What are Marine Ecological Time Series telling us about the ocean? A status report

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Annex: Directory of Time-series Programmes
7 Indian Ocean

Peter A. Thompson, Todd D. O’Brien, Kirsten Isensee, Laura Lorenzoni, and Lynnath E. Beckley

Figure 7.1. Map of IGMETS-participating Indian Ocean time series on a background of a 10-year time-window (2003–2012) sea surface temperature trends (see also Figure 7.3). At the time of this report, the Indian Ocean collection consisted of 10 time series (coloured symbols of any type), of which two were Continuous Plankton Recorder surveys (blue boxes) and one was estuarine (yellow star). Dashed lines indicate boundaries between IGMETS regions. Uncoloured (gray) symbols indicate time series being addressed in a different regional chapter (e.g. Southern Ocean, North/South Pacific, South Atlantic). See Table 7.3 for a listing of this region’s participating sites. Additional information on the sites in this study is presented in the Annex.

Participating time-series investigators

Uli Bathmann, Frank Conan, Claire Davies, Ruth Eriksen, Mitsuo Fukuchi, Anatzia Genin, Graham Hosie, Jenny Huggett, Takahashi Kunio, Felicity McEnnulty, Anthony Richardson, Malcolm Robb, Don Robertson, Karen Robinson, Yonathan Shaked, Anita Slotwinski, Peter A. Thompson, Mark Tonks, and Julian Uribe-Palomino

Figure 7.2. Major Indian Ocean Currents (adapted from Schott et al., 2009). a) Late Northeast Monsoon (March–April); b) Late Southwest Monsoon (September–October).
Chapter 7 Indian Ocean

7.1 Introduction

With a southern boundary historically ranging from 60°S to the Antarctic continent, the Indian Ocean is the fourth largest ocean with an area of up to 74 million km². As the IGMETS analysis required non-overlapping ocean regions for its spatiotemporal trend calculations, and likewise could only assign each time series to a single region, a boundary of 45°S was used to define and separate the Indian Ocean region from the Southern Ocean region. This modified area, ca. 7800 km wide and stretching from about 25°N to 45°S, had an area of 56.8 million km². Unlike the North Atlantic and North Pacific oceans, the Eurasian landmass in the north precludes high-latitude cooling of surface waters in the northern Indian Ocean. There is, however, some low-latitude exchange of water between the Pacific and Indian oceans via the Indonesian Throughflow. The only large shelf areas are the shallow seas north of Australia, which are regions of strong tidal dissipation. There are a number of significant meridional ridges and several deep basins extending below 5000 m. Within the Indian Ocean, there are several marginal seas, gulfs, and bays, with the Indian subcontinent separating the two most prominent, namely the Arabian Sea and the Bay of Bengal. The Arabian Sea, extending between roughly 10–23°N and 51–74°E, reaches depths >3000 m over most of its area and has two important regions: the Gulf of Aden to the southwest, connecting with the Red Sea, and the Gulf of Oman to the northwest, which connects with the Persian Gulf. The Bay of Bengal, which occupies an area of 2 172 000 km², receives input from a number of rivers, the most important being the Ganges–Brahmaputra river system.

This river system delivers large quantities of sediment to the Bengal Fan, which causes depth in the Bay of Bengal to decrease gradually from 4000 m south of Sri Lanka to ≤ 2000 m at 18°N (Tomczak and Godfrey, 2003; Galy et al., 2007). The circulation of the northern Indian Ocean is dominated by the monsoons and their seasonal switching between strong southwest winds during June–September and weaker northeast winds in October–March (Talley et al., 2011). The winds drive a reversal of the surface currents in both the Bay of Bengal and Arabian Sea (Figure 7.2). In particular, the western boundary current in the Arabian Sea, the Somali Current, flows northward in boreal summer and then reverses to flow largely southward in boreal winter. The summer pattern of wind and current stimulates a strong current and upwelling along the coast from Somalia to Oman (Beal and Chereskin, 2003). The northern hemisphere summer monsoons provide intense rainfall between 10–20°N and 70–120°E, which is of considerable importance to agriculture in the region. The tropical southern Indian Ocean typically has a wet monsoon between November and March that produces rainfall around 5–10°S and across the entire basin. In contrast, the austral winter (April–October) monsoon creates some upwelling along the west coasts of Java and Sumatra (Wyrtki, 1962) that tends to be strongest during El Niño events (Susanto et al., 2001). The Agulhas Current is a strong western boundary current that transports 70 Sv poleward at 31°S, with flow that varies from 9 to 121 Sv at velocities of up to 2 m s⁻¹ at 35°S (Boebel et al., 1998; Bryden et al., 2005). As the Agulhas Current reaches the southern tip of Africa, most of the water is reflected to the east by the “westerly wind drift” (Lutjeharms and van Ballegooeyen, 1988) along the subtropical front (STF). However, a modest volume of warm, saline Indian Ocean water is transported into the South Atlantic through the Agulhas leakage. It has been suggested that the Agulhas leakage is an important component of the climate system (Beal et al., 2011). The STF is a long narrow feature stretching from the east coast of South America, through the South Atlantic, Indian, and South Pacific oceans to the west coast of South America. It separates the warm, salty subtropical waters from colder, fresher Antarctic waters. The STF region is high in eddy kinetic energy and develops large coccolithophorid blooms in summer (Balch et al., 2011, 2016).

