Aragonite saturation state in a monsoonal upwelling system off Java, Indonesia

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A B S T R A C T

Aragonite saturation state (Ωarag) and its influence from upwelling along the southern coast of Java, Indonesia were examined using carbonate and hydrographic data collected from 22 September to 2 October 2013. Results showed that sea surface Ωarag was lower in the upwelling area (2.97–3.44) than in the nonupwelling area (3.45–3.57), with the lowest value in the eastern part of the study area. We used a two end-member mixing model to separate contributions on Ωarag from two processes associated with upwelling: physical transport vs. biological production. Results indicated that physical transport induced at least a Ωarag decrease of 0.8, whereas biological production caused Ωarag to increase by up to 0.6. Additionally, the influence on Ωarag of interannual upwelling variability modulated by the Indian Ocean Dipole (IOD), a unique climate phenomenon in the Indian Ocean, was roughly estimated. We argue that the effect of interannual upwelling variability modulated by IOD events was possibly larger than what was imposed on Ωarag by increasing atmospheric CO2 levels over the past decade.

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1. Introduction

While effectively alleviating the rise in atmospheric carbon dioxide (CO2) concentration and global climate change (Canadell et al., 2007; Le Quéré et al., 2014), the oceanic uptake of anthropogenic CO2 has been causing a decrease in pH, carbonate ion concentration, and carbonate mineral saturation states in the oceans, a process commonly referred to as “ocean acidification” (OA) (Caldeira and Wickett, 2003; Doney et al., 2009; Feely et al., 2004; Orr et al., 2005). OA not only leads to specialization shifts of elements in seawater (Gledhill et al., 2015), affecting their bioavailability to phytoplankton, but also impacts the ability of many marine calcifying organisms to form their calcium carbonate (CaCO3) shells and skeletons (Bednarsek et al., 2012; Doney et al., 2009).

Carbonate mineral saturation state (Ω) may better represent the effect of carbonate chemistry on biologically mediated CaCO3 precipitation, compared with pH and CO2 partial pressure or other CO2 species (Riebesell, 2004; Waldbusser et al., 2014). When Ω < 1.0, waters are undersaturated with respect to carbonate minerals, causing shells and skeletons made of CaCO3 to become susceptible to dissolution without protection from, e.g., organic coatings (Doney et al., 2009). Aragonite is usually the most abundant form of CaCO3 in shallow waters (Morse et al., 2007) and is about 50% more soluble than calcite at 25 °C (Mucci, 1983). Also aragonite is essential for many bivalve larvae (e.g., Waldbusser et al., 2014). As a result, the saturation state of aragonite (Ωarag) has received more attention, particularly in the coastal ocean (Feely et al., 2010; Harris et al., 2013; Waldbusser and Salisbury, 2014; Wallace et al., 2014; Zhai et al., 2014). In addition, the controlling mechanisms of Ωarag changes in the coastal ocean are more complex than in the open ocean due to the superposition of other processes such as eutrophication and upwelling (Cai et al., 2011; Duarte et al., 2013; Feely et al., 2008), identifying the major controlling processes of Ωarag in key coastal ecosystems is important for predicting Ωarag changes in the future with increasing atmospheric CO2 levels.

Upwelling systems are usually at a higher risk of acidification as they experience a natural decrease in carbonate ion concentration and Ωarag (Feely et al., 2008; Gruber et al., 2012; Harris et al., 2013; Lachkar, 2014; Takahashi et al., 2014), in addition to the OA effects exerted by increasing atmospheric CO2 levels. Upwelling brings cold and salty subsurface waters rich in CO2 to or near the surface, lowering carbonate ion concentration and thus Ωarag (Feely et al., 2008) (physical transport effect). At the same time, it brings waters enriched in nutrients to the surface, favoring the growth of phytoplankton (Hales et al., 2005); subsequently, biological production absorbs CO2 and increases carbonate ion concentrations and thus Ωarag (biological effect). Therefore, there are two
processes associated with upwelling that have opposite effects on the CO₂ system (Borges and Frankignoulle, 2002). The net effect of these two processes on CO₂ varies in different upwelling systems, depending on the extent of exhaustion of upwelled nutrients by biological production (Borges, 2005; Borges et al., 2005; Cao et al., 2014; Chen et al., 2004). Off the California coast (Feely et al., 2008), the transport effect of upwelling was predominant; in contrast off the Galician and Oregon coast (Borges and Frankignoulle, 2002; Hales et al., 2005), the biological effect of upwelling was found to be predominant.

