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Climate variability and ocean production in the Leeuwin Current system off the west coast of Western Australia

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Abstract

The strength of the Leeuwin Current (LC) and its eddy field are both strong during the austral winter and weak during the austral summer on the annual time scale, and are strong during the La Niña years and weak during the El Niño years on the interannual time scale. As the LC is a warm current, the sea surface (evaporative) heat loss off the west coast of Western Australia (WA), as well as the upper ocean stratification (mixing) and the nutrient fluxes, is also closely linked to the strength of the current. In this paper, recent studies on the temporal and spatial variability of the biophysical properties in the oligotrophic marine environment off the west coast of WA are reviewed. By analysing recent satellite chlorophyll *a* data and shipboard survey results, possible mechanisms that could be important to the variability of the ocean production off the west coast of WA are identified as: meridional erosion of the seasonal thermocline; vertical motion of the nitrocline; horizontal and vertical nutrient advection related to the LC eddy activity; *in situ* nitrification; and benthic-pelagic coupling. Potential impacts of climate change on the ocean production off the west coast of WA are hypothesized.

Keywords: ENSO; chlorophyll *a*; nitrocline; climate change; eddies

Introduction

Ocean circulation off the west coast of Western Australia (WA) is dominated by the poleward-flowing Leeuwin Current (LC). The LC is a warm ocean current that originates off the North West Cape of WA (22°S), and flows southwards along the edge of the continental shelf before turning eastwards around Cape Leeuwin (34°22'S, 115°08'E) and continuing into the Great Australian Bight (Cresswell & Golding 1980; Figure 1). The meridional pressure gradient in the southeast Indian Ocean, set up by the Indonesian Throughflow (ITF) in the tropics and by latent heat fluxes (cooling) in the mid-latitude, accounts for the existence of the LC (Thompson 1984; Godfrey & Ridgway 1985).

The LC deepens the thermocline and nitrocline off the west coast of WA (Thompson 1984), therefore suppresses productivity on the continental shelf, causing the oligotrophic marine environment off the coast. On the other hand, the LC is responsible for the existence of coral reefs as far south as 29°S (Collins *et al*. 1991) and the presence of tropical species along the west and south coasts of WA (Maxwell & Cresswell 1981; Hutchins & Pearce 1994). There are relatively large invertebrate populations off the coast, *e.g.*, western rock lobster (Panulirus Cygnus), Australia's most valuable single species fishery (Pearce & Phillips 1988; Caputi *et al*. 1995).

The LC has the strongest eddy energy among the mid-latitude eastern boundary current systems (Feng *et al*. 2005). The interannual variations of the LC and its eddy field respond to the El Nino/Southern Oscillation (ENSO) and many of the fisheries recruitments off WA are also associated with ENSO induced interannual variability (Caputi *et al*. 1995). The LC eddy field has vital influences on the marine pelagic production off the west coast of WA (*e.g.*, Hanson *et al*. 2005a; Feng *et al*. 2007; Koslow *et al*. 2008).

Understanding how the variability of the LC affects the ecosystems off the coast is crucial for the long term management of marine resources off WA. By reviewing the recent studies on the biophysical properties of the LC system and by combining satellite and *in situ* observations, this paper derives some key mechanisms that may be responsible for the nutrient dynamics in the LC system on seasonal and longer time scales, which is a first step to unveiling the full impact of the LC on the ecosystems off the WA coast.

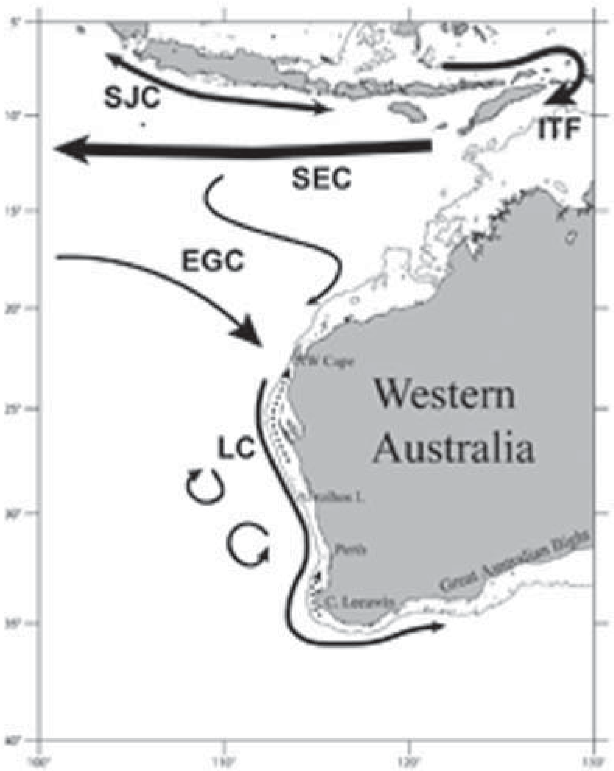

Figure 1. Regional currents in the East Indian Ocean and off the WA coast. ITF: Indonesian Throughflow; SJC: South Java Current; SEC: South Equatorial Current; EGC: East Gyral Current; LC: Leeuwin Current; NW Cape: Northwest Cape; Abrolhos I.: Abrolhos Islands; C. Leeuwin: Cape Leeuwin. The 200 m isobath of bottom bathymetry is shown as solid lines and the dashed lines denote the inshore wind-driven currents (Modified from Feng *et al*. 2003).