The Leeuwin Current forms the eastern boundary current for the Indian Ocean and is unusual, as it flows poleward albeit with much less volume (ca. 5 Sv) than the Agulhas Current (Godfrey and Ridgway, 1985). The poleward flow of warm, fresher water typically peaks in May or June (Figure 7.2). The buoyant Leeuwin Current also flows eastward through the Great Australian Bight along the south coast of Australia in winter suppressing upwelling along its length (Ridgway and Condie, 2004). The oligotrophic southern Indian Ocean central gyre has previously been estimated to be growing in size in response to climate drivers (Jena et al., 2013; Signorini et al., 2015).
Figure 7.3. Annual trends in Indian Ocean sea surface temperature (SST) (a) and sea surface chlorophyll (CHL) (b), and correlations between chlorophyll and sea surface temperature for each of the standard IGMETS time-windows (c). See “Methods” chapter for a complete description and methodology used.
7.2 General patterns in temperature and phytoplankton biomass

For the entire Indian Ocean, the overwhelming trend in temperature has been upwards (Figure 7.3; Table 7.1). During 1983–2012, ca. 98% of the Indian Ocean was warming. 81.9% showed a significant temperature increase >0.1 and ≤ 0.5°C decade⁻¹ (Table 7.1). This was the greatest proportion of warming for any ocean on the planet (Chapter 10) and is associated with a range of climate cycles including the relatively long positive phase of the Interdecadal Pacific Oscillation (Han et al., 2014). Over this same time-period, only 0.5% of the Indian Ocean was found to be significantly cooling (Table 7.1).

The analysis over multiple 5-year time-windows shows that temperature changes were more rapid and more variable over shorter intervals (Table 7.1; Figure 7.3). For example, 19.3% of the Indian Ocean was warming at a high rate of > 1.0°C decade⁻¹ over the 5-year window (2008–2012), but this rate was not observed over the 15-year time-window (1998–2012). Between 2008 and 2012, the statistically significant rates of warming ranged from −1.0 to +1.0°C decade⁻¹ (Figure 7.4a). Over the longer temporal window from 1983 to 2012, these rates tended to be 5–10-fold less variable ranging from −0.5 to +0.5°C decade⁻¹ (Figure 7.4b). Notwithstanding the influence of statistics itself, the declining variability in the rate of temperature changes suggests that the shorter-term climate cycles predominately have periods that are less than 30 years. Consider that the proportion of the Indian Ocean warming rose 0.75% year⁻¹ as the time series lengthened from 0 to 20 years, then only 0.23% year⁻¹ as the time series was extended another 10 years (Table 7.1). The slowing in the spatial expansion and consolidation of the rate at an intermediate value suggests that the variability associated with shorter-term climatic signals in the Indian Ocean (e.g. Indian Ocean Dipole, El Niño Southern Oscillation) is reduced when using a linear model and more than 20 years of data.

More than 79% of the Indian Ocean experienced a decline in surface chlorophyll $a$ over the 15-year time-period from 1998 to 2012, while only 20.3% had an increase (Figure 7.3). The proportion of the Indian Ocean experiencing a decline in chlorophyll $a$ was the greatest for any ocean (Table 7.1; Chapter 10) and highly correlated with warming (Figure 7.3). The main areas of cooling and increasing chlorophyll $a$ were associated with just four regions: (i) the Arabian Sea and surrounding areas (Red Sea and Persian Gulf), (ii) along the southwest coasts of Sumatra and the Sunda Islands, (iii) south from Madagascar, and (iv) along the subtropical front (STF) at ca. 40–45°S (Figure 7.3). The spatial pattern of increasing chlorophyll $a$ tended to vary depending on the temporal window considered, but was arguably most consistent along the STF.

There was significant cooling in the Red Sea, Arabian Sea, Persian Gulf, and along the coasts of Oman and Yemen between 1998 and 2012 (Figure 7.3). Almost none of this regional cooling was evident over the longer 30-year time-window, suggesting that it was strongly influenced by relatively short climatic cycles such as the Indian Ocean Dipole (IOD) and ENSO. It is also possible that this regional cooling was caused by greater seasonal upwelling associated with an increased frequency of stronger IOD and ENSO events that were predicted as a response to climate change (Cai et al., 2015).

The Arafura and Timor seas north of Australia also trended colder during 1998–2012. It is probable that this colder trend resulted from the strong positive IOD in 2011 and 2012 and a weakening La Niña (Meyers et al., 2007). The IOD primarily affects the pelagic ecology of this region through upwelling favourable winds (Currie et al., 2013; Kämpf, 2015).