Eastern boundary upwelling systems play an important role in marine ecosystems and commercial fisheries (Chavez and Messié, 2009). Covering less than 1% of the global ocean’s surface area, they account for about 11% of the global new production (Chavez and Toggweiler, 1995) and up to 20% of the global fish catch (Pauly and Christensen, 1995). The marine ecosystems and related fisheries economics in these areas could be greatly influenced by OA (Cooley and Doney, 2009). Thus, examining carbonate mineral saturation state and its major controlling processes in eastern boundary upwelling systems is crucial for understanding the impacts of OA on marine biogeochemistry, organisms and ecosystems as well as the fisheries economics (Lachkar and Gruber, 2012).

The South Java upwelling system, an eastern boundary upwelling system in the tropics, is unique in that it is driven by typical Indian Ocean monsoon systems (monsoon upwelling system). The upwelling is only present during the southeast monsoon from June to October (Susanto et al., 2001), unlike the relatively stable upwelling systems driven by trade winds in the Pacific and the Atlantic. It is also linked to climate phenomena such as El Niño–Southern Oscillation (ENSO) and the Indian Ocean Dipole (IOD) (Iskandar et al., 2009; Susanto et al., 2001). The South Java upwelling system hosts high levels of biological production, especially during the typical positive IOD (El Niño) years (Iskandar et al., 2009; Sachoemar and Yanagi, 2001; Susanto et al., 2006), and it acts as one of the most productive areas for oceanic migratory fish such as skipjack and tuna (Sartimbul et al., 2010). In addition, coral reefs, which are very sensitive to OA (Erez et al., 2011), are widely distributed along the Java coast (Allen and Adrim, 2013).

Influences of the South Java upwelling on nutrients, dissolved oxygen, chlorophyll a (Chl a), and fishery productivity have been explored before using satellite data such as sea surface temperature and Chl a (Iskandar et al., 2009; Ningsih et al., 2013; Sartimbul et al., 2010; Susanto et al., 2001, 2006), and sporadic field observations conducted during early periods (Sachoemar and Yanagi, 2001; Wyrkti, 1962). However, information on Ωₘₐᵦ in the South Java upwelling region and the influence of upwelling processes on it is not available to date.

Based on field observations during September–October 2013, we report the Ωₘₐᵦ distribution for the first time along the southern coast of Java, Indonesia during the decay of an upwelling event. We quantify how the two processes—physical transport vs. biological production—influence Ωₘₐᵦ using a two end-member mixing model. In addition, we explore the influence on Ωₘₐᵦ from interannual upwelling variability modulated by IOD events.

2. Study site and methods

2.1. Study site

The study site is located in the southeastern Indian Ocean, off the southern coast of Java Island, extending from 7°S to 10°S along 106°E and 114°E (Fig. 1). It is strongly influenced by seasonal monsoons, with southeast monsoons during June–October and northwest monsoons during December–February (Susanto et al., 2001). The southeasterly wind generates annual upwelling, which is eventually terminated because of the reversal of winds, associated with the onset of the northwest monsoon and impingement of the Indian Ocean equatorial Kelvin wave (Susanto et al., 2001; Wijffels et al., 1996; Wyrkti, 1962). The upwelling intensifies during positive IOD (El Niño) years and weakens during negative IOD (La Niña) years (Iskandar et al., 2009; Ningsih et al., 2013; Susanto et al., 2001). This study site is also affected by the South Java Current (SJC) and the Indonesian Throughflow (ITF) (Wijffels et al., 1996).

Our investigation was conducted from 22 September to 2 October 2013 onboard the R/V BI–8 from the Indonesian Institute of Sciences (Indonesian: Lembaga Ilmu Pengetahuan Indonesia or LIPI), Indonesia. In total, there were 16 sampling stations (three transects included, Fig. 1). At each station, water samples with depths of 3 m, 10 m, 30 m, 50 m, 75 m, 100 m 150 m, 200 m, and 300 m were collected using a rosette sampler with 10-L Niskin bottles mounted on a SeaBird 911 Conductivity Temperature Depth (CTD) sensor (SeaBird Inc., Bellevue, WA, USA), which measured temperature and salinity in the water column.