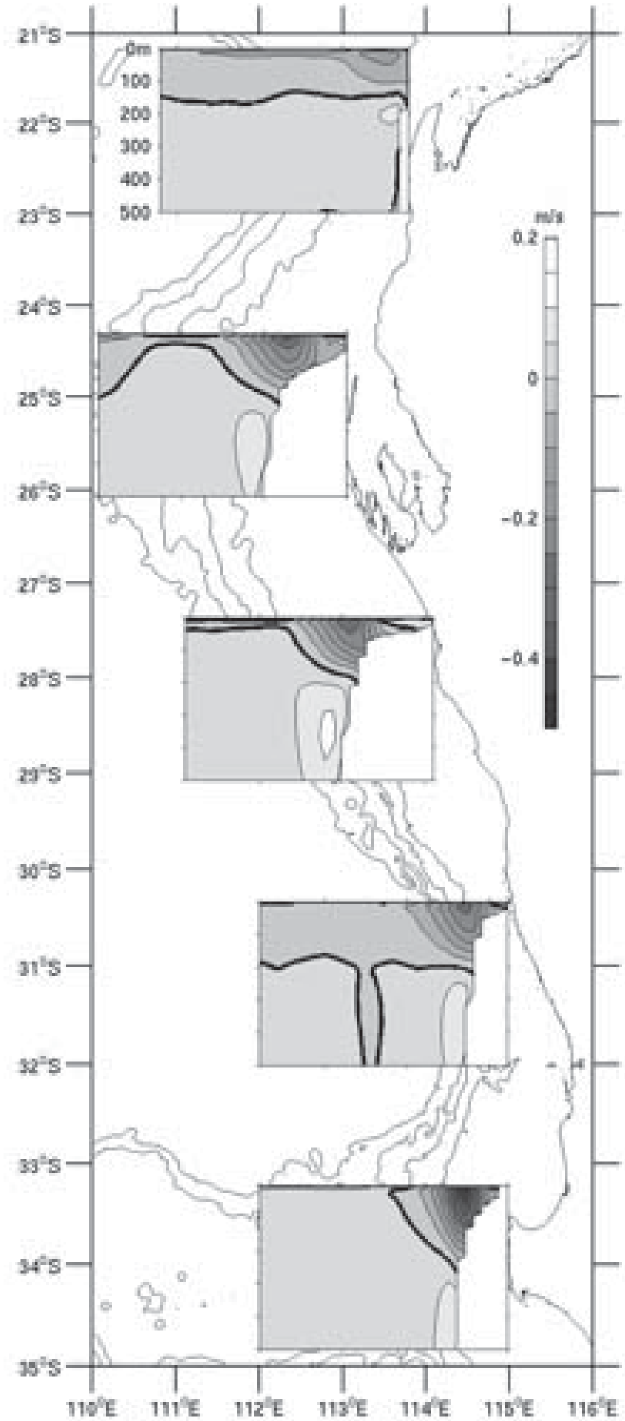

Figure 2. Mean meridional velocities at 22°, 25°, 28°, 31°, and 34°S in the upper ocean (0–500m) along the west coast of WA derived from the Ocean Forecast Australian Model (OFAM) 12-year simulation (Schiller *et al*. 2008).

Physical background

The latitudinal variations of the LC are illustrated with vertical sections of average meridional velocity from a high resolution, data-assimilating numerical simulation of the ocean system (Schiller *et al*. 2008). The LC at 22°S is a broad, shallow surface current in the upper 150m, carrying waters of the tropical origin southward (Figure 2). The LC builds up its strength along the shelf break off Shark Bay and Abrolhos Islands, due to the convergence of onshore flows. The LC becomes dispersed between Abrolhos and Perth, likely due to the strong eddy activity in this region (Feng *et al*. 2005), and further south, the LC regains its strength off the Capes near 34°S. The model description of the LC is quite consistent with earlier field observations (Smith *et al*. 1991).

