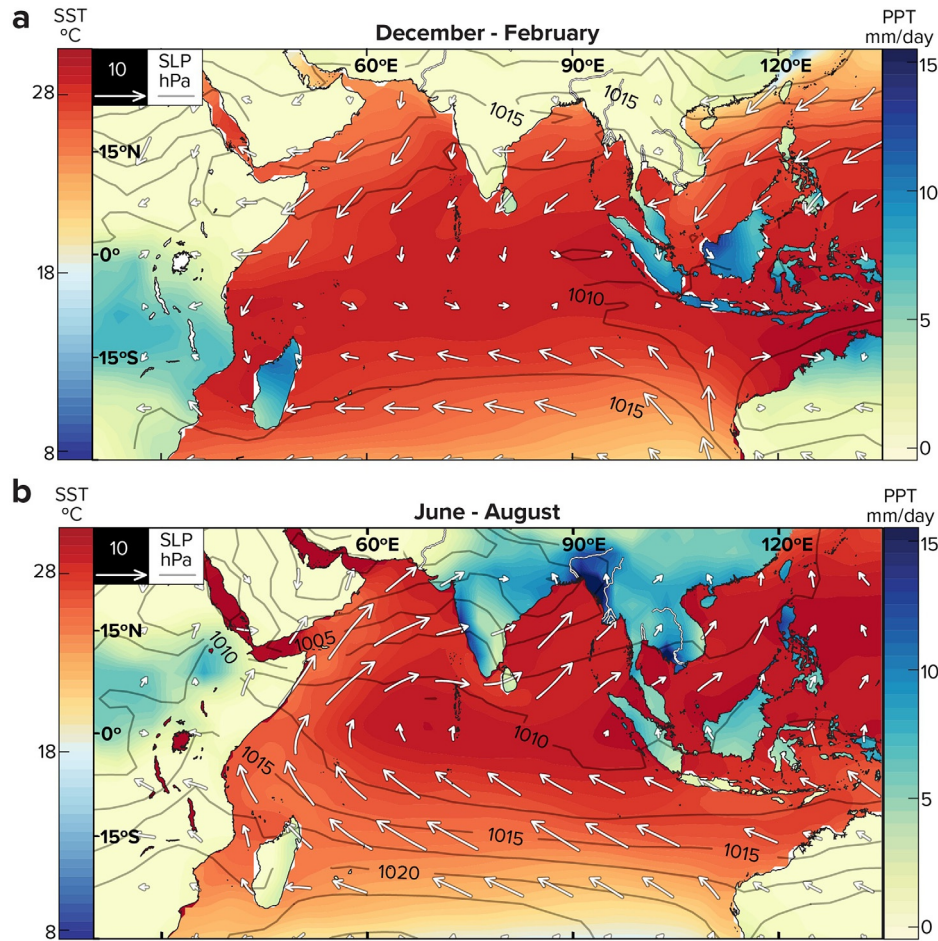


FIG. 1 Average sea surface temperature (SST in °C) with surface wind vectors (m/s), sea level pressure (SLP contours in hPa), and land precipitation (blue shading over land in mm/day) for (a) December through February and (b) June through August. Climatological conditions of SLP and winds based on NCEP/NCAR reanalysis (Kalnay, 1996), precipitation based on CMAP (Xie & Arkin, 1997), and SST on NOAA OISST version 2 (Reynolds et al., 2002) for the period 1982–2019.

current (Beal et al., 2011, 2015; Bryden et al., 2005; Hood et al., 2017, 2024a, 2024b; Sprintall et al., 2024). To the east, the Leeuwin Current is a small current that is unusual in being a southward-flowing eastern boundary current and in having the largest eddy kinetic energy among all midlatitude eastern boundary current systems (Feng et al., 2005, 2007, 2010; Hood et al., 2017, 2024a, 2024b; Phillips et al., 2021, 2024; Sprintall et al., 2024).

Finally, a significant fraction of the world's population lives in the coastal and interior regions of Indian Ocean rim countries, and they are directly impacted by the variability of the monsoons and associated rains. Moreover, many of these populations reside in low-lying coastal zones and island nations and are therefore threatened by tropical cyclones and sea level rise, both of which accelerate coastal erosion and the degradation of coastal ecosystems. Other Indian Ocean processes—such as seasonal variations in oceanic circulation and their associated biogeochemical and ecological responses—also directly and indirectly impact these populations through their influence on fisheries and food supplies (Feng et al., 2024; Hood et al., 2015; Marsac et al., 2024; Walker, 2024).

This chapter provides a brief overview of the history, geology, and physical and biogeochemical variability in the Indian Ocean. The reader is referred to the individual chapters in this book for greater detail.

2 Research history

The Indian Ocean was largely overlooked in the early days of oceanography, e.g., the Challenger expedition (1872–76) made only a single leg from Cape Town to Melbourne (see the review by Benson & Rehbock, 2002). The John Murray Expedition was the first major expedition to the Indian Ocean. This effort was focused on the Arabian Sea in 1933–34 (Sewell, 1934) and led to the discovery of the Arabian Sea Oxygen Minimum Zone (OMZ). During the International Geophysical Year (1957–58), oceanographic exploration of the southern Indian Ocean was carried out by Australian, French,

Japanese, New Zealand, and Soviet researchers. The International Indian Ocean Expedition (IIOE) in the early 1960s was the second major expedition to the Indian Ocean. The IIOE was a basin-wide interdisciplinary study that involved 46 research vessels from 14 different countries (Behrman, 1981; Hood et al., 2015). Among its many legacies, the IIOE led to the publication of the first oceanographic atlas of the basin (Wyrski et al., 1971) and a detailed map of the Indian Ocean bathymetry (Heezen & Tharp, 1966). The latter provided crucial information that contributed to the development of the theory of plate tectonics (Hood et al., 2015).

Subsequent measurements were made in the Indian Ocean under the Indian Ocean Experiment (INDEX 1979) and the Geochemical Ocean Section Study (Moore, 1984) in the 1970s, the Tropical Ocean Global Atmosphere (TOGA) program (1985–94; McPhaden et al., 1998; Webster et al., 1998), the Joint Global Ocean Flux Study (JGOFS; Fasham, 2003), and the World Ocean Circulation Experiment (WOCE) in the 1990s (Woods, 1985). The Geochemical Ocean Section Study was a global survey of the three-dimensional distributions of chemical, isotopic, and radiochemical tracers in the ocean. A key objective was to investigate the deep thermohaline circulation. The TOGA program did not do as much in the Indian Ocean as it did in the Pacific. Yet, it still led to some research progress related to physical variability, ocean-atmosphere interactions, and the influence of the El Niño-Southern Oscillation (ENSO) in the Indian Ocean. The JGOFS Arabian Sea Process Study was an interdisciplinary effort that contributed greatly to the current understanding of monsoon-forced biogeochemical variability in the Arabian Sea. WOCE provided the first systematic large-scale surface to bottom sampling of the Indian Ocean along multiple zonal and meridional transects.

The CLIVAR Repeat Hydrography project (Gould et al., 2013) in the first decade of the 21st century maintained selected transects from the WOCE sampling, and this global effort has continued through to the present under the ongoing GO-SHIP program (Talley et al., 2017). These large-scale survey programs have been consolidated with the implementation of the Indian Ocean Observing System (IndOOS) starting in 2006 (Beal et al., 2020; McPhaden et al., 2024). IndOOS is a multinational network of sustained oceanic measurements that is composed of five in situ observing networks: profiling floats (Argo), a moored tropical array (Research Moored Array for African-Asian-Australian Monsoon Analysis and Prediction; RAMA), repeat lines of temperature profiles (expendable bathythermograph, XBT, network), surface drifters, and tide gauges. The Second International Indian Ocean Expedition (IIOE-2; Hood et al., 2015) was launched in 2015. Like the IIOE, IIOE-2 is a basin-wide interdisciplinary study involving many countries. It is slated to continue through 2030. In addition to these major international efforts, there have been numerous national individual expeditions and programs that have contributed substantially to the inventory of measurements in the Indian Ocean.

Apart from the Geochemical Ocean Section Study, JGOFS, IIOE, and IIOE-2, most of the international programs have been primarily focused on atmospheric and physical oceanographic processes. As a result, understanding of the complex physical dynamics of the Indian Ocean has advanced much more rapidly compared to other disciplines. The WOCE, CLIVAR, and GO-SHIP transects have had an increasing emphasis on collecting biogeochemical measurements providing crucial information about carbon and nutrient concentrations and distributions in the Indian Ocean. Nonetheless, the complex physical dynamics that give rise to compound biogeochemical responses are not well sampled or understood in many subregions of the Indian Ocean. This is particularly true for the equatorial and Southern Hemisphere regions where much of the current understanding of biogeochemical variability is based on satellite remote sensing and numerical modeling (Hood et al., 2015, 2024a, 2024b).

3 Geology

The Indian Ocean was formed as a result of the breakup of the southern supercontinent Gondwana about 150 million years ago (Norton & Sclater, 1979). This breakup was due to the movement to the northeast of the Indian subcontinent about 125 million years ago. The Indian subcontinent collided with Eurasia about 50 million years ago, which coincided with the western movement of Africa and the separation of Australia from Antarctica. The approximate present configuration of the Indian Ocean was established by 36 million years ago. Most of the Indian Ocean basin is less than 80 million years old (Hood et al., 2015).

The oceanic ridges in the Indian Ocean are part of the worldwide oceanic ridge system where seafloor spreading occurs. The ridges form a remarkable triple junction in the shape of an inverted “Y” on the ocean floor of the Indian Ocean (Hood et al., 2015; Kennett, 1982; Talley et al., 2011; Turcotte & Schubert, 2002; Fig. 2). Starting in the upper northwest with the Carlsberg Ridge in the Arabian Sea, the ridge turns due south past the Chagos-Laccadive Plateau and becomes the Mid-Indian (or Central Indian) Ridge. Southeast of Madagascar, the ridge branches into the Southwest Indian Ridge, which continues to the southwest until it merges into the Atlantic-Indian Ridge south of Africa, and the Southeast Indian Ridge which extends to the east until it joins the Pacific-Antarctic Ridge south of Tasmania. These tectonically complex and active spreading centers have significant impacts on deep ocean chemical distributions (Hood et al., 2015; Nishioka et al., 2013; Vu & Sohrin, 2013), and they support diverse hydrothermal vent deep-sea communities (see <http://www.interridge.org/WG/VentEcology>).

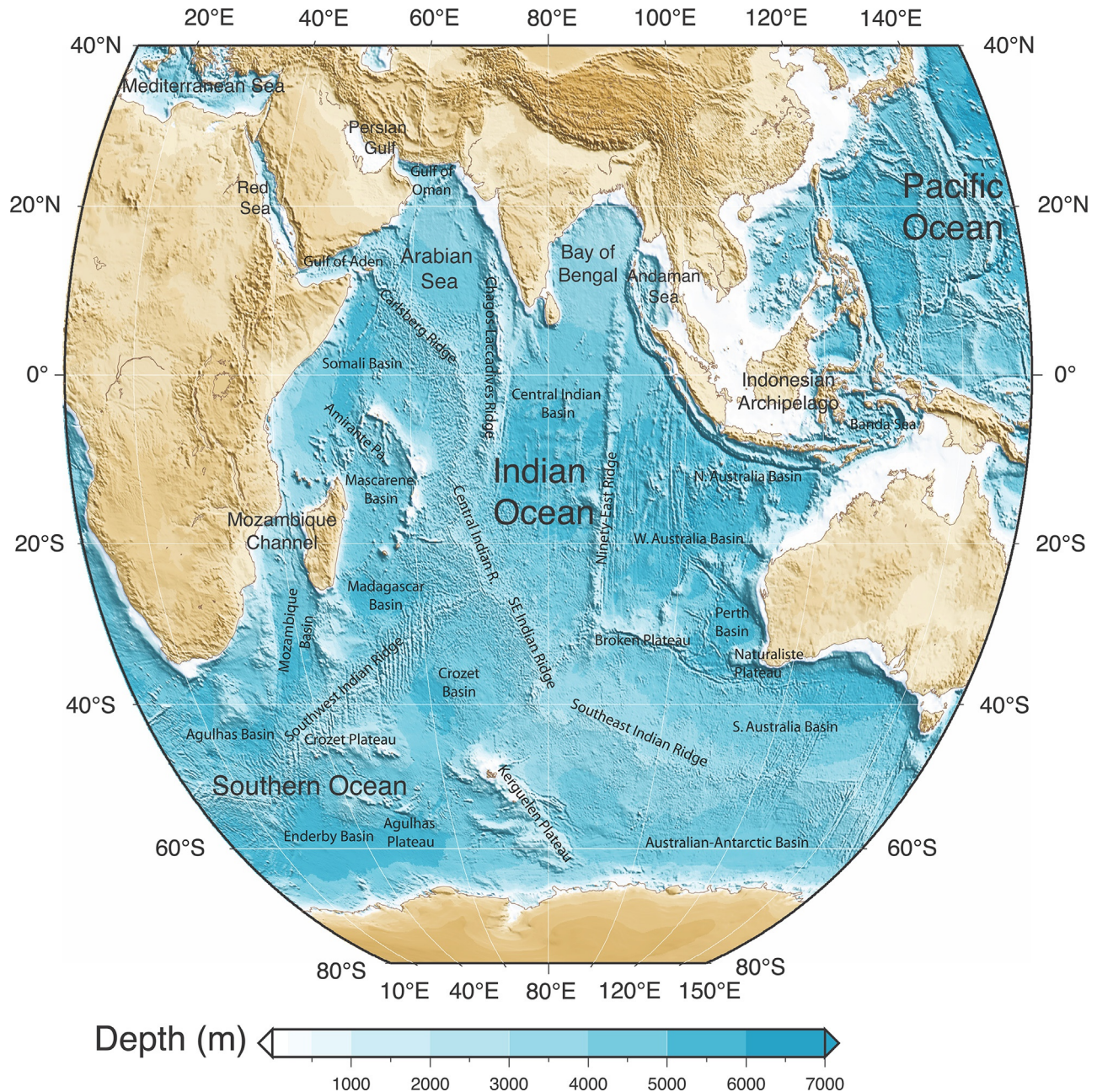


FIG. 2 Map of Indian Ocean topography, derived from ETOPO2 bathymetry data from NOAA NGDC. (From Talley *et al.* (2011).)