There was a relatively broad region of cooling off southeast Africa below Madagascar where the Southeast Madagascar and Agulhas currents normally transport considerable amounts of warm water southward (Yamagami and Tozuka, 2015). Although the Southeast Madagascar Current flow is associated with ENSO, this cooling was consistent throughout the different time-windows (Figure 7.3), suggesting a longer-term effect. The source of this surface cooling is unclear. A possibility is the multidecadal rise in subtropical wind stress and increased Southeast Madagascar Current flow (Backeberg et al., 2012). The increased South Madagascar Current resulted in a substantial rise in eddy kinetic energy and, in turn, promoted greater vertical mixing for this region (Backeberg et al., 2012). Similarly, the region showed mesoscale patches of warming, consistent with increased Agulhas and Southeast Madagascar Current flow.
Table 7.1. Relative spatial areas (% of the total region) and rates of change within the Indian Ocean region that are showing increasing or decreasing trends in sea surface temperature (SST) for each of the standard IGMETS time-windows. Numbers in brackets indicate the % area with significant ($p < 0.05$) trends. See “Methods” chapter for a complete description and methodology used.

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<tr>
<td><strong>Area (%) w/ increasing SST trends</strong> ($p &lt; 0.05$)</td>
<td>66.7% (26.4%)</td>
<td>76.0% (47.4%)</td>
<td>82.6% (58.1%)</td>
<td>96.7% (87.9%)</td>
<td>96.2% (89.4%)</td>
<td>97.8% (91.9%)</td>
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<tr>
<td><strong>Area (%) w/ decreasing SST trends</strong> ($p &lt; 0.05$)</td>
<td>33.3% (4.2%)</td>
<td>24.0% (4.5%)</td>
<td>17.4% (4.6%)</td>
<td>3.3% (0.8%)</td>
<td>3.8% (0.7%)</td>
<td>2.2% (0.5%)</td>
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<td>&gt; 1.0°C decade$^{-1}$ warming ($p &lt; 0.05$)</td>
<td>24.8% (19.3%)</td>
<td>7.8% (7.8%)</td>
<td>0.0% (0.0%)</td>
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<tr>
<td>0.5 to 1.0°C decade$^{-1}$ warming ($p &lt; 0.05$)</td>
<td>18.8% (6.1%)</td>
<td>19.0% (18.3%)</td>
<td>6.6% (6.5%)</td>
<td>2.9% (2.9%)</td>
<td>0.1% (0.1%)</td>
<td>0.0% (0.0%)</td>
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<tr>
<td>0.1 to 0.5°C decade$^{-1}$ warming ($p &lt; 0.05$)</td>
<td>18.1% (0.9%)</td>
<td>38.5% (20.9%)</td>
<td>60.3% (49.2%)</td>
<td>86.2% (83.3%)</td>
<td>84.2% (83.3%)</td>
<td>82.2% (81.9%)</td>
</tr>
<tr>
<td>0.0 to 0.1°C decade$^{-1}$ warming ($p &lt; 0.05$)</td>
<td>5.0% (0.0%)</td>
<td>10.7% (0.5%)</td>
<td>15.7% (2.3%)</td>
<td>7.6% (1.7%)</td>
<td>11.9% (6.0%)</td>
<td>15.6% (10.0%)</td>
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<tr>
<td>0.0 to –0.1°C decade$^{-1}$ cooling ($p &lt; 0.05$)</td>
<td>4.1% (0.0%)</td>
<td>8.7% (0.1%)</td>
<td>7.1% (0.1%)</td>
<td>2.0% (0.1%)</td>
<td>2.7% (0.1%)</td>
<td>1.7% (0.1%)</td>
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<tr>
<td>–0.1 to –0.5°C decade$^{-1}$ cooling ($p &lt; 0.05$)</td>
<td>15.0% (0.5%)</td>
<td>12.7% (2.4%)</td>
<td>9.7% (3.9%)</td>
<td>1.2% (0.6%)</td>
<td>1.1% (0.6%)</td>
<td>0.5% (0.3%)</td>
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<td>–0.5 to –1.0°C decade$^{-1}$ cooling ($p &lt; 0.05$)</td>
<td>8.7% (1.5%)</td>
<td>2.3% (1.8%)</td>
<td>0.6% (0.6%)</td>
<td>0.1% (0.1%)</td>
<td>0.0% (0.0%)</td>
<td>0.0% (0.0%)</td>
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<tr>
<td>&gt; –1.0°C decade$^{-1}$ cooling ($p &lt; 0.05$)</td>
<td>5.5% (2.2%)</td>
<td>0.3% (0.2%)</td>
<td>0.1% (0.1%)</td>
<td>0.0% (0.0%)</td>
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Figure 7.4. The range of SST trends observed over different temporal windows in the Indian Ocean. (a) From 2008 to 2012, 33.3% (66.7%) of the area was cooling (warming); while 4.2% (26.4%) was cooling (warming) significantly ($p < 0.05$). Rates of cooling and warming ranged from –0.50 to +0.50°C year$^{-1}$. (b) Over the longer-term from 1983 to 2012, the rates of temperature change were much more constrained, ranging from –0.05 to 0.07°C year$^{-1}$. Over this 30-year period, only 2.2% (97.8%) of the area was cooling (warming), while 0.5% (91.9%) was cooling (warming) significantly ($p < 0.05$).
It is suggested that stronger increasing eddy kinetic energy observed across the Agulhas retroflection (Swart et al., 2015) added to this spatial mosaic of mesoscale variability in warming and cooling trends. The spatial pattern of patchy cooling extended across the entire Indian Ocean at ca. 40°S, suggesting that this effect can be seen a long way eastward along the edge of the STF. There was also cooling south of Africa between 20°–60°E and 50°–60°S. This probably relates to the long-term positive trend in the SAM (Swart et al., 2015), which increases the westerly flow along 60°S, decreasing SST to lower-than-normal values (Lovenduski and Gruber, 2005; Verdy et al., 2006).