The LC is strong during the austral winter and weak during the austral summer (Smith *et al*. 1991; Feng *et al*. 2003), mostly due to the seasonal variations of surface winds. From a monthly upper ocean thermal climatology off the west coast of WA, the annual mean geostrophic transport of the LC is about 3.4 Sv (10^6 m^3s^{-1}) referenced to 300 m, and the LC has its peak transport during May–July (Figure 3; Feng *et al.* 2003). The seasonal variation of the LC induces a 20 cm annual variation of the Fremantle sea level. During the austral summer, pulses of northward winds overcome the meridional pressure gradient and drive the episodic northward inshore currents (Cresswell *et al*. 1989). However, coastal upwelling is only significant along narrow strips of the shelf, such as off the Capes in the south (Gersbach *et al*. 1999) and north of Abrolhos (Woo *et al*. 2006).

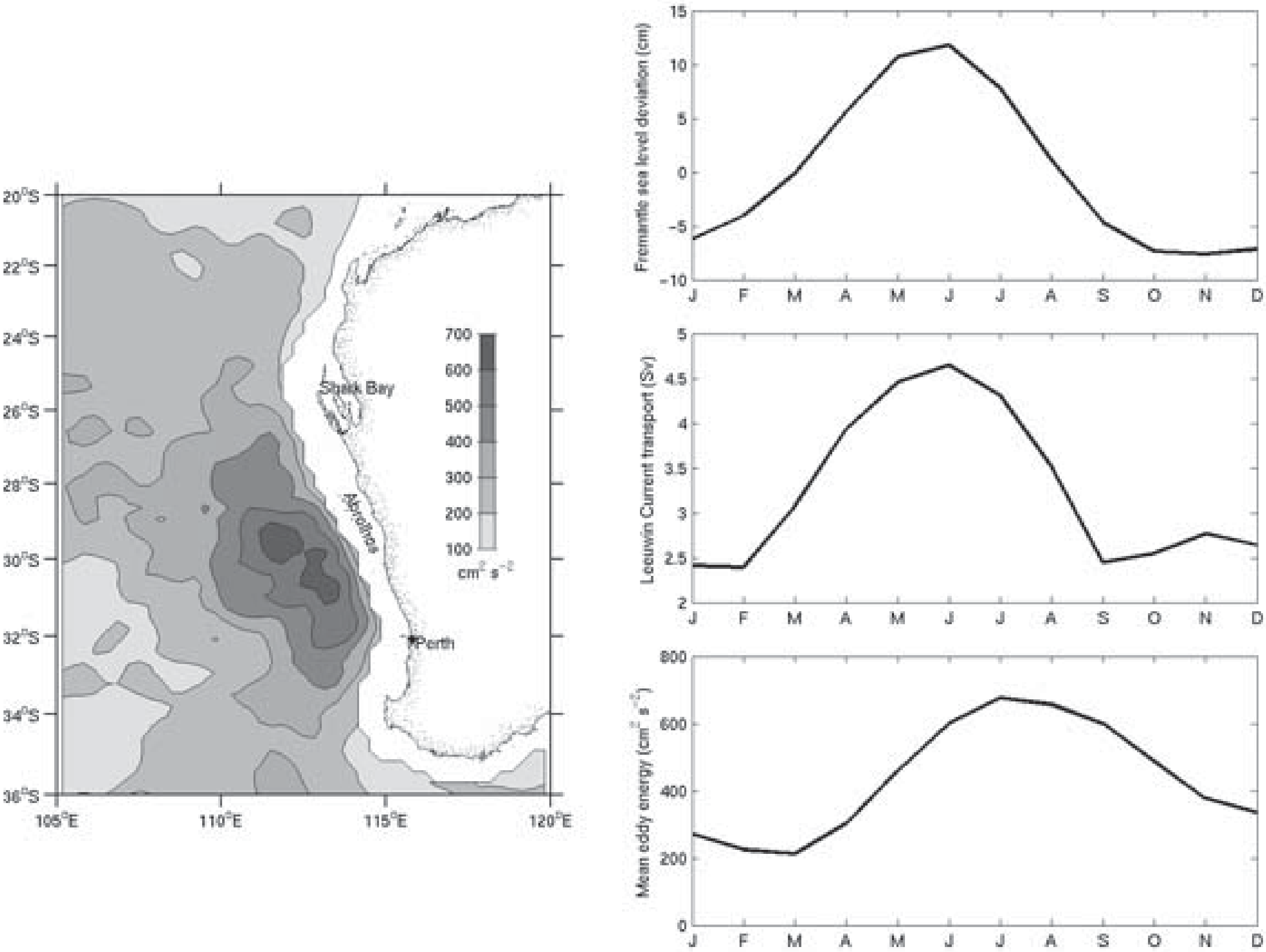

Figure 3. Long-term mean surface eddy kinetic energy derived from satellite altimeter data (left panel). The right panels (from top to bottom) are seasonal cycle of the Fremantle sea level, Leeuwin Current transport (Geostrophic & Ekman), and surface eddy kinetic energy between Abrolhos and Perth.