In addition, there are several prominent aseismic ridges in the Indian Ocean (Fig. 2). Perhaps the most striking is the Ninety East Ridge. It is the straightest and longest ridge in the world ocean (Hood *et al.*, 2015; Kennett, 1982). It runs northward along the 90° E meridian for 4500 km from the zonal Broken Ridge at latitudes 31°S to 9°N. Other important aseismic ridges in the Indian Ocean include the Madagascar, Chagos-Laccadive, and Mascarene plateaus (Fig. 2). These ridges, which extend to the surface to form island chains in many places, can have a profound impact on both surface and deep circulations in the Indian Ocean, and they can dramatically enhance biological productivity in the surface waters in regions where surface currents interact with topographic features and island chains, which can lead to nutrient and/or trace metal fertilization and pronounced island-wake effects (e.g., Strutton *et al.*, 2015).

The deep Indian Ocean basins are characterized by relatively flat plains of thick sediment that extend from the flanks of the oceanic ridges (Divins, 2003; Hood *et al.*, 2015). The Indian Ocean's ridge topography defines several separate basins

that range in width from 320 to 9000 km across. They include the Arabian, Somali, Mascarene, Madagascar, Mozambique, Agulhas, and Crozet basins in the west and the Central Indian Ocean, and the Wharton and South Australia basins in the east (Fig. 2).

The continental shelves of the Indian Ocean are, for the most part, relatively narrow, extending to an average width of only 120 km (Talley et al., 2011; Fig. 2). The widest shelves are found off India near Mumbai (Bombay) and off north-western Australia. The shelf break is typically found at a depth of about 150 m. The Ganges, Indus, and Zambezi rivers have carved particularly large canyons into the shelf breaks and slopes in the Bay of Bengal, Arabian Sea, and off Mozambique, respectively.

Finally, it should be noted that the tectonic activity associated with subduction zones of the Java Trench and the Sunda Arc trench system has generated numerous tsunamis and volcanic eruptions over geologic time, which have had widespread impacts in the Indian Ocean (Feng et al., 2024; Shen-Tu, 2016). Most recently, the Indian Ocean Tsunami of December 26, 2004 claimed more than 283,000 human lives in 14 countries, inundating coastal communities with waves up to 30 m high (Feng et al., 2024; Lay et al., 2005; Shen-Tu, 2016). This tsunami was one of the deadliest natural disasters in recorded history. Indonesia was the hardest-hit country, followed by Sri Lanka, India, and Thailand. Moreover, the geography of Indonesia is dominated by volcanoes that are generated by these subduction zones. As of 2012, Indonesia has 127 active volcanoes, and about 5 million people live and work within the volcanic danger zones (Hood et al., 2015).

4 Oceanography

4.1 Ocean circulation

4.1.1 Near-surface currents

The atmospheric and oceanic circulation of the Indian Ocean are different from those in the Pacific and Atlantic, largely due to geography. The Asian landmass limits the northern extent of the Indian Ocean to only $\sim 25^\circ\text{N}$, so there is no dense water formation at high northern latitudes as observed in the Atlantic and Pacific. As discussed earlier, the intense seasonal heating and cooling of air masses over Asia drives the seasonal monsoons (Fig. 1). Nonetheless, the timing of the onset and relaxation of the monsoons, their strength, and the associated wet and dry periods in the northern Indian Ocean rim countries are somewhat variable due to the influence of multiple larger-scale climate modes and smaller-scale ocean-atmosphere interactions (Phillips et al., 2024). The seasonally reversing winds drive corresponding surface coastal ocean currents in the northern Indian Ocean (Fig. 3), i.e., anticyclonic upwelling circulations during the Southwest Monsoon and cyclonic downwelling circulations during the Northeast Monsoon (Hood et al., 2017; Phillips et al., 2024). During the spring and fall inter-monsoon transitions, the winds relax and drive the equatorial Wyrki Jets that flow rapidly from the west to the east along the equator (McPhaden et al., 2015; Strutton et al., 2015; Wyrki, 1973).

The connection with the Pacific Ocean through the ITF in the southeastern tropical Indian Ocean also strongly influences Indian Ocean circulation patterns. The warm and fresh ITF into the Indian Ocean sets up a north-south pressure gradient that drives westward transport in the South Equatorial Current (Phillips et al., 2024) and southward transport in the Leeuwin Current (Hood et al., 2017; Fig. 3). The Leeuwin Current is a relatively small ($<5\text{ Sv}$) eastern boundary current (Godfrey & Ridgway, 1985; Feng et al., 2003; Hood et al., 2017; Sprintall et al., 2024) with elevated kinetic energy that sheds relatively high chlorophyll-*a* (Chl*a*), warm-core, downwelling eddies westward into open ocean waters (Feng et al., 2005, 2007, 2010; Hood et al., 2017; Waite et al., 2007). The South Equatorial Current crosses the southern tropical Indian Ocean between $\sim 10^\circ\text{S}$ and 20°S and feeds into the northward-flowing East African Coastal Current and the southward-flowing Agulhas Current (Fig. 3). The Agulhas is the largest Southern Hemisphere western boundary current (60–85 Sv) with sources derived from the Mozambique Channel, the Southeast Madagascar Current, and the southwest Indian Ocean subgyre (Hood et al., 2017; Phillips et al., 2024; Stramma & Lutjeharms, 1997). It flows southwestward along the coast of southern Africa from $\sim 25^\circ\text{S}$ to 40°S where most of it retroflects sharply eastward back into the Indian Ocean (becoming the eastward-flowing Agulhas Return Current), while shedding eddies (“Agulhas Rings”) that propagate into the Atlantic (Hood et al., 2017; Lutjeharms, 2006; Sprintall et al., 2024; Fig. 2). Meanders and eddies in the Agulhas Current propagate alongshore and interact with winds and topographic features which gives rise to seasonally variable localized upwelling and downwelling (Hood et al., 2017; Lutjeharms, 2006; Phillips et al., 2024).

In the central-eastern South Indian Ocean, the surface flow is generally eastward between 20°S and 30°S (Fig. 3; Godfrey & Ridgway, 1985; Phillips et al., 2024; Schott et al., 2009; Sharma, 1976; Sharma et al., 1978). This flow is driven by the large-scale, poleward drop in sea surface height (Godfrey & Ridgway, 1985; Phillips et al., 2024; Schott et al., 2009) that is associated with the transition from the warm and fresh South Equatorial Current waters to the cooler, saltier, and

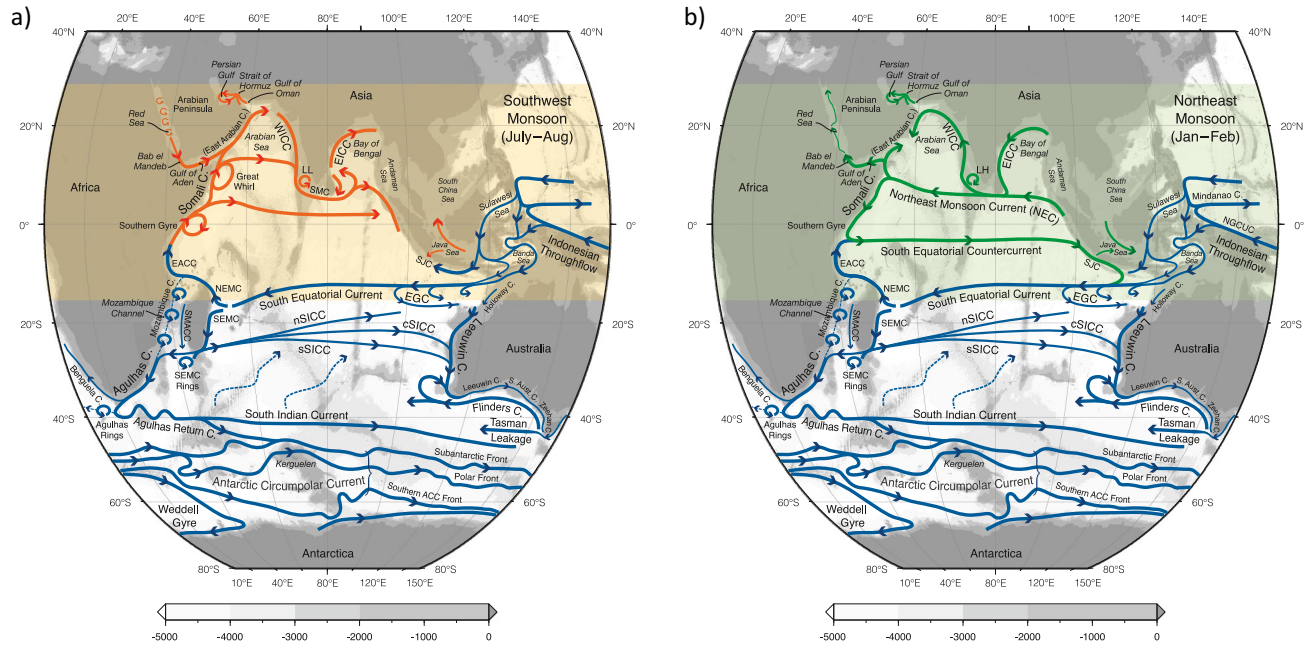


FIG. 3 Schematic near-surface circulation during the Southwest Monsoon (July–August, left panel). Schematic near-surface circulation during the Northeast Monsoon (January–February, right panel). *Blue*: year-round mean flows with no seasonal reversals. *Orange*: monsoonally reversing circulation (after [Schott & McCreary, 2001](#)). The ACC fronts are taken directly from [Orsi et al. \(1995\)](#). Acronyms: *EACC*, East African Coastal Current; *NEMC*, Northeast Madagascar Current; *SEMC*, Southeast Madagascar Current; *SMACC*, Southwest Madagascar Current; *WICC*, West Indian Coastal Current; *EICC*, East Indian Coastal Current; *LH* and *LL*, Lakshadweep high and low; *SJC*, South Java Current; *EGC*, Eastern Gyral Current; *SICC*, South Indian Countercurrent (south, central, and southern branches); *NEC*, Northeast Monsoon Current. Updated from [Talley et al. \(2011\)](#), originally based on [Schott and McCreary \(2001\)](#). The *light gray* shading shows seafloor bathymetry. (Modified from [Phillips et al. \(2021\)](#).)

denser waters to the south. Narrower eastward jets are embedded in this general eastward flow ([Fig. 3](#); [Divakaran & Brassington, 2011](#); [Maximenko et al., 2009](#); [Menezes et al., 2014](#); [Phillips et al., 2024](#)) that starts as a single flow near the southern tip of Madagascar around 25°S and then divide into separate jets at the Central Indian Ridge (65°E–68°E) ([Fig. 3](#); [Menezes et al., 2014](#); [Phillips et al., 2024](#)).

Variability in the atmospheric and oceanic circulation of the Indian Ocean is the result of complex interactions that are both internal and external to the Indian Ocean. The major drivers of this variability are still not fully understood ([Beal et al., 2019, 2020](#); [Phillips et al., 2024](#)).

4.1.2 The overturning circulation

The Indian Ocean overturning cells are important for the redistribution and transport of heat and other properties southward from the tropical Indian Ocean ([Figs. 4 and 5](#); [Beal et al., 2020](#); [Phillips et al., 2024](#); [Schott et al., 2002](#)). The shallow overturning (<500m) in the Indian Ocean ([Figs. 4 and 5](#)) consists of the cross-equatorial cell and the southern cell ([Beal et al., 2020](#); [Miyama et al., 2003](#); [Phillips et al., 2024](#); [Schott et al., 2004](#)). The ascending branches of these cells, which connect to different upwelling zones in the southern and northern Indian Ocean ([Figs. 4 and 5](#)), play an important role in regulating the heat balance in the tropical Indian Ocean ([Beal et al., 2020](#); [Lee, 2004](#); [Lee & McPhaden, 2008](#); [Sprintall et al., 2024](#)). In a deeper cell (>500m), highly oxygenated mode and intermediate waters of various density classes enter the Indian Ocean from the Southern Ocean at lower thermocline to intermediate depths and move northward. The upper part of this cell mixes with lighter water above and upwells to the sea surface, and then returns southward via wind-driven Ekman transport of near-surface waters ([Fig. 4](#); [Beal et al., 2020](#); [Schott et al., 2009](#); [Sprintall et al., 2024](#)). Deep and abyssal water masses flow equatorward in deep western boundary currents along the Madagascar and African coasts in the western Indian Ocean, and along the Southeast Indian and Ninety East Ridges in the eastern Indian Ocean. These deep waters are upwelled within the Indian Ocean and then mix together with the lower part of the mode and intermediate waters and return south to exit the basin at shallower depths via the Agulhas Current as part of the return flow within the meridional overturning circulation ([Beal et al., 2020](#); [Sprintall et al., 2024](#)).