Between 1998 and 2012, there were upward trends in chlorophyll a within the southern Red Sea, Persian Gulf, and patches through the Gulf of Oman and Arabian Sea off the coasts of Yemen and Oman that largely coincide with regions of declining SST. The latter are regions of upwelling that are known to respond to stronger winds during the summer southwest monsoon season (Yi et al., 2015). The trends in chlorophyll a are clearly dependent on the temporal window selected with downward trends for most of this region over the shorter 10-year window from 2003 to 2012. Over the shortest temporal window from 2008 to 2012, trends in chlorophyll a were mixed across the region, although quite strongly negative in the Persian Gulf (Figure 7.3). A 60-year reconstruction of summer blooms based on SST suggested that regional summer chlorophyll a concentrations might have peaked during the very strong upwelling of 1966 (Roxy et al., 2016).

The rise in chlorophyll a from 1998 to 2012 along the coasts of Java and Sumatra, weakly through the Timor and Arafura seas, and into the Gulf of Carpentaria was in regions that respond to positive ENSO and IOD conditions (Currie et al., 2013). These climatic indices were positive on average and trending positive throughout this 15-year period. Indeed, the 15-year period started with predominant El Niño episodes and progressed to include several moderate-to-strong La Niña events in the latter half. The mechanisms potentially driving an increase in chlorophyll a across this diverse region include upwelling favorable winds off Java and Sumatra, increased runoff into the Gulf of Carpentaria, and greater interocean exchange from the Pacific to the Indian Ocean for the shallow Arafura and Timor seas.

There were patches of increased chlorophyll a southeast of Africa and south of Madagascar observed across all three temporal windows. These patches were also evident across the Indian Ocean sector near the STF. In these regions, the patchy spatial distribution of the increasing chlorophyll a has a strong resemblance to the spatial pattern of increased SSH and increasing eddy kinetic energy observed during 1993–2009 (Backeberg et al., 2012). Thus, the spatial nature of the increased phytoplankton in this region can be hypothesized to be associated with increased eddy pumping (Falkowski et al., 1991). Eastward, along the STF, it is likely that eddies with increased deep mixing at this convergence zone have created this mosaic of increased and decreased chlorophyll a. The most pronounced increases in phytoplankton along the Indian Ocean sector of the STF were observed close to Tasmania, where the STF interacts with a strengthening East Australian Current (Figure 7.3).

Mostly at latitudes >45°S, although occasionally closer to the equator, there were scattered regions where chlorophyll a was trending upwards. A positive SAM has been associated with changes in the ocean meridional overturning circulation, including increased upwelling of nutrient-rich waters in the region of 60°S (Hall and Visbeck, 2002), as well as a shallower surface mixed layer depth (Lovenduski and Gruber, 2005). The ENSO also exerts an influence on phytoplankton in this region; for example, when a positive SAM aligns with a positive ENSO event, eddy kinetic energy increases significantly (Langlais et al., 2015). Between 35 and 60°S, the spatial patterns of SST and chlorophyll a trends were quite consistent across all temporal windows examined. However, during 2003–2012, the warming and greening was broader, but patchier. The trend and average condition of both ENSO and SAM cycles were positive during these 10 years, factors that have been previously linked to increased chlorophyll a in these regions (Lovenduski and Gruber, 2005). During this time, the deeply mixed surface layer of the Subantarctic Zone (SAZ) and Polar Front Zone (PFZ) apparently became more conducive for phytoplankton growth. The mechanism for this response is hypothesized to be the upwelling of iron or shallowing of the surface mixed layer (Carranza and Gille, 2015).

During 2008–2012, the south Indian central gyre cooled, and a broad increase in chlorophyll a was also observed in that area. The explanation for this strong reversal of the longer-term trend is not evident at this point and may merit further investigation. Indeed, most research has shown the subtropical gyres to be expanding, warming, and declining in phytoplankton (Jena et al., 2013; Signorini et al., 2015).
Table 7.2. Five-year trends (TW05, 2008–2012) in the time series of observations from \textit{in situ} sites in the South Pacific (not including Continuous Plankton Recorder sites).