2.2. Determination of dissolved inorganic carbon, total alkalinity, and Ωₘₐᵦ

Water samples for dissolved inorganic carbon (DIC) and total alkalinity (TA) were collected from Niskin bottles and were stored in borosilicate glass flasks after adding saturated HgCl₂ solution (0.02% of the total volume) as a preservative. DIC was analyzed with an infrared CO₂ analyzer by acidifying seawater with phosphoric acid (H₃PO₄) and quantifying the release of CO₂; TA was measured by open cell Gran titration (Huang et al., 2012; Wang and Cai, 2004). DIC and TA measurements had a precision of 0.1% (Huang et al., 2012; Wang and Cai, 2004). During the analysis, Certified Reference Materials (CRMs) from Scripps Institution of Oceanography were used to check the accuracy of the system (~4 μmol kg⁻¹ for DIC and TA, Ωₘₐᵦ at in situ temperature (Ωₘₐᵦ(Ωₘₐᵦ sant) and a constant temperature of 25 °C (Ωₘₐᵦ(Ωₘₐᵦ)) were calculated with DIC and TA data as well as phosphate and silicate data (see Fig. S1 in the Supplementary material) using the CO₂sys program (Lewis and Wallace, 1998) and the CO₂ system coefficients of Mehrbach et al. (1973) reffit by Dickson and Millero (1987).
2.3. Chl a measurement

Chl $a$ concentrations were measured fluorometrically using a F3010 fluorescence spectrophotometer (Hitachi Co., Tokyo, Japan) as in Zhang et al. (2010). Briefly, seawater samples were filtered under low vacuum (<0.05 MPa) through 47 mm Whatman GF/F filters and then Chl $a$ was extracted by soaking the filters in 10 ml 90% acetone for 24 h.

3. Results

3.1. Hydrographic data

Sea surface temperature (SST) was relatively low (25.97–26.76 °C) along Transects B and C as well as in the nearshore area of Transect A (stations A0–A2), and was high (27.43–27.85 °C) in the offshore area of Transect A (Fig. 2). This pattern was probably associated with relatively strong upwelling in the eastern part of the study area (Kuswardani and Qiao, 2014). In contrast, sea surface salinity (SSS) was relatively high (34.14–34.27) along Transects B and C as well as in the nearshore area of Transect A, and low (33.71–33.95) in the offshore area of Transect A (Fig. 2).

The water column was highly stratified, with temperature decreasing (salinity and sigma-theta increasing) from the surface to the deep layer (Fig. 3). Shoreward ascending of isolines of temperature, salinity, and sigma-theta was observed from Transect A. Waters at a 50 m depth can go all the way and reach the surface (Fig. 3). This is a typical hydrologic phenomenon occurring in this region during this period, i.e., South Java upwelling, which has been reported in a number of studies (Kuswardani and Qiao, 2014; Ningsih et al., 2013; Sachoemar and Yanagi, 2001; Susanto et al., 2001; Wyrkti, 1962). Although it seems that the upwelling feature along Transect B was not as clear as that along Transect A (Fig. 3), Transect B was in fact more affected by upwelling than Transect A (Kuswardani and Qiao, 2014; Sachoemar and Yanagi, 2001). This can be seen from the similar hydrographic properties between Transect B and the upwelling area of Transect A (Fig. 4) as well as the surface and vertical distributions of temperature, salinity, and sigma-theta (Figs. 2 and 3). The inconspicuous upwelling visually along Transect B was probably related to station locations, which were all located within the upwelling area, making along-transect comparison harder. Thus, hereafter our discussion will be focused on Transect A, which was clearly influenced by upwelling in the nearshore area and not in the offshore area (Figs. 2 and 3).

3.2. Chlorophyll $a$

Sea surface Chl $a$ showed a similar distribution pattern to SSS, with relatively high values of 0.24–1.99 mg m$^{-3}$ along Transects B and C as well as in the nearshore area of Transect A (upwelling area), and low values of ~0.15 mg m$^{-3}$ in the offshore area of Transect A (nonupwelling area, Fig. 2). Vertical maxima of Chl $a$ levels greater than 0.4 mg m$^{-3}$ occurred in the upwelling affected area (Fig. 3). Furthermore, Chl $a$ concentrations in the upper 75 m of Transect B were higher than those along Transect A, also implying relatively stronger upwelling along Transect B than Transect A (Kuswardani and Qiao, 2014).

3.3. DIC, TA, and aragonite saturation state

Vertical distributions of DIC and TA along Transects A and B were shown in Fig. 5. Their distribution patterns were generally similar to salinity (Fig. 3), with high DIC and TA waters at the subsurface layer upwelling to the surface at Transect A. In addition, both DIC (1884–1973 μmol kg$^{-1}$) and TA (2189–2240 μmol kg$^{-1}$) values at the surface were slightly lower than those in the neighboring open ocean (Indian Ocean, ~1950 μmol kg$^{-1}$ for DIC and ~2250 μmol kg$^{-1}$ for TA) (Bates et al., 2006). These low values at our coastal site were possibly related to input of low-salinity water carried by the SJC originating from runoff in the Java Sea or along the west coast of Sumatra (Wijffels et al., 1996).