Indian Ocean Features

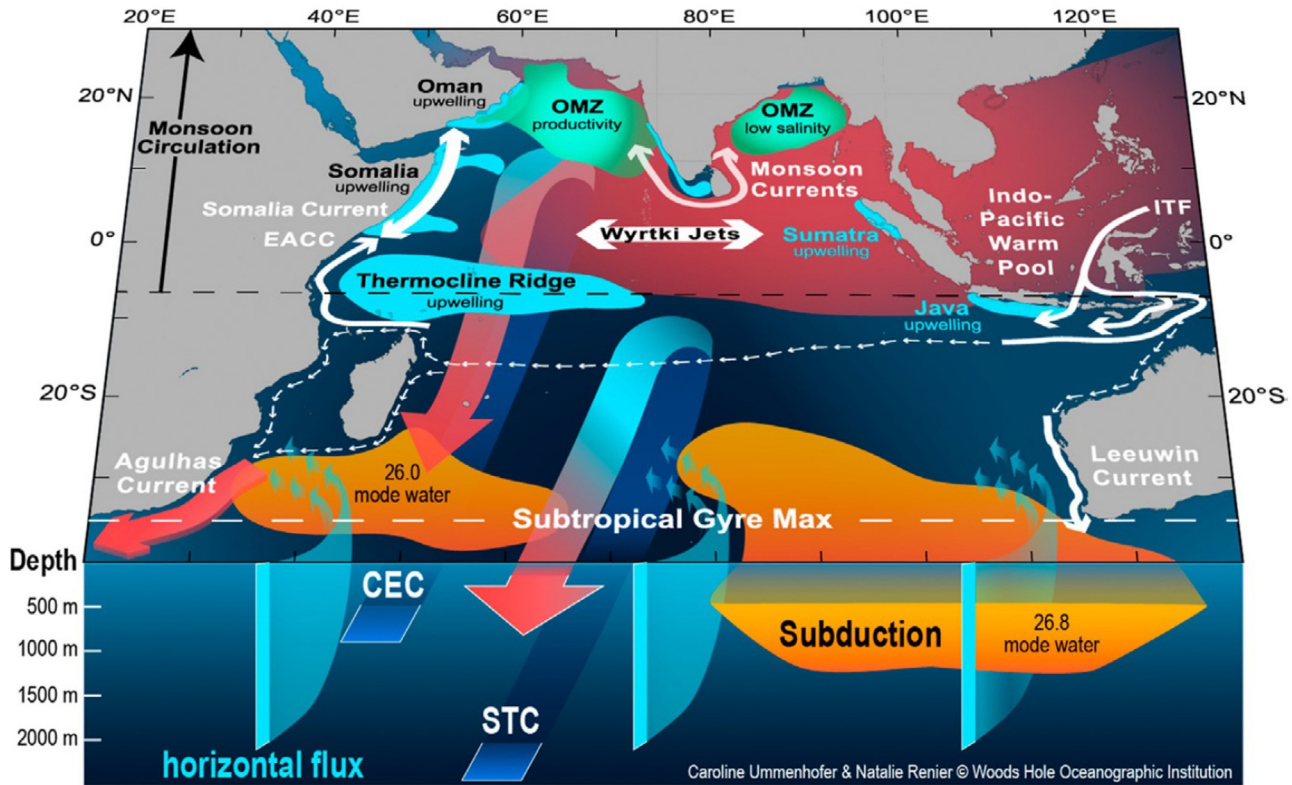


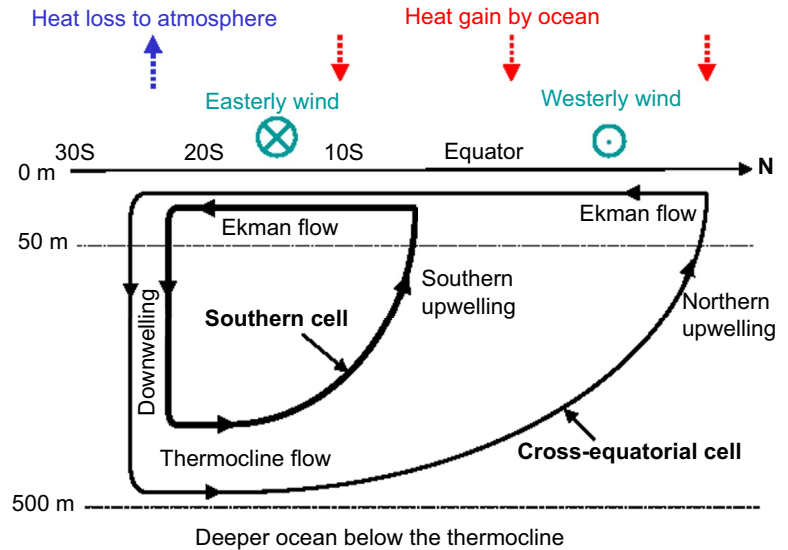
FIG. 4 Indian Ocean main oceanographic features and phenomena. The surface circulation seasonally reverses north of 10°S under the influence of monsoons. The summer monsoon also promotes the intense Somali current as well as upwelling and high productivity in the western Arabian Sea. High surface layer productivity, sinking of biomass, and its remineralization at depth also lead to the formation of subsurface oxygen minimum zones (OMZs) in the Arabian Sea and Bay of Bengal. The Indo-Pacific warm pool is a region of intense air-sea interactions, where the Madden-Julian oscillation, monsoon intraseasonal oscillation, and Indian Ocean dipole develop. The Indian Ocean is a gateway of the global oceanic circulation, with inputs of heat and freshwater through the Indonesian Throughflow, which exit the basin through boundary currents, mainly the Agulhas Current along Africa, but also the Leeuwin Current along Australia. There are two vertical overturning cells connecting subducted waters south of 30°S to the tropical Indian Ocean: the shallow subtropical overturning cell where water upwells in the “thermocline ridge” open-ocean upwelling region, and the cross-equatorial cell where water upwells farther north in the Arabian Sea off the coast of Somalia and Oman. These cells are the main source of subsurface ventilation due to the presence of continents to the north. (From Beal et al. (2020).)

4.2 Upper-ocean structure

4.2.1 Sea surface temperature (SST)

The tropical Indian Ocean includes the largest fraction of SST warmer than 28°C (relative to the size of the basin) in any ocean basin and has warmed faster than either the tropical Pacific or Atlantic (Fox-Kemper et al., 2021; Han et al., 2014; Phillips et al., 2024), with implications for primary productivity (Roxy et al., 2014, 2016, 2024). Over the past decade, the Indian Ocean has accounted for 50%–70% of the total global upper-ocean (700 m) heat uptake, due to global warming (Beal et al., 2020; Lee et al., 2015; Phillips et al., 2024; Ummenhofer et al., 2021). The patterns of SST variability in the Indian Ocean are largely driven by the spatial and seasonal variability in net heat fluxes at the sea surface (Beal et al., 2020; Phillips et al., 2024; Yu et al., 2007). In the southern Indian Ocean, this results in a seasonally driven SST pattern with the lowest SST during the austral winter and the highest SST during the austral summer (Fig. 1). In contrast, in the northern Indian Ocean, there is cooling and lower SST during the Northeast Monsoon due to the influence of the monsoon winds. Warming and higher SST occur there during the rest of the year and especially during spring and fall inter-monsoon periods when the winds are weak (Beal et al., 2020; Phillips et al., 2024; Yu et al., 2007). One important exception to this general pattern is the SCTR, where the seasonal cycle of SST is also strongly influenced by substantial intraseasonal variations in upwelling

FIG. 5 Conceptual illustration of the time-mean meridional overturning circulation of the upper Indian Ocean that consists of a southern and a cross-equatorial cell. The time-mean zonal wind and surface heat flux are also shown schematically. This flow is believed to partially supply the cross-equatorial thermocline flow. (From Lee (2004).)



(Beal et al., 2020; Foltz et al., 2010; Hermes & Reason, 2008; Phillips et al., 2024; Ummenhofer et al., 2021; Vialard et al., 2008). The seasonally reversing monsoon winds also drive upwelling and downwelling patterns that lead to more complex SST variability in the western Arabian Sea and in the coastal zones of Java and Bali (Beal et al., 2020; Chowdary et al., 2015; Hood et al., 2017; Phillips et al., 2024; Sprintall et al., 2024; Yu et al., 2007).

4.2.2 Upper-ocean stratification

Mixed layer depth in the northern Indian Ocean exhibits a strong seasonal cycle that is associated with the seasonally reversing monsoon winds (Beal et al., 2020; de Boyer Montegut et al., 2007; McCreary & Kundu, 1989; Phillips et al., 2024; Prasad, 2004; Rao et al., 1989; Rao & Sivakumar, 2003; Schott et al., 2002; Sreenivas et al., 2008). In the Arabian Sea, the winds induce Ekman transport that generates convergences and divergences that vary with both the monsoon phase and from the coastal to offshore regions. The Bay of Bengal exhibits weaker seasonal mixed layer depth variations due to the strong salinity stratification induced by river runoff that prevents entrainment mixing, particularly in the northern part of the Bay (Babu et al., 2004; Beal et al., 2020; Gopalakrishna et al., 2002; Narvekar & Kumar, 2006; Phillips et al., 2024; Prasad, 2004; Rao et al., 1989; Shenoi et al., 2002). In addition, temperature inversions are generated by river runoff and wintertime surface cooling near the surface in the northern Bay of Bengal. In this region, the surface water becomes cooler than the subsurface in winter owing to surface cooling, but still, the stratification is stable, due to freshwater supply by the rivers (Girishkumar et al., 2013; Nagura et al., 2015; Shetye et al., 1996; Thadathil et al., 2002). At the equator, the predominantly eastward flow causes a deepening of mixed layer depth from the west to the east (Ali & Sharma, 1994; Beal et al., 2020; O'Brien & Hurlburt, 1974; Phillips et al., 2024), although heavy precipitation results in shoaling of the mixed layer depth west of Sumatra during the Northwest Monsoon (Beal et al., 2020; Du et al., 2005; Masson et al., 2002; Phillips et al., 2024; Qu & Meyers, 2005). Thin mixed layers are associated with the SCTR due to upwelling (Beal et al., 2020; McCreary et al., 1993; Phillips et al., 2024; Resplandy et al., 2009; Vialard et al., 2009). In general, seasonal mixed layer depth shoaling and deepening in the southern Indian Ocean is related to the annual cycle of net surface heat flux and wind, with mixed layers deepening during the austral winter due to increased surface heat loss and increased winds and vice versa during the austral summer (Beal et al., 2020; Foltz et al., 2010; McCreary et al., 1993; Phillips et al., 2024).

Fresher surface layers can result in shallow mixed layers overlying a strongly stable salt-stratified pycnocline within a deeper isothermal layer: the region between the mixed layer depth and the isothermal layer is referred to as the barrier layer (Lukas & Lindstrom, 1991; Sprintall & Tomczak, 1992). The thickest barrier layers in the Indian Ocean are in the northern Bay of Bengal as a result of river runoff and in the equatorial waters off Sumatra as a result of precipitation and perhaps advection of fresh water from the Bay of Bengal (Fig. 6; Felton et al., 2014; Qu & Meyers, 2005; Sprintall & Tomczak, 1992). Significant seasonal variability of the barrier layer thickness in the Indian Ocean arises due to the interplay of surface water convergence, wave propagation, and precipitation associated with monsoon forcing (Felton et al., 2014; Qu & Meyers, 2005). Barrier layer thickness can impact the summer monsoon, development of the Indian Ocean Dipole (IOD), and tropical cyclone-induced upwelling in the Indian Ocean (Balaguru et al., 2012; DeMott et al., 2024; Feng

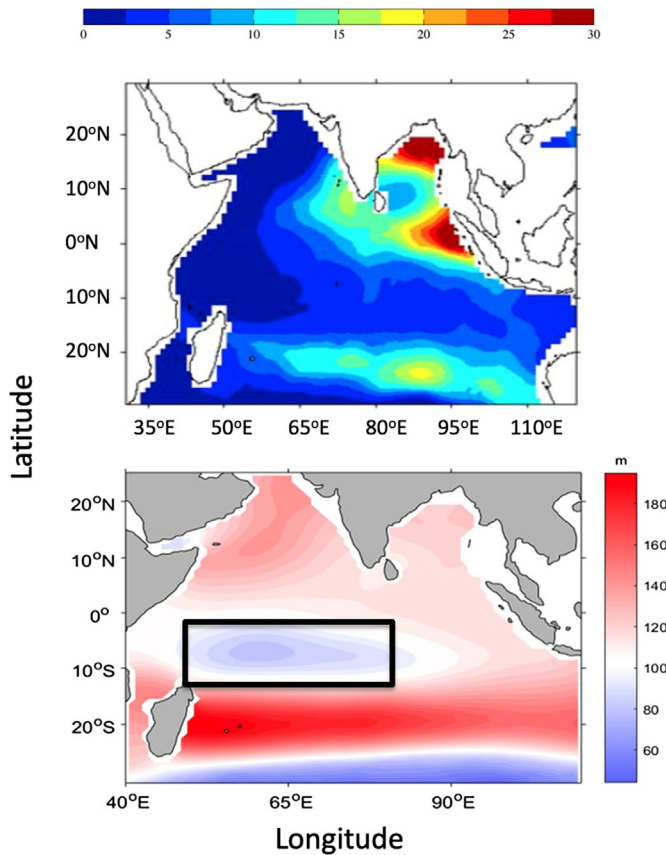


FIG. 6 Top panel: Annual mean barrier layer thickness in the Indian Ocean estimated from Argo data. Bottom panel: Climatological distribution of the thermocline depth in the Indian Ocean. The box denotes the approximate location of the SCTR. (*Top panel: Modified from Felton et al. (2014). Bottom panel: Modified from Yuan (2020).*)

et al., 2024; Guo et al., 2013; Kumari et al., 2018; Masson et al., 2005; Qiu et al., 2012; Sengupta et al., 2008; Tozuka et al., 2024; Yamagata et al., 2024).