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\( p > 0.05 \) negative \( p < 0.05 \) positive
7.3 Trends from in situ time series

There is a dire paucity of ecological long-term time-series data openly available for the Indian Ocean (Figure 7.5). Most of the small number of time series extend for <10 years. The lack of these data makes it nearly impossible to even describe the current status of the ecosystem and its pelagic biota or their temporal trends at the basin scale.

The most heavily sampled region is the coastal zone around Australia, extending from Darwin (12°S 131°E) to Tasmania (42°S 148°E), mostly in the Longhurst biogeographical province No. 2 (Longhurst, 2007), the Indonesian and Australian Coastal Province (Figure 7.6). The footprint of these stations indicates that they can represent temperature and chlorophyll a temporal trends across a large portion of the adjacent shelf (Oke and Sakov, 2012; Jones et al., 2015). Within the Australian portion of this province, the Leeuwin Current (LC) is a strong influence bringing warm, low-salinity, silicate-rich water southward from the tropics (Thompson et al., 2011). During 2008–2012, two stations, Esperance and Rottnest Island (Table 7.2), showed the most signs of greater LC effects, such as increasing SST, decreases in salinity, declining dissolved oxygen (DO), and fewer diatoms (Figure 7.5). Copepod biomass rose at both these sites, as did chlorophyll a at Esperance (Figure 7.5). The latter result is in strong contrast to the general trend of declining chlorophyll a within this region (Figure 7.3), possibly reflecting a more localized effect at this nearshore station or the increased intrusion of productive STF eddies onto the shelf (Schodlok et al., 1997; Cresswell and Griffin, 2004). The nearby estuarine site in the Swan River had 10-year trends of increasing temperature, chlorophyll a, and DO, but declines in phosphate and dinoflagellates. These are strongly influenced by declining rainfall (Thompson et al., 2015). The Kangaroo Island site is also located in Longhurst No. 2 near the extreme eastern end of the Indian Ocean. This site is influenced by wind-driven upwelling with weak links to ENSO and SAM (Nieblas et al., 2009). The site showed declines in nitrate, zooplankton, diatoms, and the diatom/dinoflagellate ratio, with a modest increase in salinity from 2008 to 2012 (Table 7.2; Figure 7.5). The declines in the proportion of diatoms and diatom/dinoflagellate were found at three stations along the bottom of Australia. Consistent with satellite data, the time series from the Gulf of Elat site in the northern Red Sea showed a strong increase in temperature and dissolved oxygen and a weak increase in chlorophyll a between 2008 and 2012 (Table 7.2).

Salinity, nitrate, phosphate, and silicate all trended down over the 5-year time-window between 2008 and 2012. The only other time series currently available from the Indian Ocean is at the western extreme at Mossel Bay on the Agulhas Bank. This is one of two regions on the Bank that have been sampled since 1979 (Hutchings et al., 1995). There was a strong negative trend in zooplankton biomass over the 15-year time-window from 1998 to 2012 (Table 7.2).

In summary, across the few in situ time series available for the Indian Ocean, the trends in temperature were generally upward, while salinity trended down. Trends in nitrate were also generally downward, but other nutrients and biology were variable. These results suggest that local or shorter-term complexities in the oceanography are important in determining ecosystem responses even in the presence of a pervasive warming trend.

7.4 Consistency with previous analysis

Over the 15-year time-period from 1998 to 2012, the average IOD and ENSO indices were positive and trending positive, while the SAM was positive, but trending slightly more negative. These shorter climatic cycles impact most significantly on the Indian Ocean environment and its ecology. The environmental impacts tend to oscillate, with some level of periodicity associated with these climate cycles (Eccles and Tziperman, 2004; Yamagata et al., 2003; Fogt et al., 2009).

The ENSO climatic driver is the primary source of interannual variability in climate throughout most of the Pacific Ocean and portions of the Indian Ocean (Weiqing et al., 2014; Cai et al., 2015). It is a coupled ocean and atmosphere cycle that can be represented by the Southern Oscillation Index (SOI) based on the difference in surface air pressure between Tahiti and Darwin (Australia).