Calculated $\Omega_{arag}$ at sea surface showed a similar spatial distribution to SST, with relatively low values of 2.97–3.44 along Transects B and C as well as in the nearshore area of Transect A (upwelling area), and high values of 3.45–3.57 in the offshore area of Transect A (nonupwelling area) (Fig. 2). In the eastern part of the study area (station C1), we found the lowest surface $\Omega_{arag}$ of 2.97, approaching the threshold (3.0) that indicates the lower limit of conditions in which coral reefs can recover after damage (Manzello et al., 2014). Vertically, $\Omega_{arag}$ was oversaturated in the upper 300 m of the water column, though it decreased from the surface to the deep layer (Fig. 5). Low $\Omega_{arag}$ waters in the subsurface layer were observed to upwell towards the surface at Transect A (Fig. 5). For example, the $\Omega_{arag}$ contour of 3.4 near the 50 m depth reached all the way to the surface in the nearshore area.

![Fig. 2. In situ SST (a), satellite SST on 23 September 2013 (b), in situ Chl $a$ (c), satellite Chl $a$ on 23 September 2013 (d), SSS (e), and $\Omega_{arag}$ in situ (f). Satellite data are from Aqua MODIS (http://coastwatch.pfeg.noaa.gov/erddap/griddap/). Dashed lines in b and d show the locations of Transects A and B.](image-url)
area, indicating the influence of upwelling. Furthermore, the surface $\Omega_{\text{arag}}$ along Transect B was almost identical to that in the upwelling area of Transect A (Fig. 5).

4. Discussion

4.1. Influence of upwelling events on $\Omega_{\text{arag}}$

We used Transect A as an example to examine the effects of the two processes associated with upwelling (physical transport vs. biological production) on $\Omega_{\text{arag}}$. Figs. 2 and 3 show that the nearshore area of Transect A was clearly influenced by upwelling-driven physical transport, which can bring subsurface waters with high CO$_2$ and nutrients to or near the surface (Feely et al., 2008; Sachoemar and Yanagi, 2001), and by subsequent strong biological production. This area (stations A0–A2) had lower temperature, higher salinity, and higher Chl $a$ than the nonupwelling area (Figs. 2 and 3). Despite the strong biological production in the upwelling area, $\Omega_{\text{arag}}$ was lower in the upwelling area than in the surrounding nonupwelling area (Figs. 2 and 3), indicating the dominant role of physical transport.

Mixing in the upper 150 m of the water column can be approximated as a two end-member mixing process, as indicated by the temperature–salinity plot (Fig. 4) and the conservative behavior of TA (Fig. 6). This allowed us to use a two end-member mixing model to separate the contribution of physical transport and biological production associated with upwelling (Benitez-Nelson et al., 2007; Cao et al., 2014; Chen et al., 2008). We used the 10 m depth water at station A6 and the 150 m depth water at station A5 as the surface and deep layer end-members, respectively. The justifications are: (1) station A6 was farthest from Java Island among the investigated stations and was not affected by upwelling (Figs. 1–3); (2) the 10 m depth water was not disturbed by evaporation and precipitation as directly or strongly as the surface water; (3) subsurface waters at a water depth of 150 m can ascend to the surface (Sachoemar and Yanagi, 2001; Wyrtki, 1962); and (4) the biological production at these two end-members was relatively low as indicated by Chl $a$ level (Fig. 3).

Considering that $\Omega_{\text{arag}}$ was not linear, we first calculated DIC (DIC$_{\text{mix}}$) and TA (TA$_{\text{mix}}$) under conservative mixing conditions based on the two end-member mixing model Eq. (1a) and (1b) and then calculated the conservative mixing value of $\Omega_{\text{arag}}$ ($\Omega_{\text{mix}}$) using these DIC and TA values via the CO$_2$SYS program Eq. (2). Note that the calculation of $\Omega_{\text{arag}}$ performed at a constant temperature of 25 °C and did not take account of the temperature effect. In fact, the temperature effect on $\Omega_{\text{arag}}$ was very minor (Kuchinke et al., 2014), which can be seen from almost the same vertical distributions of the aragonite saturation state at in situ temperatures and a constant temperature of 25 °C (Fig. 5, also see the discussion about temperature effect in the supplementary material).