As with mixed layer depth, the thermocline depth is strongly influenced by the monsoon winds and Ekman pumping in the northern Indian Ocean, with coastal upwelling during the Southwest Monsoon causing shoaling of the thermocline in coastal waters and deepening of the thermocline due to convergence and downwelling further offshore (Beal et al., 2020; Phillips et al., 2024; Fig. 6). However, remotely wind-forced planetary waves also impact thermocline depth in the northern Indian Ocean. For example, coastal Kelvin waves that originate in the eastern equatorial Indian Ocean can alter coastal currents, upwelling and downwelling signatures, and thermocline depth in both the Bay of Bengal and Arabian Sea (Beal et al., 2020; Hood et al., 2017; Phillips et al., 2024). The shallowest thermocline in the tropical Indian Ocean is associated with upwelling in the SCTR (Beal et al., 2020; McCreary et al., 1993; Phillips et al., 2024; Resplandy et al., 2009; Vialard et al., 2009; Fig. 6). The thermocline depth in the SCTR varies seasonally, shoaling in May and December and deepening in February and September, in response to the interaction between local wind-driven Ekman pumping and remotely forced downwelling Rossby waves that also influence interannual thermocline variability (Beal et al., 2020; Nyadjro et al., 2017; Phillips et al., 2024; Tozuka et al., 2010; Trenary & Han, 2012; Yu et al., 2005). At higher latitudes, the wind-induced Ekman pumping drives convergence in the subtropics deepening the thermocline and divergence further south leads to a shoaling thermocline (Fig. 6; Phillips et al., 2024; Sprintall & Tomczak, 1992).

4.3 Interocean basin connections and heat transport

The signature of the ITF water masses is readily identifiable as a low-salinity surface layer (Gordon et al., 1997; Hood et al., 2024a; Sprintall et al., 2024) separated from a low-salinity, high-silicate intermediate depth core that stretches across nearly the entire Indian Ocean within the South Equatorial Current (Sprintall et al., 2024; Talley & Sprintall, 2005). The ITF water masses then feed into the East African Coastal Current and the Agulhas Current, the latter via the Mozambique Channel and the Southeast Madagascar Current (Sprintall et al., 2024; Fig. 3). A smaller ITF contribution exports fresh water and heat into the poleward-flowing Leeuwin Current. The ITF nutrient inputs support a substantial amount of new production in the

Indian Ocean and significantly impact basin-wide biogeochemical cycles (Ayers et al., 2014; Hood et al., 2024a). The ITF itself is influenced by regional climate forcing from intraseasonal to interannual to decadal timescales. Yet, there remains a relatively poor understanding of the water mass mixing within the Indonesian seas and nutrient fluxes via the ITF due to the lack of in situ physical observations, and particularly the lack of biogeochemical measurements (Sprintall et al., 2024).

A portion of the western boundary East Australian Current Extension that is not reflected back into the Pacific Ocean turns south around Tasmania, Australia, and finds its way into the Indian Ocean as the “Tasman leakage” (Fig. 3; Ridgway & Dunn, 2007; Sprintall et al., 2024) and the Flinders Current that follows the southern Australian shelf-break (Fig. 3; Duran et al., 2020; Sprintall et al., 2024). Together with the ITF from the Pacific to the north of Australia, the Pacific water leakage south of Australia constitute important interocean contributions to the surface return flow of the global thermohaline circulation. The water masses derived from the ITF and Tasman leakage undergo significant changes along their pathway through the Indian Ocean, ultimately feeding into the Atlantic via the Agulhas Current (“Agulhas Rings”) (Fig. 3; Speich et al., 2002; Sprintall et al., 2024).

The Indian Ocean has the world’s largest southward meridional heat transport (Lumpkin & Speer, 2007), with an estimated 1.5 PW exiting the basin across 32°S. This export of heat balances the net heat gain from the atmosphere and the heat transport into the Indian Ocean from the Pacific via the ITF (Fig. 3) that is then advected out of the basin via horizontal currents and the aforementioned shallow overturning cells (Sprintall et al., 2024). The South Equatorial Current transports heat gained from the atmosphere and the ITF westward and the Agulhas Current then moves this heat poleward and into the Atlantic at surface and intermediate depths (Bryden & Beal, 2001; Sprintall et al., 2024). The Leeuwin Current makes a much smaller contribution to the poleward flow of heat from the ITF (Feng et al., 2003; Furue et al., 2017; Smith et al., 1991; Sprintall et al., 2024). However, the mesoscale eddies generated by the Leeuwin Current carry heat into the interior of the Indian Ocean, which ultimately contribute to heat export across 32°S (Dilmahamod et al., 2018; Domingues et al., 2006, Feng et al., 2007; Phillips et al., 2024; Sprintall et al., 2024).

5 Atmosphere

The South Asian monsoon represents the largest monsoon system on Earth. It is also a lifeline for nearly 2 billion people in the region, who receive 80% of their annual precipitation during the summer monsoon (Fig. 1), setting the rhythm of agricultural production (Singh, 2016; Ummenhofer et al., 2024a). Even small perturbations in the timing of onset, seasonal distribution, and monsoon intensity can have disproportionate effects on agriculture, most of which is rain-fed (Kumar et al., 2005; Parthasarathy et al., 1988). Historically, droughts have occasioned widespread famine and even today trigger large-scale impoverishment and rural-to-urban migration. For example, a 20% reduction in monsoon rainfall in 2002 resulted in billions of dollars in damage to the Indian economy (Gadgil et al., 2004); weakening of the monsoons and ensuing drought have had major societal effects and even upheavals throughout Asia (e.g., Cook et al., 2010; Mohtadi et al., 2024; Ummenhofer et al., 2024a).

Traditionally characterized to first order as continent-scale “sea breezes,” where in summer the land heats faster than the ocean, causing warm air to rise over the continent and moist air to be drawn in from the ocean, the South Asian monsoon exhibits seasonally reversing wind and precipitation patterns (Geen et al., 2020; Wang, 2006; Wang et al., 2005; Webster et al., 1998). Yet, more recently, theoretical advances have led to the emergence of the monsoon as a regional manifestation of the seasonal variation of the global tropical atmospheric overturning and migration of the associated convergence zones (Geen et al., 2020). Monsoon wind strength and precipitation vary on intraseasonal timescales, with so-called active and break spells, on interannual timescales in response to modes of climate variability, such as ENSO, as well as on multidecadal and longer timescales.

The Pacific Ocean’s influence on the Asian monsoon is well established, going back to the late 19th century, when a catastrophic drought and famine in India in 1877 coincided with unusually high pressure over Australia (Taschetto et al., 2020). The atmospheric pressure seesaw between the Pacific and Indian Oceans, first reported by Hildebrandsson (1897) and coined the Southern Oscillation by Walker (1925), became one of the key predictors of the strength of the South Asian monsoon. The pressure changes are indicative of zonal displacements of the ascending and descending branches of the tropical Walker Circulation; El Niño events are associated with anomalous descent of dry air over South Asia that tends to reduce monsoonal precipitation (e.g., Kumar et al., 1999), while an enhanced South Asian summer monsoon is typically observed during La Niña events when an intensified ascending branch is located over Indonesia (Ummenhofer et al., 2024a).

This traditional model of ENSO-South Asian monsoon interaction has worked well in the past, yet forecasting centers failed to predict the wet monsoon season in 1997 during one of the strongest El Niño events on record, and the failure of the monsoon in 2002 during a weak El Niño (Taschetto et al., 2020). Recent studies suggest that the well-established

relationships between ENSO and the Asian monsoon seem to be changing (e.g., [Kumar et al., 1999](#)), or that ENSO diversity plays a significant role. Regarding the latter, the location of maximum warming in the Pacific during an El Niño event, and the associated shift of the Walker circulation, appear to be important factors in determining whether or not an Indian summer monsoon failure materializes. Additionally, the role of other climate modes, such as the IOD and Indian Ocean Basin Mode, also modulate Indian summer monsoon rainfall ([Anil et al., 2016](#); [Ashok et al., 2001](#); [Behera et al., 1999](#); [Chang & Li, 2000](#); [Li et al., 2001](#); [Li & Zhang, 2002](#); [Yamagata et al., 2024](#)) and can modulate ENSO's influence on multi-decadal timescales ([Ummenhofer et al., 2011, 2024a](#)). Recent advances in regional modeling not only produce improved skill in simulating Indian summer monsoon rainfall but also better capture the transition of the active/break cycles and remote impacts by ENSO ([Shinoda et al., 2024](#)).

The Indian Ocean is characterized by very active intraseasonal variability (ISV, see [DeMott et al., 2024](#)), with far-reaching implications for regional climate. ISV of the atmosphere and ocean refers to disturbances with characteristic time-scales spanning approximately 30–90 days. In the tropics, atmospheric ISV is chiefly driven by the intraseasonal oscillation (ISO)—a large-scale atmospheric disturbance often featuring organized cloud clusters so large that they span nearly the entire Indian Ocean basin. The ISO remotely affects global weather through numerous teleconnections across timescales, from mesoscale to interannual ([Zhang, 2005, 2013](#)). Perturbations in the atmosphere introduced by ISV modify the ocean via atmosphere-ocean coupled feedbacks, and vice versa ([DeMott et al., 2015, 2024](#)). These feedbacks act across the ocean-atmosphere interface through the fluxes of heat, freshwater, and momentum, which are themselves regulated by processes that control the wind speed, temperature, and moisture content of the atmospheric boundary layer and the heat content of the upper-ocean mixed layer. Theories and global modeling studies have demonstrated that local air-sea coupling is critical for a realistic simulation of the northward propagating ISV ([DeMott et al., 2015, 2024](#); [Fu et al., 2007, 2008](#)): in the tropics, local air-sea coupling is effectively communicated to the deep troposphere via their impacts on convection, leading to well-defined regional air-sea coupled effects on monsoon precipitation and its ISV ([DeMott et al., 2015, 2024](#); [Shinoda et al., 2024](#)).

Changes in the broader Indian Ocean region as seen in recent decades ([Roxy et al., 2024](#)), including a warming Indian Ocean, contribute to increasing monsoon droughts and floods, and premonsoon heatwaves over South Asia ([Rohini et al., 2016](#); [Roxy et al., 2015, 2017, 2024](#); [Ummenhofer et al., 2024a](#); [Wang et al., 2021](#)). Yet, there is still considerable uncertainty about the future of the South Asian monsoon under anthropogenic forcing. Observations and climate models indicate a decreased Indian summer monsoon in the second half of the 20th century, primarily due to anthropogenic aerosol forcing, while in the long term, the South Asian monsoon is projected to increase over the 21st century and exhibit enhanced inter-annual variability ([Doblas-Reyes et al., 2021](#); [Douville et al., 2021](#); and references therein). The strong warming of the Indian Ocean in recent decades also has far-reaching implications for regional and global climate and weather patterns ([Ummenhofer et al., 2024a, 2024b](#)) and for Indian Ocean productivity ([Roxy, 2014](#); [Roxy et al., 2016, 2024](#)).