A persistently positive SOI (lower pressure over Darwin than Tahiti) is a La Niña event. During such an event, strong Pacific trade winds and surface currents push warm water in the tropics westward causing increasing SST, increased steric height, greater rainfall, and reduced SSS. The increased volume of seawater transported from the Pacific Ocean to the Indian Ocean by a La Niña event has been estimated at 5 Sv (Meyers, 1996), and this contributes to a stronger Leeuwin Current.
Figure 7.5. Map of Indian Ocean region time-series locations and trends for select variables and IGMETS time-windows. Upward-pointing triangles indicate positive trends; downward triangles indicate negative trends. Gray circles indicate time-series sites that fell outside of the current study region or time-window. Additional variables and time-windows are available through the IGMETS Explorer (http://IGMETS.net/explorer). See “Methods” chapter for a complete description and methodology used.
The intensity of the IOD is represented by an anomalous SST gradient between the western equatorial Indian Ocean (50°-70°E and 10°S-10°N) and the southeastern equatorial Indian Ocean (90°-110°E and 10°S-0°N; Saji et al., 1999). This gradient is also known as the Dipole Mode Index (DMI). Typically, significant anomalies appear around June, intensify in the following months, and peak in October. During a positive IOD, anomalously strong winds push warm water west towards Africa, decreasing upwelling in that region, and increasing upwelling (lower SST) along the west coast of Sumatra (Alory et al., 2007). A positive IOD is also associated with reduced rainfall in Indonesia and northern Australia. Lastly, during a positive Southern Annular Mode (SAM) event, the southern hemisphere westerly winds tend to move farther southward and increase in intensity (Gong and Wang, 1999; Thompson et al., 2000). This results in stronger cold water upwelling at high latitudes, anomalous downwelling around 45°S, and a strengthening of the Antarctic Circumpolar Current (Hall and Visbeck, 2002; Oke and England, 2003). There is significant temporal variability in the SAM, but over the last 50+ years, the SAM index has been positive and on the rise (Thompson et al., 2000; Abram et al., 2014); this rise has been linked to anthropogenic factors that include ozone depletion (Fyfe et al., 1999).

The analyses presented here are broadly consistent with previous work in the Indian Ocean in terms of warming and the biological responses. However, there are also some discrepancies. For example, some studies of phytoplankton biomass using remote sensing have reported declines in parts or most of the north Indian Ocean (Gregg and Rousseaux, 2014; Roxy et al., 2016), in the Equatorial Indian Ocean (Gregg and Rousseaux, 2014), and in the central gyre of the southern Indian Ocean (Jena et al., 2013; Signorini et al., 2015). The choice of time and space scale used in the analysis clearly influences the result. Shorter time-periods are prone to detecting responses to shorter climatic cycles. For example, the 5-year analysis presented here showed increasing chlorophyll a in the south Indian Ocean central gyre (Longhurst 33) during 2008-2012, suggesting a possible link to an intermediate climate cycle (e.g. ENSO or IOD). In addition, the analysis of trends at smaller spatial scales is clearly demonstrating important patterns that can be overlooked if averaging at a larger spatial scale.

Long-term changes to sea surface temperature, surface salinity, and dissolved oxygen in the Indian Ocean have all been previously investigated (e.g. Durack and Wijffels, 2010; Stramma et al., 2010; Cai et al., 2015). The physical changes in the Indian Ocean are the best studied and are the best resolved through a combination of considerable observational effort and modeling. Chemical and biological studies that are sufficient to detect significant temporal trends are very scarce. The long-term trends of increasing CO2 and decreasing O2 at depth tend to change slowly making reasonable predictions of their trends possible from low-frequency sampling. The Bay of Bengal has a shallow low DO layer that extends south of the equator and is expanding (Stramma et al., 2010), with the potential to disrupt fisheries (Stramma et al., 2012). This low DO layer appears to be high in nitrate and CO2, and low in pH (Waite et al., 2013). The seasonal dynamics of other chemical and the biological components, especially in the euphotic zone, require monthly sampling to resolve trends (Henson, 2014).

The lack of biological monitoring in the Indian Ocean severely limits our ability to understand the effects of climate variability on ecological processes and biodiversity (Dobson, 2005). In addition, the heterogenetic spatial and temporal distributions of biota can make it much more challenging to detect long-term trends or broad spatial patterns. For example, the seasonal variability in phytoplankton biomass can be significant and change annually, requiring sustained and consistent sampling to detect any type of longer-term trend. Thus, detecting a relatively small climate signal requires long-term, carefully designed sampling regimes. Trends in the basic biology of phytoplankton, primary production, zooplankton, and secondary production are known from only a few points in the Indian Ocean. More numerous are reports of significant and rapid range expansions by pelagic biota in response to the changing climate (McLeod et al., 2012; Sunday et al., 2015). The lack of fisheries-independent stock assessments, the potential for unreported fishing effort, and the instability in catch per unit effort (cpue) means that a robust measure of trends for many populations of fish species is not available at the basin scale.

In the next section, we present some of the most notable changes that have been reported in the literature for the Indian Ocean and compare them to the available data for this report using a geographic framework (Longhurst, 2007).
Figure 7.6. The Indian Ocean and surrounding marginal seas separated into biogeographical provinces after Longhurst (1995, 2007). The provinces are based on the types of physical forces that determine the pelagic ecology, particularly the distribution of phytoplankton (et al., 2013). Indian Ocean provinces include the: Northwest Arabian Upwelling (17), Red Sea and Persian Gulf (19), East Africa Coastal (18), Australia–Indonesia Coastal (2), Subtropical Convergence (52), Subantarctic (53), East India Coastal (11), West India Coastal (22), Archipelago Deep Basin (29), India Monsoon Gyres (32), and India Subtropical Gyre. (33).