\[
\text{DIC}_{\text{mix}} = \text{DIC}_{10} + (S - S_{10}) \times (\text{DIC}_{150} - \text{DIC}_{10})/(S_{150} - S_{10}) \\
\text{TA}_{\text{mix}} = \text{TA}_{10} + (S - S_{10}) \times (\text{TA}_{150} - \text{TA}_{10})/(S_{150} - S_{10}) \\
\Omega_{\text{mix}} = f(\text{DIC}_{\text{mix}}, \text{TA}_{\text{mix}}, S, 25) 
\]

where DIC$_{10}$ (DIC$_{150}$), TA$_{10}$ (TA$_{150}$) and S$_{10}$ (S$_{150}$) are the DIC, TA, and salinity at the surface (deep) layer end-member, respectively; $S$ is the salinity at the given point between the surface and the 150 m depth; and $f$ denotes $\Omega_{\text{arag}}$, a function of DIC, TA, salinity, and temperature.

Assuming that pre-upwelling conditions in the upwelling area were the same as in the nonupwelling area such as at station A6, we were able to conclude that upwelling-driven physical transport resulted in a salinity increase of at least 0.4, a DIC increase of 110 μmol kg$^{-1}$, and a $\Omega_{\text{arag}}$ decrease of 0.8 (Fig. 6). We also found that when Chl $a$ was less than 0.2 mg m$^{-3}$, most of the DIC and $\Omega_{\text{arag}}$ changes can be explained by salinity (Fig. 6). In contrast, when Chl $a$ was greater than 0.2 mg m$^{-3}$, DIC was in deficit (below the theoretic mixing line) and $\Omega_{\text{arag}}$ was above the theoretic mixing line, indicating that biological uptake of...
CO₂ was becoming more important. In the upwelling-affected area where there was very strong biological production (Chl a concentration was greater than 0.4 mg m⁻³), DIC decreased by up to 70 µmol kg⁻¹, and Ωₘarag increased by up to 0.6 units due to biological CO₂ uptake (Fig. 6). Overall, the effect of physical transport was greater than that of biological production in this region, with a net effect of increasing DIC and decreasing Ωₘarag.

The separation of physical transport from biological production provides a quantitative view of how different aspects of upwelling affect Ωₘarag and can help better understand the OA mechanism in the upwelling system. However, one needs to be aware of the uncertainties related to this method. One uncertainty is from the assumption that pre-upwelling conditions in the upwelling area were the same as in the nonupwelling area (Benitez-Nelson et al., 2007; Chen et al., 2008). This will result in underestimation of physical transport, considering that the upwelling-affected area was much closer to the shore than the nonupwelling area. In addition, terrestrial input may also contribute to lowering Ωₘarag (Jiang et al., 2010). However, the higher salinity in the upwelling area compared with the nonupwelling area (Fig. 2) indicated that terrestrial input in the upwelling area was minor.

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**Fig. 5.** Profiles of DIC, TA, aragonite saturation state at in-situ temperature (Ωₘarag@in situ) and 25 °C (Ωₘarag@25) along Transects A and B.

**Fig. 6.** Scatterplots of TA vs. salinity (a), DIC vs. salinity (b), and Ωₘarag@25 vs. salinity (c) in the upper 200 m of the water column along Transect A. In (a), the dashed line is the linear regression line; in (b) and (c), the theoretical mixing line (dashed line) between two end-members and influences of physical transport of upwelling and biological production are shown. Details can be found in text. Also Chl a concentration (b and c) is shown with the color bar. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)
4.2. Impact of IOD events on upwelling and Ωarag

Climate phenomena such as ENSO and IOD could alter the strength of the South Java upwelling (Iskandar et al., 2009; Rixen et al., 2006; Sartimbul et al., 2010; Susanto et al., 2001) and consequently affect Ωarag. Considering that the IOD is a unique feature in the Indian Ocean basin (Saji et al., 1999; Sun et al., 2015), characterized by the dipole mode index (DMI, http://www.jamstec.go.jp/frsgc/research/d1/iod/iod/dipole_mode_index.html), we explore its influences on the South Java upwelling and hence on Ωarag. We chose the SST anomaly (SSTA) in August (when the upwelling strength is greatest) as an indicator of the upwelling strength. We first calculated the average SST in August of each year from 2002–2013 for the study area (7.81°S–9.86°S; 106.76–113.30°E) and surface SST from SST, Aqua MODIS, NPP, Indonesia, Daytime, http://coastwatch.pfeg.noaa.gov/erddap/griddap/erdMPsstdmday.html, then took an average for the data over twelve years, and finally obtained the SSTA. A negative SSTA denotes intensified upwelling and a positive SSTA weakened upwelling.