Paleoclimate proxies provide a long-term context for Asian monsoon variability ([Mohtadi et al., 2024](#); [Ummenhofer et al., 2024a](#)). A number of proxies have been used to reconstruct summer-monsoon wind strength in the Bay of Bengal and Arabian Sea where winds are 90% steady from the southwest at $\sim 15 \text{ ms}^{-1}$ during the summer-monsoon months. Although the close association between the onset of summer monsoon rains over India and the abrupt strengthening of the south-westerly Somali Jet over the Arabian Sea is well established in the modern climatology ([Boos & Emanuel, 2009](#)), none of these wind-related proxies directly record rainfall. More broadly, millennial-scale variability in temperature and productivity of the Indian Ocean and rainfall over the surrounding continents appear to be controlled by changes in the strength of the oceanic global thermohaline circulation. Cooling of the North Atlantic (e.g., during Heinrich Stadials and the Younger Dryas) provokes rapid changes in the atmospheric circulation by displacing the Intertropical Convergence Zone (ITCZ) to the warmer Southern Hemisphere, inducing a weaker Northern Hemisphere monsoon and a drier and warmer northern Indian Ocean. As such, periods of a less vigorous Atlantic Meridional Overturning Circulation (AMOC) during the last glacial period are generally associated with anomalously dry conditions for the South Asian monsoon ([Berkelhammer et al., 2012](#); [Dutt et al., 2015](#)). In contrast, periods of stronger AMOC and a relatively warmer North Atlantic result in generally wetter conditions at or north of the equator and drier conditions over the Southern Hemisphere monsoon regions ([Mohtadi et al., 2014, 2024](#); [Wurtzel et al., 2018](#)).

6 Hydrology and hydrography

The Indo-Pacific Warm Pool ([Fig. 7b](#)), defined by the 28°C isotherm and associated with the ascending branch of the Walker Circulation, represents a key regional and global feature ([Ummenhofer et al., 2024b](#)). The Indo-Pacific Warm Pool is characterized by high climatological mean rainfall rates, with the precipitation maximum tapering off toward the western part of the basin ([Fig. 7a](#)). Lower precipitation is found in the subtropics, with minima near $\sim 20^\circ\text{--}30^\circ\text{S}$ off the coast of

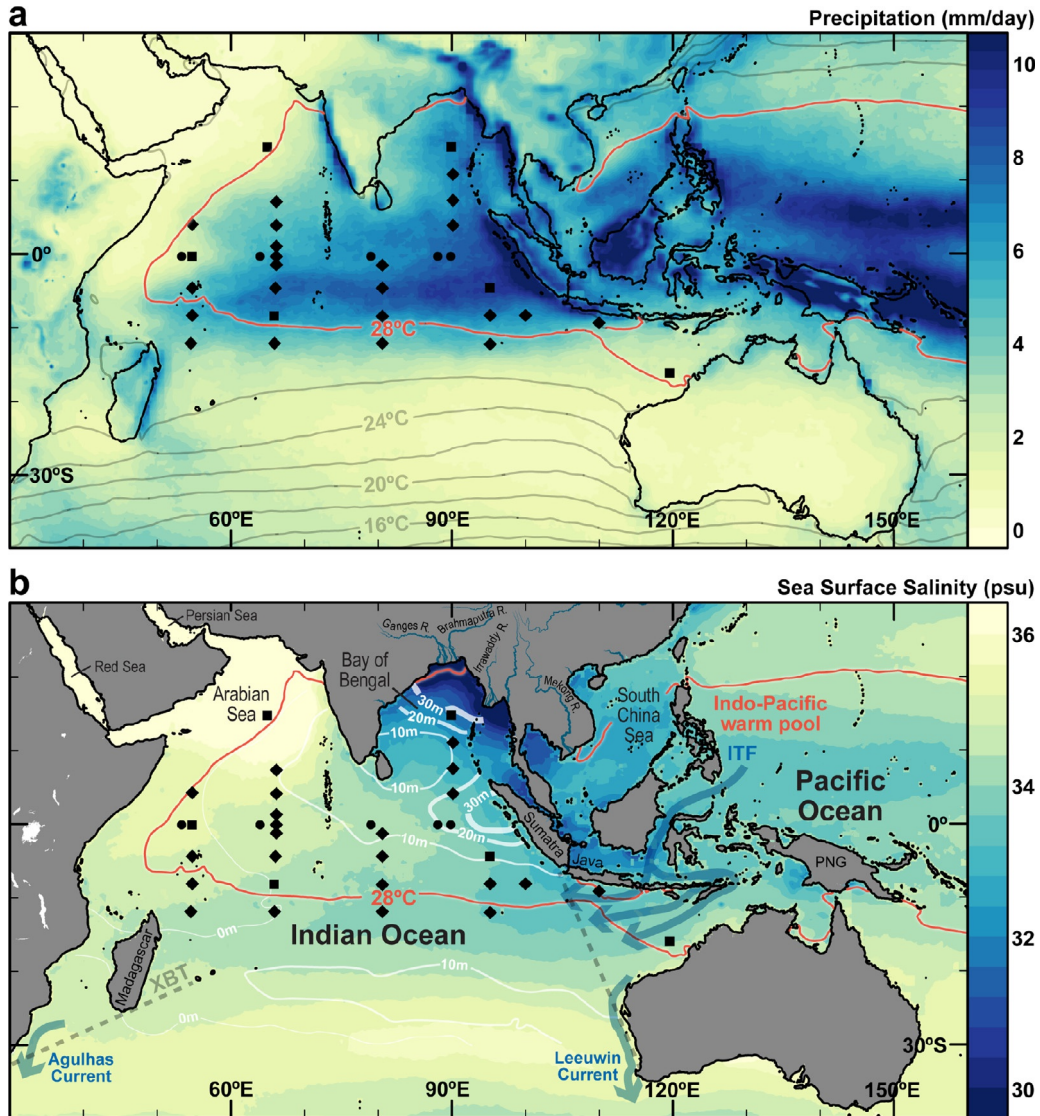


FIG. 7 (a) Average precipitation (shaded; in mm/day) and surface ocean temperature contours superimposed (in °C); (b) sea surface salinity (SSS, shaded; in psu) and barrier layer thickness (in m). The 28°C isotherm defines the Indo-Pacific Warm Pool. The black symbols denote positions of the RAMA array moorings in 2018, gray dashed lines key XBT lines; and major geographical features, rivers, and currents are indicated as well. (Adapted from Ummenhofer et al. (2021) and barrier layer thickness from Felton et al. (2014).)

Western Australia and in the northwestern Arabian Sea (Adler et al., 2017; Ummenhofer et al., 2024b). Both areas are also associated with high evaporation (not shown). There are strong spatial variations in precipitation, particularly in the tropics, near orographic features such as the west coasts of India, Myanmar, northern Thailand, Sumatra, and Borneo (Fig. 7a). Similarly, large variability in precipitation is caused by the seasonal march of the moisture transport of the monsoon system and the interannual movement of the Walker circulation in both zonal extent and magnitude associated with ENSO and IOD variability (Ummenhofer et al., 2024b; Yang et al., 2010).

To first order, the observed sea surface salinity (SSS; Fig. 7b) is expected to reflect the evaporation-minus-precipitation patterns (Ummenhofer et al., 2021). Using the same color scheme, Fig. 7 illustrates where the time-mean SSS pattern corresponds to the time-mean precipitation pattern; locations with a mismatch between the two fields are indicative of regions where effects from river discharge and ocean dynamics dominate. The maritime continent region is relatively fresh due to the direct input from extensive convection and precipitation as part of the rising branch of the Walker circulation, as well as Southeast Asian riverine input, for example, from the Mekong River system entering the South China Sea (Fig. 7b).

Freshwater from the Indonesian Seas lowers the salinity from the surface to ~ 600 m depth from where the ITF enters the southeast Indian Ocean at $\sim 10^{\circ}$ – 15° S all the way across the Indian Ocean to the coast of Madagascar (Fig. 7b; Gordon, 1986; Hu et al., 2019; Sprintall et al., 2024; Talley & Sprintall, 2005). To the south of the ITF influence, the low surface salinity along 15° S is located about 6° – 8° south of the rainfall maximum, and so is likely largely contributed through southward Ekman transport in response to the dominant easterly winds found south of 10° S in the Indian Ocean (Sengupta et al., 2006). On the other hand, the part of the fresh ITF that enters the poleward flowing Leeuwin Current acts to erode the effect of the local evaporation-precipitation maximum in this region such that relatively low SSS is found against the coast of Western Australia (Fig. 7b).

The northern Indian Ocean is characterized by stark salinity contrasts between the Arabian Sea and Bay of Bengal. River runoff from the Ganges, Brahmaputra, and Irrawaddy River systems (Fig. 7b) accounts for 60% of total riverine input north of 30° S (Sengupta et al., 2006). This produces the freshest surface waters of all the global tropical oceans (Fig. 7b) and is associated with strong salinity stratification in the upper 50–80 m in the Bay of Bengal (Mahadevan et al., 2016a, 2016b). However, the uncertainty in the freshwater distribution and mixing pathways for riverine input is high, and shallow, salinity-controlled mixed layers and the presence of barrier layers significantly affect the upper-ocean temperature (Sengupta et al., 2006; Thadathil et al., 2002; Wijesekera et al., 2016). In contrast, the saltiest waters of the Indian Ocean basin are found in the northern Arabian Sea influenced by the high-salinity inflow from the marginal Red and Persian Seas (Al-Yamani et al., 2024) that is largely evaporatively driven (Zhai et al., 2015). Subsurface salinity patterns in the upper 200 m within the Indian Ocean basin primarily reflect the SSS patterns (Hu et al., 2019).

7 Biogeochemistry, productivity, and fisheries

7.1 Nutrients, phytoplankton, and zooplankton

In the Arabian Sea, the strongest upwelling occurs during the Southwest Monsoon, driving nutrient enrichment and elevated phytoplankton productivity in coastal waters off Somalia, Yemen, and Oman and along western India (Fig. 8; see also Hood et al., 2017, 2024a, 2024b, and references cited therein). Multiple lines of evidence indicate that production in these upwelled waters is limited by silicate (Si) and iron (Fe) (Hood et al., 2024a; Moffett & Landry, 2020; Moffett et al., 2015; Naqvi et al., 2010; Wiggert et al., 2006; Fig. 9). In contrast, the Northeast Monsoon drives downwelling except in the northern central Arabian Sea where primary production increases due to wind-driven nutrient entrainment (Hood et al., 2017, 2024a; Wiggert et al., 2000, 2005). Yet, the seasonal and spatial variability in mesozooplankton biomass in the Arabian Sea is surprisingly weak (Baars, 1999; Hood et al., 2024a, 2024b; Madhupratap et al., 1992). The relative grazing contributions of meso- and microzooplankton vary spatially and seasonally in a way that is consistent with spatially separated co-regulation by grazing and iron limitation (Hood et al., 2024a; Moffett & Landry, 2020).

The biogeochemistry and productivity in the Bay of Bengal are influenced by the same factors that impact the Arabian Sea, but seasonal wind effects are less pronounced due to weaker winds and strong freshwater stratification (Gomes et al., 2000; Hood et al., 2017, 2024a, 2024b; Kumar et al., 2002, 2007; Thushara et al., 2019; Vinayachandran et al., 2002, 2005; Vinayachandran, 2009; Wijesekera et al., 2016; Fig. 8). Although Bay of Bengal surface waters are generally oligotrophic, there are regions of high Chl *a* and production that are associated with river plumes, wind-induced nutrient entrainment, and upwelling eddies (Hood et al., 2024a; Vinayachandran, 2009; Fig. 8). This variability in productivity drives significant fluctuations in mesozooplankton biomass (Fernandes & Ramaiah, 2019; Hood et al., 2024a; Muraleedharan et al., 2007; Ramaiah et al., 2010). As a result of the low oxygen concentrations in waters beneath the pycnocline, mesozooplankton in the Bay of Bengal are likely forced to inhabit a thin mixed layer where trophic interactions are more concentrated compared to other regions in the Indian Ocean (Hood et al., 2024a).

There are strong nutrient, Chl *a*, production and zooplankton responses to monsoon wind forcing in the SCTR. Vertical sections show that the nutricline and deep Chl *a* maximum shoal sharply in the SCTR region (Hood et al., 2024a). In the SCTR, the highest Chl *a* and primary production is observed in the austral winter (June–August) due to the strong Southeast Monsoon winds that increase wind stirring and induce upwelling (Dilmahamod, 2014; Hood et al., 2024a; Resplandy et al., 2009; Fig. 8c). Zooplankton biomass is relatively low for most of the year in the SCTR region with a pronounced peak during the Southeast Monsoon upwelling in August (austral winter) (Hood et al., 2024a). Southeast Monsoon winds along the southern coasts of the Indonesian island chains, combined with the upwelling-favorable South Java Current, drive nutrient inputs that give rise to substantial increases in Chl *a* concentration and primary production (Hood et al., 2017, 2024a; Fig. 8c). This elevated productivity during the Southeast Monsoon supports order of magnitude increases in zooplankton biomass compared to open-ocean waters further south (Hood et al., 2024a; Tranter & Kerr, 1969, 1977).