7.4.1 Northwest Arabian upwelling province (Longhurst 17), Red Sea and Persian Gulf province (Longhurst 19)

A series of cruises to this region during the 1990s produced significant insights into the pelagic ecology (Smith, 2005), especially the biological responses to monsoonal forcing (Wiggert et al., 2005). Unfortunately, there are limited in situ time-series data to assess long-term ecological trends. A regional peak in satellite chlorophyll a was observed in 2003 associated with a negative SLA and low SST (Prakash et al., 2012). Similarly, a very low SST and strong upwelling was observed in 1966 (Roxy et al., 2016). The patchy increases in satellite chlorophyll a scattered across the northwest Arabian upwelling province seen over the 5- and 15-year temporal windows may reflect positive IOD and negative ENSO events during the early monsoon season (Currie et al., 2013). At this time, the considerable interannual variability and local effects of climate cycles make it difficult to conclude whether the region will experience a longer-term trend towards increased upwelling under prolonged climate change (Bakun, 1990; Goes et al., 2005; Prakash and Ramesh, 2007; Narayan et al., 2010; Sydeman et al., 2014). The significant positive chlorophyll a trend in the Gulf of Oman and Persian Gulf, especially observed in the 15-year time-window, is consistent with reports of large blooms of the green form of Noctiluca scintillans (Gomes et al., 2014). This mixotrophic dinoflagellate, which grows fastest when grazing, has been increasing globally (Harrison et al., 2011). Gomes et al. (2014) linked the rise in N. scintillans with eutrophication and indicated that it was coincident with declining subsurface dissolved oxygen. The 15-year trend of increasing satellite-detected chlorophyll a in the Red Sea reported herein has been investigated by Raitos et al. (2015), who suggest that the physical mechanism stimulating primary production is related to ENSO climate.
variability in monsoonal winds and nutrient injection from the Indian Ocean, allowing the southern Red Sea to bloom. Unfortunately, the only IGMETS time-series data available from the Red Sea is from the Gulf of Eilat in the far north and well away from this 15-year trend of increasing phytoplankton. In the Gulf of Eilat, the warming and declining nutrients are consistent with the generally expected responses to climate change (Laufkötter et al., 2015).

7.4.2 East Africa coastal province (Longhurst 10)

The zooplankton biomass on the Agulhas Bank has declined 57% since 1988 (J. Huggett, pers. comm.). In particular, abundance of the large calanoid zooplankton Calanus agulhas has decreased significantly. Available information suggests that increasing SST and predation by anchovy (Engraulis capensis) and sardine (Sardinops sagax) may be the primary causes (J. Huggett, pers. comm.), a relationship that has been observed nearby on the west coast of Africa (Verheye et al., 1998). Increasing SST has been frequently associated with declining populations of large macrozooplankton (Daufresne et al., 2009). This province has spatially heterogeneous trends in physical and ecological characteristics that are apparently associated with changing coastal boundary currents (Backeberg et al., 2012).

7.4.3 Australia–Indonesia coastal province (Longhurst 2)

Lower SST, greater upwelling, and more chlorophyll a off Java and Sumatra have been previously linked to a positive IOD and negative ENSO during September–November (Currie et al., 2013). Based on in situ data from Rottnest Island (1951–2002), a warming trend of ca. 0.012° year⁻¹ has been reported for the west coast of Australia (Thompson et al., 2009), which is consistent with the 30-year SST trend reported here. The La Niña event of 2011 was associated with a pronounced increase in water temperature (Feng et al., 2013), a decline in zooplankton biomass, and significantly more of the pico-plankton Prochlorococcus and Synechococcus, which are most abundant in the tropics (Thompson et al., 2015).

7.4.4 Subtropical convergence province (Longhurst 52)

The subtropical convergence and the subtropical front separates the more saline subtropical waters from the fresher subantarctic waters. The physical drivers are spatially and temporally dynamic (Graham and de Boer, 2013).

The province stretches across the South Atlantic, Indian, and South Pacific oceans and has high eddy kinetic energy and substantial westward flow. In this report, the STF showed evidence of increased chlorophyll a across the South Atlantic, South Pacific, and South Indian oceans. In the Indian Ocean, it is possible that this increase is related to large blooms of coccolithophores, which occur in mid-summer (Brown and Yoder, 1994; Balch et al., 2011). There is evidence of a diversity of Emiliania huxleyi morphotypes (Cubillos et al., 2007; Cook et al., 2011) and a poleward expansion of these taxa (Winter et al., 2014).