From Fig. 7, we see that the IOD events had important influences on SSTA, that is, on the upwelling strength. For example, in 2006 a SSTA of positive SSTA weakened upwelling. The interannual variability in upwelling strength may cause large variability in Ωarag.

Although the direct temperature effect on Ωarag was minor (Fig. S2 in the Supplementary material), there was a strong relationship between temperature and Ωarag, since low temperatures in deep water usually correspond to high salinity and low Ωarag (Figs. 3–5). Thus, we can use the slope between Ωarag and temperature (Fig. 7) to make a rough estimate of the influence of upwelling variability on Ωarag. The slope varied between 0.107 °C⁻¹ for subsurface water with temperatures less than 25 °C and 0.196 °C⁻¹ for surface water with temperatures greater than 25 °C, which can be regarded as lower and upper limits of slopes. Here we used the low slope value of 0.107 °C⁻¹ to make a conservative estimate of the influence of IOD events on the interannual variability in Ωarag. In 2006 during a strong positive IOD year, SST was −0.91 °C and surface Ωarag was expected to be at least 0.10 (0.91 + 0.107) lower than normal. In 2010 during a strong negative IOD year, SST was 1.91 °C and surface Ωarag would be 0.20 (1.91 + 0.107) higher than normal. In 2013 during a weak positive IOD year (Fig. 7), SST was 0.76 °C and surface Ωarag would be 0.08 (0.76 + 0.107) higher than normal. Lenton et al. (2015) also found large interannual variability in Ωarag (>0.05) in the far eastern Tasman Sea. A more recent study showed that the mean rate of Ωarag decrease in the eastern equatorial Indian Ocean during 1962–2012 was only 0.01 year⁻¹ (Xue et al., 2014). Thus we argue that over the past decade the impact of IOD events on Ωarag (interannual variability) could be larger than that from anthropogenic CO₂ intrusion caused OA. Furthermore, this impact will become more important in the future, considering that the frequency of extreme IOD events due to global warming is increasing in the years to come (Cai et al., 2014).

The Ωarag–temperature relationship should be used with caution to estimate the effects of IOD on Ωarag, especially during a positive IOD year. Satellite images showed that during a positive IOD year there was stronger upwelling, which could induce higher Chl a concentrations (Ningsih et al., 2013). This implies that the effect of physical transport by strong upwelling will be reduced to some extent by enhanced biological activity, which will cause uncertainties for estimating Ωarag variations via the Ωarag–temperature relationship. However, we can use this relationship at least to estimate the influence of IOD on Ωarag, when the strength of biological activity was not stronger than during our cruise. Even when there was very strong upwelling, we can use the Ωarag–temperature relationship to estimate the influence of IOD on Ωarag for the period when anomalous upwelling occurred but strong biological activity did not yet develop, since phytoplankton needs time (1–2 weeks) before thriving. Recently Evans et al. (2015) demonstrated that anomalous upwelling resulted in negligible biological responses, which are greatest only after intense upwelling ceases. Overall, here we made a rough estimate of the influence of IOD events on Ωarag in a region with limited observations, and more efforts based on in situ observations should be made in the future to examine the actual impact of IOD events.

5. Summary and conclusions

Sea surface Ωarag along the southern coast of Java, Indonesia was affected by upwelling, with lower values in the upwelling area than in the nonupwelling area. Contributions on Ωarag from two processes associated with upwelling (physical transport vs. biological production) were separated by a two end-member mixing model, which can help better understand the OA mechanism in the upwelling system. During the study period, physical transport of upwelling resulted in a Ωarag decrease of 0.8 units, whereas biological production in the upwelling area caused Ωarag to increase by up to 0.6 units. This study also roughly estimated the impact of IOD events on Ωarag and found that the interannual variability in Ωarag associated with IOD events was possibly larger than that due to increasing atmospheric CO₂ over the past decade. This impact will become more important in the future, given the increasing frequency of extreme IOD events due to greenhouse warming in the years to come. More efforts based on in situ observations should be made in the future to examine the actual impact of IOD events.

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Appendix A. Supplementary data

Supplementary data to this article can be found online at http://dx.doi.org/10.1016/j.jmarsci.2015.08.003.

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