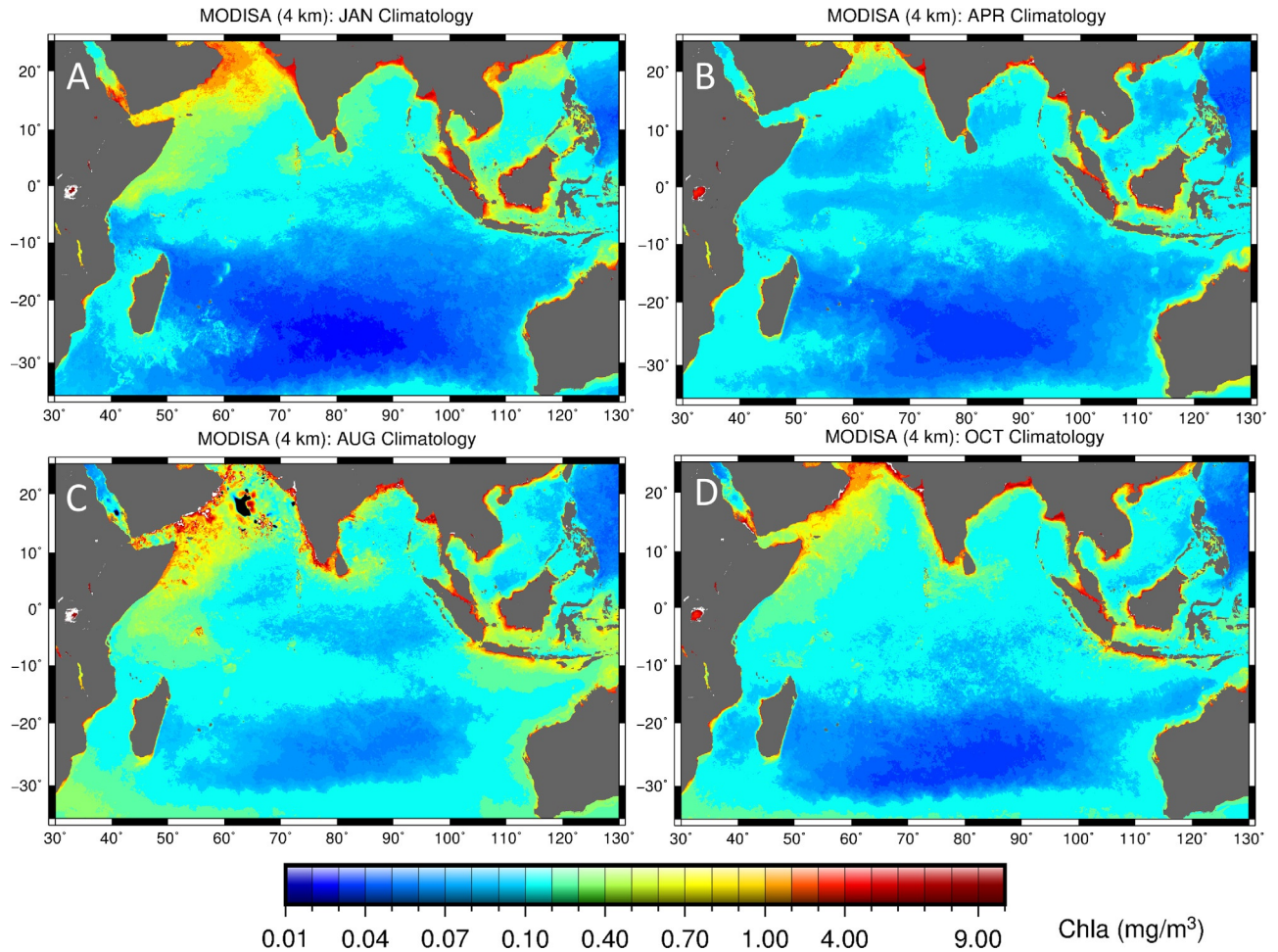


FIG. 8 Monthly climatology of MODIS-Aqua (4-km resolution) chlorophyll: (a) January, (b) April, (c) August, and (d) October. The climatology fields were obtained from the Goddard DAAC (<http://daac.gsfc.nasa.gov>). (From Hood et al. (2017).)

The ITF has a strong influence on biogeochemical properties and productivity in the eastern Indian Ocean and in the Leeuwin Current (Hood et al., 2017; Sprintall et al., 2024). The tropical origins of the Leeuwin Current combine with its southward-flowing downwelling favorable tendency to create a warm oligotrophic current (Hood et al., 2017, 2024a). However, local wind forcing during the austral summer can override this general tendency and drive localized upwelling (Gersbach et al., 1999; Hanson et al., 2005a, 2005b; Hood et al., 2017, 2024a; Pearce & Pattiaratchi, 1999; Rossi et al., 2013a, 2013b). Seasonal nutrient climatologies from the southwestern Australian shelf, the Leeuwin Current, and offshore suggest nitrogen (N) limitation of primary production (Lourey et al., 2006). Primary production is lowest during the austral summer (Fig. 8a) when the water column is most stratified (Hanson et al., 2007). However, primary production in nearshore upwelling regions can attain very high levels (Furnas, 2007; Hood et al., 2017). In the poleward-flowing Leeuwin Current, tropical zooplankton are transported southward (Buchanan & Beckley, 2016; Sutton & Beckley, 2016) and warm-core anti-cyclonic eddies carry moderately high Chla coastal water offshore (Hood et al., 2017, 2024a; Paterson et al., 2008; Waite et al., 2007, 2015). In open ocean waters of the southeastern Indian Ocean, the highest mesozooplankton biomass is observed in tropical waters south of Java and to the north of northwest Australia driven by upwelling during the Southeast Monsoon (Hood et al., 2024a; Tranter & Kerr, 1969, 1977; Fig. 8).

In the southwest Indian Ocean meanders and topographic interactions of alongshore-propagating eddies in the Mozambique Channel and Agulhas Current drive upwelling and downwelling and lateral transport which have a strong influence on nutrient concentrations and primary production (Hood et al., 2017, 2024a; Jose et al., 2014, 2016; Lamont et al., 2014; Lutjeharms, 2006; Meyer et al., 2002; Roberts et al., 2014). The Agulhas is an oligotrophic current that is derived from oligotrophic surface waters from the southwestern tropical Indian Ocean (Lutjeharms, 2006). Nutrient concentrations, Chla and production in surface waters of the Agulhas are particularly low during the austral summer and higher in the

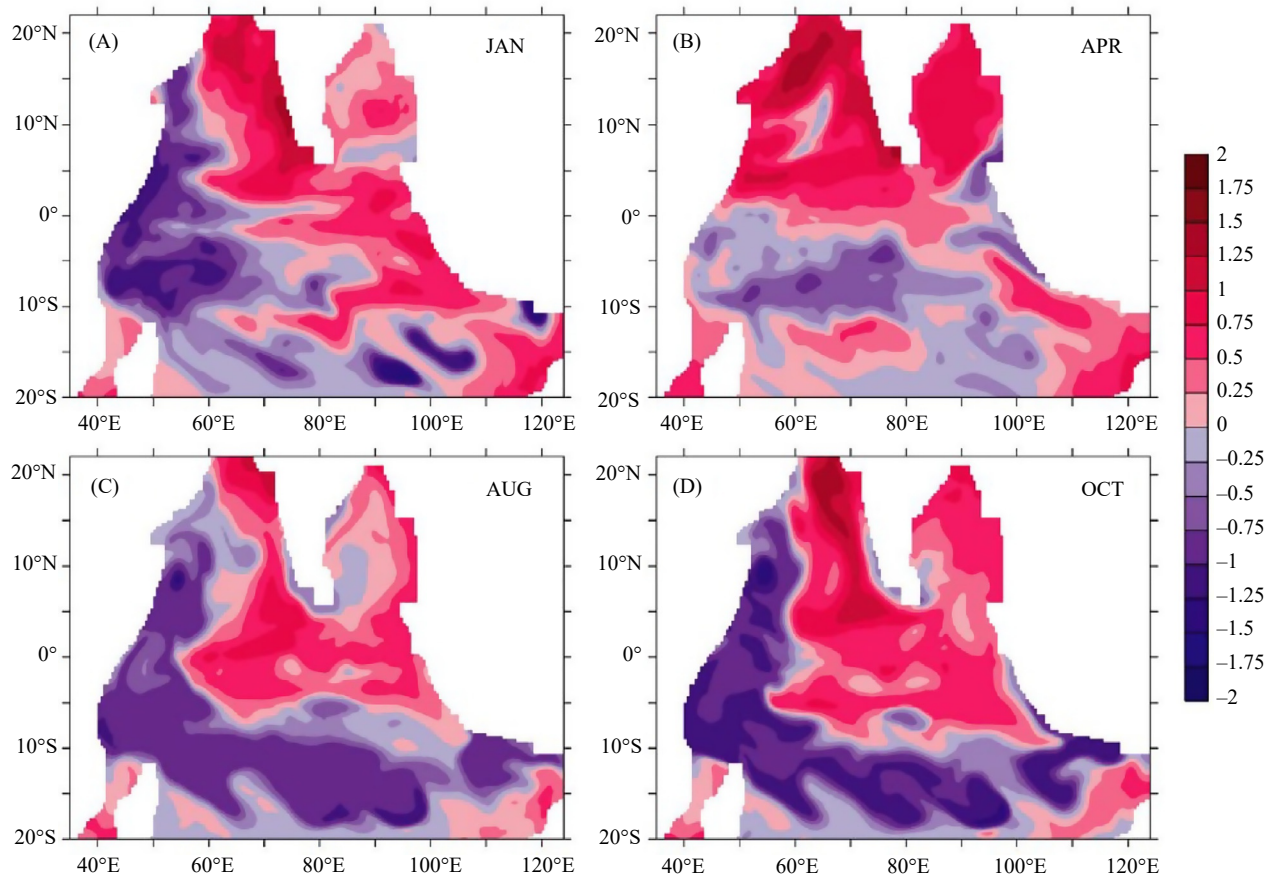


FIG. 9 Model-simulated seasonal evolution of most limiting surface nutrient for net plankton with blue (red) indicating Fe (N) limited growth (i.e., red is iron replete). The four seasons consist of (a) January (NEM); (b) April (Spring Intermonsoon); (c) August (SWM); and (d) October (Fall Intermonsoon). (Modified from Wiggert *et al.* (2006).)

austral winter (Hood *et al.*, 2017, 2024a; Machu & Garçon, 2001; Fig. 8a, c, respectively). However, the Agulhas Current can elevate primary production in the coastal zone through meandering and topographic interactions that drive upwelling and conversely can suppress primary production when the current impinges onto the shelf (Hood *et al.*, 2017, 2024a; Schumann *et al.*, 2005). In general, zooplankton biomass is also low in these waters, except in coastal/shelf regions (Huggett & Kyewalyanga, 2017). The Agulhas Current transports Indo-Pacific zooplankton southward (De Decker, 1973; Hood *et al.*, 2024a). Over the Agulhas Bank, primary production is somewhat elevated, and the zooplankton community is dominated by the large calanoid copepod *Calanus agulhensis* in the austral spring (De Decker & Marska, 1991; Hood *et al.*, 2024a; Huggett & Richardson, 2000).

In general, the southern subtropical gyre of the Indian Ocean is extremely oligotrophic (Harms *et al.*, 2019; Hood *et al.*, 2024a; Fig. 8). However, during the austral summer and fall, the South-East Madagascar bloom often occurs off the south-eastern tip of Madagascar extending eastward into the subtropical gyre (Longhurst, 2001; Fig. 8a). It has been suggested that the bloom is fueled by N_2 fixation, with eddies and/or the South Indian Counter Current advecting and dispersing it eastward (Dilmahamod *et al.*, 2019; Hood *et al.*, 2024a; Huhn *et al.*, 2012; Poulton *et al.*, 2009; Raj *et al.*, 2010; Srokosz & Quartly, 2013; Srokosz *et al.*, 2004, 2015; Uz, 2007). Phytoplankton production (Harms *et al.*, 2019) and mesozooplankton abundance (Rochford, 1969, 1977) are particularly low in the western subtropical gyre, and variability is associated with specific water masses that have different Chl *a* concentrations (Cedras *et al.*, 2020; Hood *et al.*, 2024a). Zooplankton biomass is highest in the eastern subtropical gyre, likely due to the influence of increased winds and eddy activity associated with Leeuwin Current and proximity to the Australian land mass (Hood *et al.*, 2024a). Grazing rates estimated from the southern subtropical gyre of the Indian Ocean indicate that the dominant pico-sized phytoplankton largely escape direct feeding by mesozooplankton, and microzooplankton are the major grazers as is typical for highly oligotrophic waters (Hood *et al.*, 2024a; Jaspers *et al.*, 2009).

7.2 Oxygen, carbon, and pH

The Arabian Sea has the thickest OMZ in the world where oxygen concentrations in intermediate water ($\sim 200\text{--}800\text{m}$) decline to nearly zero (e.g., Hood et al., 2024b, and references cited therein; Morrison et al., 1999; Naqvi et al., 2005). These low oxygen concentrations have significant impacts on nitrogen cycling, which include denitrification-driven depletion of subsurface nitrate (NO_3^-) concentrations, generation of nitrite (NO_2^-) maxima, production of greenhouse gases (nitrous oxide, N_2O), and globally significant fluxes of N_2 gas into the atmosphere from the ocean (Codispoti et al., 2001; Naqvi et al., 2005; Ramaswamy et al., 2001). In contrast, in the Bay of Bengal, these biogeochemical impacts are much less pronounced where dissolved oxygen concentrations in intermediate waters remain just above the threshold for denitrification (Hood et al., 2024b; Naqvi et al., 2005). In both the Arabian Sea and the Bay of Bengal, the low oxygen conditions in the water column are associated with high biological oxygen demand and slow ventilation. It appears that the more intense OMZ in the Arabian Sea is due to a weaker mineral (e.g., plankton shell) ballast effect and a higher biological production that support a higher biological oxygen demand compared to the Bay of Bengal, which, in turn, balances the higher physical oxygen supply (Hood et al., 2024b; Rixen et al., 2019). Elsewhere in the Indian Ocean, oxygen concentrations generally remain above the hypoxic threshold of 2 mg/L with concentrations increasing southward due to the influence of northward transport of well-oxygenated intermediate waters from the Southern Hemisphere. Highly oxygenated, intermediate water is formed along the northern edge of the Antarctic Circumpolar Current, and subsequently spreads throughout the Indian Ocean (McCreary et al., 2013).