7.4.5 Subantarctic province (Longhurst 53)

This province is only rarely within the 45°S limit ascribed to the Indian Ocean for the purpose of this report (Orsi et al., 1995). Unfortunately, in situ time-series data from this province and within the Indian Ocean sector are very rare. One exceptional dataset was published by Hirawake et al. (2005) from regular cruises between Tokyo and the Antarctic continent once a year. The authors reported an increase in chlorophyll a over their time series (1965–2002, a trend which is consistent with the remotely-sensed data analyzed herein.
7.5 Conclusions

The Indian Ocean had the greatest extent of warming of all oceans, with 91.8% of its area showing a significant positive trend over 30 years, compared with the Atlantic (88.6%), Pacific (65.9%), Arctic (79.2%), and Southern (31.8%) oceans. In addition to having a high degree of warming, the Indian Ocean also had the greatest proportion of its area (55.1%) showing a significant \( p<0.05 \) decline in chlorophyll \( a \) between 1998 and 2012. The 51 million km\(^2\) of warming in the Indian Ocean may reflect the fact that the northern Indian Ocean is landlocked at ca. 25°N and, therefore, has no high-latitude, seasonal deep mixing to supply cold surface water.

Across the entire Indian Ocean, a few relatively small and mostly upwelling regions, previously identified as productive (Carr et al., 2006), have shown remarkable resilience to warming or declining chlorophyll \( a \) over the past 30 or 15 years, respectively. These regions off Sumatra and Java, off Somalia and Oman, in the Arabian and Red Seas, and along the STF all showed areas of temperature stability or decline and an increase in chlorophyll \( a \) during 1998–2012. Upwelling was predicted to intensify under global warming (Bakum, 1990), and there is evidence of this happening in the Indian Ocean. The rise in phytoplankton along the STF is consistent with some coarse-scale modeling of the ocean in 2090 (Marinov et al., 2013), although the mechanism and proposed taxonomic shift towards diatoms cannot be confirmed. It should be noted that these regions are the overwhelming minority of the Indian Ocean, with only 0.5% of the area tending significantly downward in temperature over 30 years and 4.8% trending significantly upward for chlorophyll \( a \) over a 15-year time-period.

Given the spatial scale of warming in the Indian Ocean, it seems likely that climate impacts on marine ecosystems will be most pronounced in this ocean. At the same time, the Indian Ocean has very few \textit{in situ} biogeochemical time series that can be used to assess impacts of climate change on biota or biodiversity. The few time series that exist in the IGMETS database are all on the continental shelves of just two continents, leaving vast areas completely unmonitored. Still, new insights have arisen from examining trends over sequential time-steps and from the effort to couple physical observations with observed biological responses, including \textit{in situ} observations. These \textit{in situ} observations allow unprecedented insights into trends in ecology driven by climate. Previous research on ecological responses of the Indian Ocean to climate change at the basin-scale have been largely model-based (Currie et al., 2013), limited to trends detected by remote sensing, or a few local case studies. This IGMETS report is the first effort to bring multiple \textit{in situ} time series together to provide a global synthesis and basin-scale comparisons of long-term trends.
Table 7.3. Time-series sites located in the IGMETS Indian Ocean region. Participating countries: Australia (au), Israel (il), and South Africa (za). Year-spans in red text indicate time series of unknown or discontinued status.

| No. | IGMETS-ID | Site or programme name | Year-span | T | S | Oxy | Ntr | Chl | Mic | Phy | Zoo |
|-----|------------|------------------------|-----------|---|---|-----|-----|-----|-----|-----|-----|-----|
| 1   | au-10101   | Swan River Estuary: S01 Blackwall Reach (Southwestern Australia) | 1994–present | X | X | X | X | X | - | X | - |
| 2   | au-40114   | SO-CPR Aurora 140-160-B4245 (Southern Ocean) see Southern Ocean Annex A4 | 2008–present | X | X | - | - | X | - | - | X |
| 3   | au-40205   | AusCPR MEAD Line (Australian Coastline) see Southern Ocean Annex A4 | 2010–present | - | - | - | - | X | - | X | X |
| 4   | au-50102   | IMOS National Reference Station Darwin (Northern Australia) | 2011–present | X | X | - | X | X | X | X | X |
| 5   | au-50103   | IMOS National Reference Station Esperance (Southern Australia) | 2009–present | X | X | X | X | X | X | X | X |
| 6   | au-50104   | IMOS National Reference Station Kangaroo Island (Southern Australia) | 2008–present | X | X | - | X | X | X | X |
| 7   | au-50106   | IMOS National Reference Station Ningaloo (Western Australia) | 2010–present | X | X | - | X | X | X | X |
| 8   | au-50108   | IMOS National Reference Station Rottnest Island (Southwestern Australia) | 2009–present | X | X | - | X | X | X | X | X |
| 9   | il-10101   | Gulf of Elat Aqaba NMP Station A (Gulf of Elat – Gulf of Aqaba) | 2003–present | X | X | X | X | X | - | - | - |
| 10  | za-30202   | ABCTS Mossel Bay Monitoring Line (Agulhas Bank) | 1988–present | - | - | - | - | - | - | - | X |
7.6 References