It is anticipated that global warming will lead to the intensification of monsoon winds (Goes et al., 2005). Regional modeling studies have suggested that stronger Southwest Monsoon winds intensify the upwelling, increasing biological productivity and respiration in excess of the increase in wind-driven ventilation and are also responsible for deepening of the OMZ (Hood et al., 2024b; Lachkar et al., 2018). Increases in Arabian Sea denitrification are expected to cause increases in the N_2 and N_2O production. However, studies of long-term OMZ variability in the global ocean reveal only a small decline in oxygen concentrations in the Arabian Sea and the Bay of Bengal OMZs compared to in OMZs in the Pacific and Atlantic Oceans (Hood et al., 2024b; Ito et al., 2017; Naqvi, 2019; Schmidtke et al., 2017; Stramma et al., 2008). Moreover, outbreaks of hydrogen sulfide (H_2S) have so far not been reported in the northern Indian Ocean during the past 50 years (Rixen et al., 2020), other than in bottom waters on the Indian shelf (Naqvi et al., 2000). The absence of H_2S suggests that the balance of physical oxygen supply and biological oxygen demand has been fairly constant in the Arabian Sea and the Bay of Bengal OMZs and has maintained the current hypoxic/anoxic conditions (Hood et al., 2024b; Rixen et al., 2020). However, some of the most recent studies suggest that oxygen levels are now declining in the Arabian Sea and that these declines are giving rise to major biogeochemical and ecological impacts (Goes & Gomes, 2016; Goes et al., 2020; Gomes et al., 2014; Lachkar et al., 2019, 2020). Earth System Model-projected future changes in the northern Indian Ocean OMZs are highly uncertain (Bopp et al., 2013; Cabré et al., 2015; Kwiatkowski et al., 2020; Rixen et al., 2020). This uncertainty largely arises from differences in the magnitude and timing of the ventilation and biological oxygen demand in the models and how strongly they offset each other (Hood et al., 2024b; Resplandy, 2018; Rixen et al., 2020).

Inorganic and organic carbon pools and air-sea carbon fluxes in the Indian Ocean are sparsely sampled compared to other ocean regions (Hood et al., 2024b; Fig. 10). The Indian Ocean is responsible for $\sim 20\%$ of the oceanic uptake of atmospheric CO_2 (Takahashi et al., 2002). The Arabian Sea is a significant source of CO_2 to the atmosphere due to Southwest Monsoon-driven upwelling of water with elevated pCO_2 (Fig. 10; see also De Verneil et al., 2021; Takahashi et al., 2009, 2014). In contrast, whether the Bay of Bengal is a source or sink of CO_2 to the atmosphere is still uncertain due to the lack of measurements (Fig. 10; Bates et al., 2006; Hood et al., 2024b). The Indian Ocean appears to be a strong net sink of atmospheric CO_2 south of 14°S . This sink is due to the combined effects of both the biological pump (biologically mediated carbon export) and the solubility pump (CO_2 dissolution and physical transport and mixing) (Fig. 10; Hood et al., 2024b; Valsala et al., 2012). Some of the lowest pH values in the world have been measured in surface waters of the Indian Ocean (Feely et al., 2009). Further acidification is anticipated in the Indian Ocean over the coming decades given projected increases in oceanic CO_2 concentrations. This acidification will likely have severe negative impacts on coral reefs and other calcifying organisms (Bednarsek et al., 2012; de Moel et al., 2009; Hoegh-Guldberg et al., 2007; Hood et al., 2015, 2024b; IPCC, 2019).

As observed globally, dissolved organic carbon (DOC) concentrations in the Indian Ocean are high in stratified near-surface tropical and subtropical waters because these are regions where DOC is produced and accumulates. In contrast, the lowest DOC concentrations are observed in the deep ocean where heterotrophic consumption of DOC is much greater than autotrophic production (Hansell, 2009; Hansell et al., 2009). Furthermore, as observed globally, near-surface particulate organic carbon (POC) concentrations in the Indian Ocean are elevated in coastal regions (Gardner et al., 2006), with the

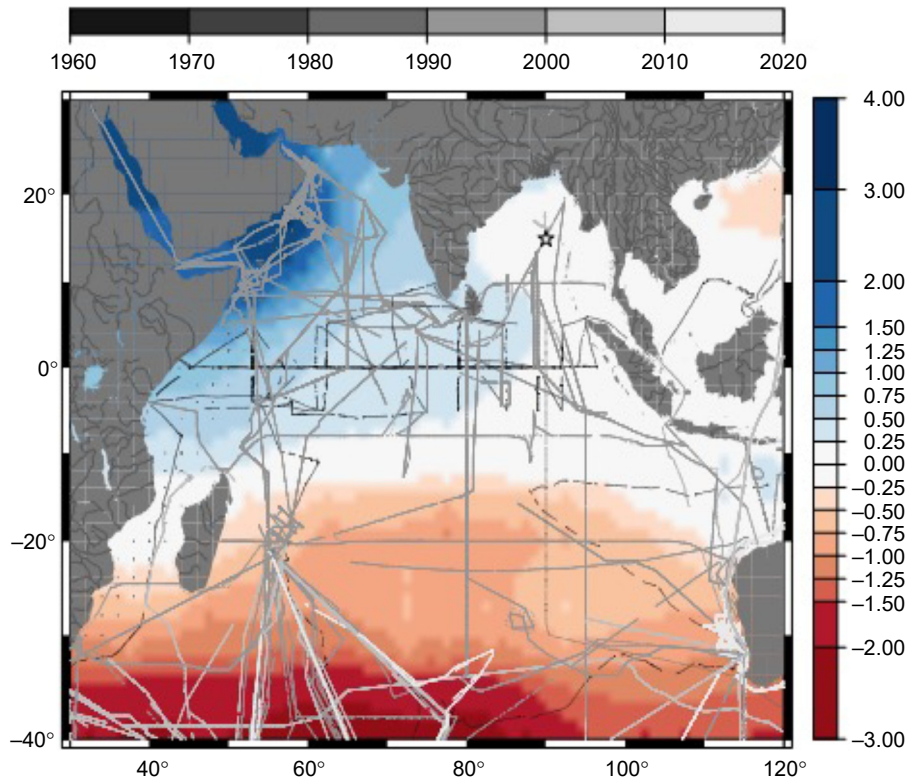


FIG. 10 Annual CO_2 flux ($\text{mol C m}^{-2} \text{yr}^{-1}$) referenced to the year 2000 (Takahashi et al., 2009) over the Indian Ocean. Data points colored by year of collection are overlaid as points. Major rivers are also delineated over the continents. (From Hood et al. (2024a).)

highest POC concentrations observed in the northwestern part of the Arabian Sea and off the coast of Somalia (Hood et al., 2024b; Fig. 11). These high POC values are coincident with high Chl a concentrations (Gardner et al., 2006; Fig. 11). A tongue of elevated POC and Chl a concentration is also associated with the Agulhas retroflection (Gardner et al., 2006; Fig. 11). The lowest POC (and Chl a) concentrations in the Indian Ocean are found in the oligotrophic southern subtropical gyre (Gardner et al., 2006; Figs. 8 and 11).

The spatial and temporal variability in POC export flux in the Indian Ocean is similar to the spatial and temporal variability in POC and Chl a concentration, which is consistent with the idea that organic carbon fluxes are largely controlled by primary production (Hood et al., 2024b; Rixen et al., 2019). However, the spatial variability of POC flux is also strongly influenced by lithogenic matter (ballast) in regions like the Bay of Bengal that have large river inputs (Fig. 7; Hood et al., 2024b; Rixen et al., 2019). In the Arabian Sea, the highest POC fluxes are observed during the Southwest Monsoon. In the Bay of Bengal, POC export flux is generally lower and the seasonal variability is weaker compared to the Arabian Sea (Hood et al., 2024b; Rixen et al., 2019). In eastern-central equatorial waters of the Indian Ocean, ^{234}Th -estimates of POC export flux have revealed elevated values (Subha Anand et al., 2017), which is surprising given the downwelling-favorable equatorial circulation (Hood et al., 2024b; Strutton et al., 2015) and the observed low burial rates of organic carbon (Jahnke, 1996). There are very few POC export flux measurements off western Australia (Hood et al., 2024b). The highest organic carbon burial rates in the southwestern Indian Ocean are observed near the coast off southeastern Africa and in the open ocean in the Agulhas retroflection (Jahnke, 1996). Export fluxes are extremely low in the southern subtropical gyre of the Indian Ocean (Harms et al., 2021; Hood et al., 2024b).

7.3 Productivity and fisheries

The spatial distribution and temporal variability of primary productivity, from regional to basin-scale and seasonal to interannual, have been quantified in the Indian Ocean using satellite ocean color data (Hood et al., 2017, 2024a;

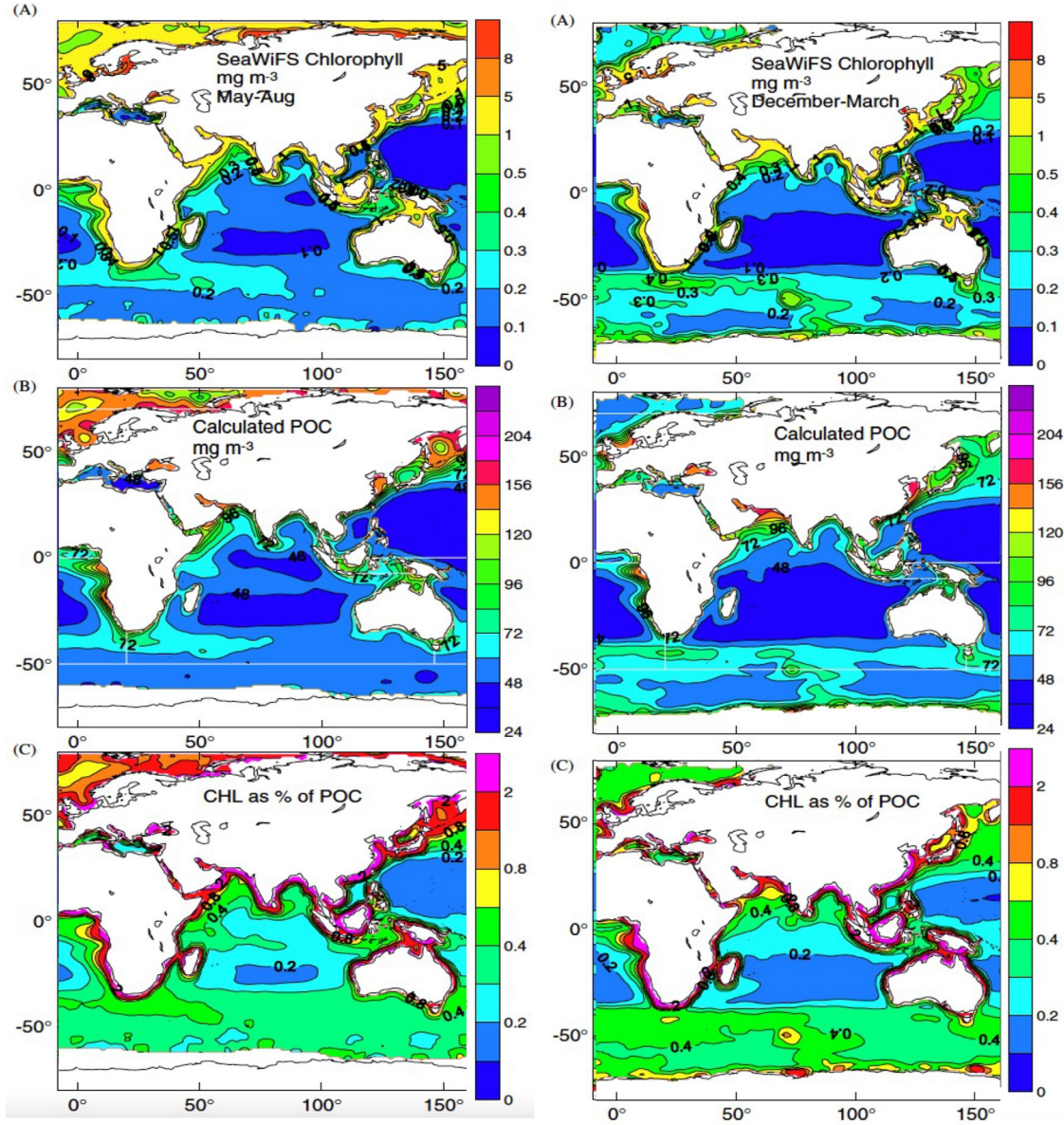


FIG. 11 Distribution of: (a) SeaWiFS Chl (mg m^{-3} , level 3, reprocessing 4 data); (b) average POC (mg m^{-3}) over one attenuation depth calculated from $K_{490}; c_p; \text{POC}$; (c) Chl as a percentage of POC for: (left panels) summer season (1997–2002, May–August, 20 months); (right panels) winter season (1997–2002, December–March, 20 months). White lines in (b) mark boundaries separating ocean basins. (Modified from Gardner et al. (2006).)

Marsac et al., 2024). The spatial patterns in Chla and POC concentration and primary production are described in the previous sections. Goes et al. (2005) used a seven-year (1997–2004) satellite ocean color time series to document an increase in Chla in the Arabian Sea that they attributed to global warming-induced strengthening of upwelling-favorable Southwest Monsoon winds in the western Arabian Sea. Others have argued that the increase in Chla observed by Goes et al. (2005) reflects decadal variability (Prakash et al., 2012; Prakash & Ramesh, 2007; Rixen et al., 2014). In contrast, Roxy et al. (2016) examined a broader area of the northwestern Indian Ocean (including the Arabian Sea) and a longer time series (1997–2013) and came to the opposite conclusion that Chla and primary production has been decreasing due to greater stratification in a warming ocean. Based on a 23-year time series, Marsac et al. (2024) concluded that there is no evidence of any long-term trends in the mean Chla concentration or the areal extent of extremely oligotrophic waters. Clearly, there are conflicting assessments of long-term changes in Chla concentration and primary production in the Indian Ocean that need to be resolved (Hood et al., 2024a).

Several studies have investigated the impact of interannual ENSO and IOD variability on the primary productivity of the basin (Currie et al., 2013; Keerthi et al., 2017; McCreary et al., 2009; Wiggert et al., 2009). For example, the IOD generates significant increases in Chl a concentration and primary production in September along the eastern boundary of the Indian Ocean in tropical waters that are normally highly oligotrophic driven by IOD-induced increases in upwelling off Sumatra (Wiggert et al., 2009). Positive Chl a and productivity anomalies are also apparent in the southeastern Bay of Bengal due to increased upwelling and/or mixing, while negative anomalies are observed over much of the Arabian Sea and SCTR where upwelling is suppressed (Marsac et al., 2024; Wiggert et al., 2009).

In 2019, tuna catches in the Indian Ocean represented 23% of the world tuna and billfish catch (FAO, 2020), and 90% of the neritic tunas are caught by the coastal/artisanal fisheries. The spatial distribution of actively feeding tuna schools is very patchy (Fonteneau, 1986; Marsac et al., 2024; Ravier & Fromentin, 2001, 2004), which reflects the patchiness in the distribution of prey. Patchiness in the distribution of tuna prey can often be related to mesoscale eddies (Tew-Kai & Marsac, 2009), indicating that mesoscale features often promote aggregations of tuna (Digby et al., 1999). Several studies have shown linkages between large-scale climate signals such as the ENSO and IOD and the Pacific Decadal Oscillation and changes in mesoscale activity (e.g., Palastanga et al., 2006; Tew-Kai & Marsac, 2009; Zheng et al., 2018). However, predictions on the future state of mesoscale activity in the ocean are highly uncertain because coupled climate models do not resolve the mesoscale. Based on the present understanding, it seems likely that tropical tunas will face challenging conditions in the open ocean if climate change results in a reduction of mesoscale activity (Marsac et al., 2024).

The distribution of tuna fleets in the Indian Ocean have been observed to shift at interannual timescales, moving from the west to the east in the equatorial Indian Ocean during the 1997–98 IOD (Marsac, 2017; Marsac & Le Blanc, 1998, 1999; Marsac et al., 2024; Fig. 12). As discussed earlier, a positive IOD disrupts the typical seasonal evolution of phytoplankton bloom dynamics over the whole Indian Ocean basin. The purse seiner fleet could not operate efficiently in the western Indian Ocean during the 1997–1998 positive IOD due to the absence of tuna schools and therefore had to move to areas with higher biological productivity in the eastern equatorial Indian Ocean (Fig. 12; Marsac, 2017; Marsac et al., 2024). It appears that the IOD-modulated prey enhancement is a major driver of tuna and tuna fleet relocation in the Indian Ocean.

Multiple studies suggest that tuna fisheries will be significantly impacted on multidecadal timescales by climate change (e.g., Lehodey et al., 2006; Ravier & Fromentin, 2001, 2004). Roxy et al. (2016) predict dramatic declines in net primary production in the northwestern Indian Ocean in the future due to climate change. These changes are also projected to expand over the 21st century (Dueri et al., 2014; Roxy et al., 2024). Thanks to their ability to adapt to changing conditions, tunas will likely survive these environmental changes, unless poor management results in severe depletion of tuna stocks. Regardless, tuna fishers in the Indian Ocean will very likely have to adapt to a significant reshaping of the tuna habitat, both horizontally and in depth (Moustahfid et al., 2018). These changes will almost certainly impact socio-economic conditions in Indian Ocean rim countries that are highly dependent on their domestic tuna fisheries (Marsac et al., 2024).

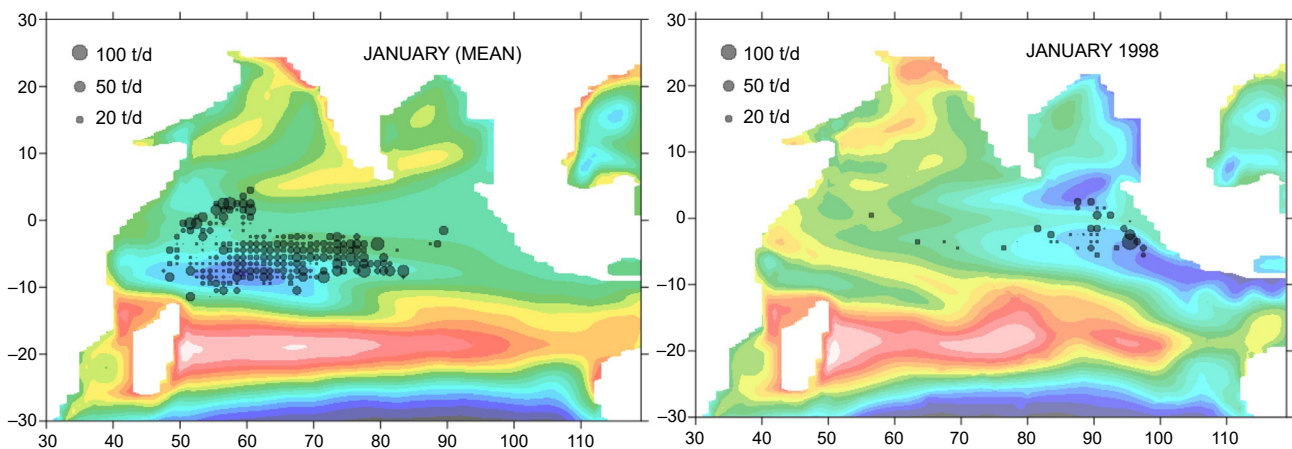


FIG. 12 Distribution of purse seine tuna catch per unit effort (circles, t day^{-1}) on free swimming schools. Left panel: January mean, catch per unit effort 1991–2002 and 20°C isothermal depth (m, color shaded), 1980–2005. Right panel: Catch per unit effort and 20°C depth in January 1998 (positive dipole). (From Marsac et al. (2024).)

8 Summary and conclusions

The seasonally reversing winds drive corresponding ocean surface currents in the northern Indian Ocean, with anticyclonic upwelling circulations during the Southwest Monsoon and cyclonic downwelling circulations during the Northeast Monsoon. And, unlike the Atlantic and Pacific, these winds drive equatorial currents that are, on average, downwelling favorable. The ITF forces and feeds the Leeuwin Current, the only poleward-flowing eastern boundary current in the world. Near-surface waters derived from the ITF and the Tasman leakage south of Australia cross the Indian Ocean basin before feeding into the poleward-flowing Agulhas Current, which sheds eddies into the Atlantic to complete the global thermohaline circulation. Variability in the circulation of the Indian Ocean is the result of complex interactions that are both internal and external to the Indian Ocean on many different timescales, and these interactions are not fully understood.

The monsoon winds also influence SST. As is observed globally, there is lower SST at higher latitudes in the Southern Hemisphere of the Indian Ocean where there is strong surface cooling, whereas in the northern Indian Ocean, there is cooling and lower SST during the Northeast Monsoon due to the influence of the monsoon winds, and warming and higher SST during the rest of the year, especially during the spring and fall inter-monsoon seasons when winds are weak. In contrast, salinity in the Indian Ocean is controlled not only by net air-sea fluxes (evaporation minus precipitation), but also freshwater inflow from large rivers, inflow of relatively fresh waters from the Pacific Ocean via the ITF, and the inflow of saltier waters from the Red Sea and the Persian and Arabian Gulfs. These different drivers combine to give the Indian Ocean a unique salinity distribution, i.e., the Arabian Sea is relatively salty and the Bay of Bengal is relatively fresh, whereas the tropics are relatively fresh, and the southern subtropics are relatively salty. Fresher surface layers also give rise to the formation of barrier layers—especially in the northern Bay of Bengal—that can influence the summer monsoon, development of the IOD, and tropical cyclone-induced upwelling. However, the uncertainty in the freshwater distribution and mixing pathways for riverine input is high, and shallow, salinity-controlled mixed layers and the presence of barrier layers can significantly affect upper-ocean processes in ways that are not fully understood.

The monsoonal wind forcing is also a major driver of biogeochemical and ecological variability with both stimulatory (Southwest Monsoon) and suppressing (Northeast Monsoon) effects throughout many Indian Ocean regions (Arabian Sea, Bay of Bengal, SCTR, Java, central/eastern subtropical gyre). In addition, there are regionally specific processes that significantly modulate the biogeochemical and ecological responses. For example, Fe/Si limitation in the Arabian Sea; freshwater and stratification in the Bay of Bengal; the influence of ITF nutrient inputs, poleward transport, downwelling, and seaward-propagating eddies in the Leeuwin Current; and alongshore-propagating eddies, meanders, upwelling, and poleward transport in the Agulhas Current. The southern subtropical gyre of the Indian Ocean is extremely oligotrophic, yet the Southeast Madagascar bloom occurs during the most stratified season (austral summer/fall). All of these regionally specific processes have significant biogeochemical consequences, yet many have received relatively little research attention.

And don't forget that the South Asian monsoon represents the largest monsoon system on Earth, the Indian Ocean has the world's largest southward meridional heat transport, and the Indian Ocean is characterized by very active intraseasonal variability with far-reaching implications for regional climate. Moreover, the Arabian Sea has the thickest OMZ in the world with globally significant biogeochemical impacts, and some of the lowest pH values in the world have been measured in surface waters of the Indian Ocean. And in 2019, tuna catches in the Indian Ocean represented 23% of the world tuna and billfish catch, with 90% of the neritic tunas caught by the coastal/artisanal fisheries.

Adding to all this complexity (or perhaps a consequence of it) are the conflicting reports as to whether or not primary production is changing in the Indian Ocean in response to global warming, and highly uncertain future projections of climate change impacts on biogeochemistry. Moreover, both interannual studies in the Indian Ocean and multidecadal studies in other ocean basins suggest that fisheries in the Indian Ocean will be significantly impacted by climate change and that these changes will impact socioeconomic conditions in Indian Ocean rim countries that are highly dependent on their domestic tuna fisheries. Clearly, more work needs to be done to reassess the potential effects of climate variability and change on the biogeochemistry and ecology of the entire Indian Ocean basin. This need is made even more pressing by the fact that a significant fraction of the world's population lives in the coastal and interior regions of Indian Ocean rim countries and they are directly impacted by this atmospheric and oceanic variability.

9 Educational resources

- Ocean Data View, free software for plotting oceanographic data. Available at: <https://odv.awi.de>.
- World Ocean Atlas, a collection of objectively analyzed, quality controlled temperature, salinity, oxygen, phosphate, silicate, and nitrate means based on profile data from the World Ocean Database. Available at: <https://www.ncei.noaa.gov/products/world-ocean-atlas>.

- Surface Ocean CO₂ Atlas (SOCAT) is a synthesis of quality-controlled, surface ocean fCO₂ (fugacity of carbon dioxide) observations by the international marine carbon research community. Available at: <https://www.socat.info>.
- Satellite ocean color data. Available at: <https://oceancolor.gsfc.nasa.gov>.
- GFDL Sea Surface Temperature Simulation for the Indian Ocean. Available at: <https://www.youtube.com/watch?v=ZVssbK0K4wc>.
- Educational video of the Asian monsoon. Available at: <https://www.youtube.com/watch?v=RQkOhKEtC20>.
- Wikipedia Page on the Joint Global Ocean Flux Study: https://en.wikipedia.org/wiki/Joint_Global_Ocean_Flux_Study.
- Information on and resources related to the GO-SHIP Program: <https://www.go-ship.org/About.html>.

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Author contributions

RRH conceived and led the chapter overall, with contributions in particular to Sections 1, 2, 3, and 7. HP wrote Section 4.1 and drafted Fig. 3. JS wrote Section 4.3. CCU wrote Sections 5 and 6 and drafted Fig. 7. All authors contributed to the discussion of content and overall chapter structure and provided feedback on the entire chapter.

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