Different physical processes dominate the dynamic balances of the LC at different latitudes (Feng *et al*. 2005): north of Abrolhos, there is a balance between the southward pressure gradient and the northward wind stress; while south of Abrolhos, the pressure gradient is stronger and the alongshore wind stress is weaker so that the current is highly unstable and generates a strong mesoscale eddy field. South of Abrolhos, the Reynolds stress exerted by the eddy field is more important than the wind stress in balancing the southward pressure gradient (Figure 3; Feng *et al*. 2005). The LC eddies are important in the offshore transport of the momentum and heat from the LC (Fang & Morrow 2003; Feng *et al.* 2005; Domingues *et al*. 2006). On the seasonal cycle, the LC eddy field is strong during the austral winter and weak during the austral summer, such that the peak eddy energy occurs about 1 month later (July) compared to that of the peak LC transport (Figure 3). This would be expected since the eddy field draws its energy from the instability of the LC.

The ENSO related upper ocean variations propagate poleward as coastal Kelvin waves along the northwest to west WA coasts (Meyers 1996; Feng *et al.* 2003). The waves transmit high coastal sea levels (deep thermocline) and induce strong LC transports (4.2 Sv) during the La Niña years, and transmit low sea levels (shallow thermocline) and induce weak LC transports (3 Sv) during the El Niño years (Feng *et al.* 2003). A significant linear relationship between the Fremantle sea level and the volume transport of the LC across 32°S on the annual and interannual time scales can be derived. There is also a strong association between ENSO and the altimeter derived eddy energetics, $\frac{1}{2}(u^2+v^2)$, averaged between Abrolhos and Perth (Figure 4). Strong eddy energetics occurred during the La Niña years, *e.g*., 1996, 1999, and 2000, while weak LC eddy energetics were observed during the El Nino years, *e.g*., 1994, 1997, and 2002. During 1993–2006, the linear correlation between the annual mean Southern Oscillation Index (SOI) and the eddy energy is 0.94, demonstrating the strong sensitivity of the LC system to ENSO.

Another important feature of the physical environment in the LC is the strong surface heat loss along the southward flowing warm current. The heat loss is mostly due to the evaporative cooling (latent heat flux) when warm sea surface temperature in the LC meets the cold air temperature in the south and the frequent occurrence of winter storms originated from the Southern

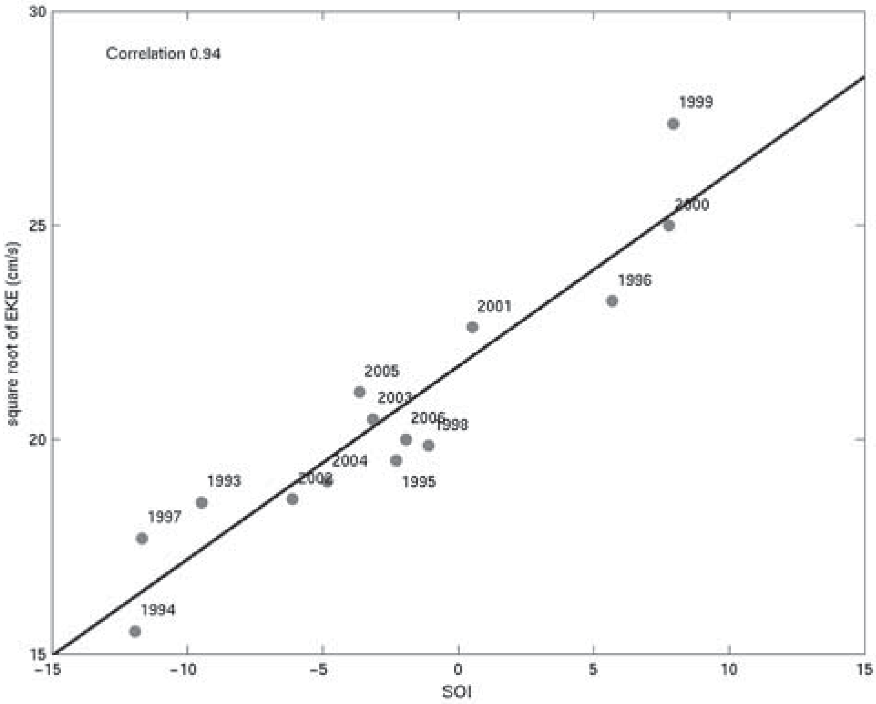

Figure 4. Relationship between annual mean Southern Oscillation Index (SOI) and the square root of the altimeter derived surface eddy kinetic energy (EKE) between Abrolhos and Perth ($R^2 = 0.88$).

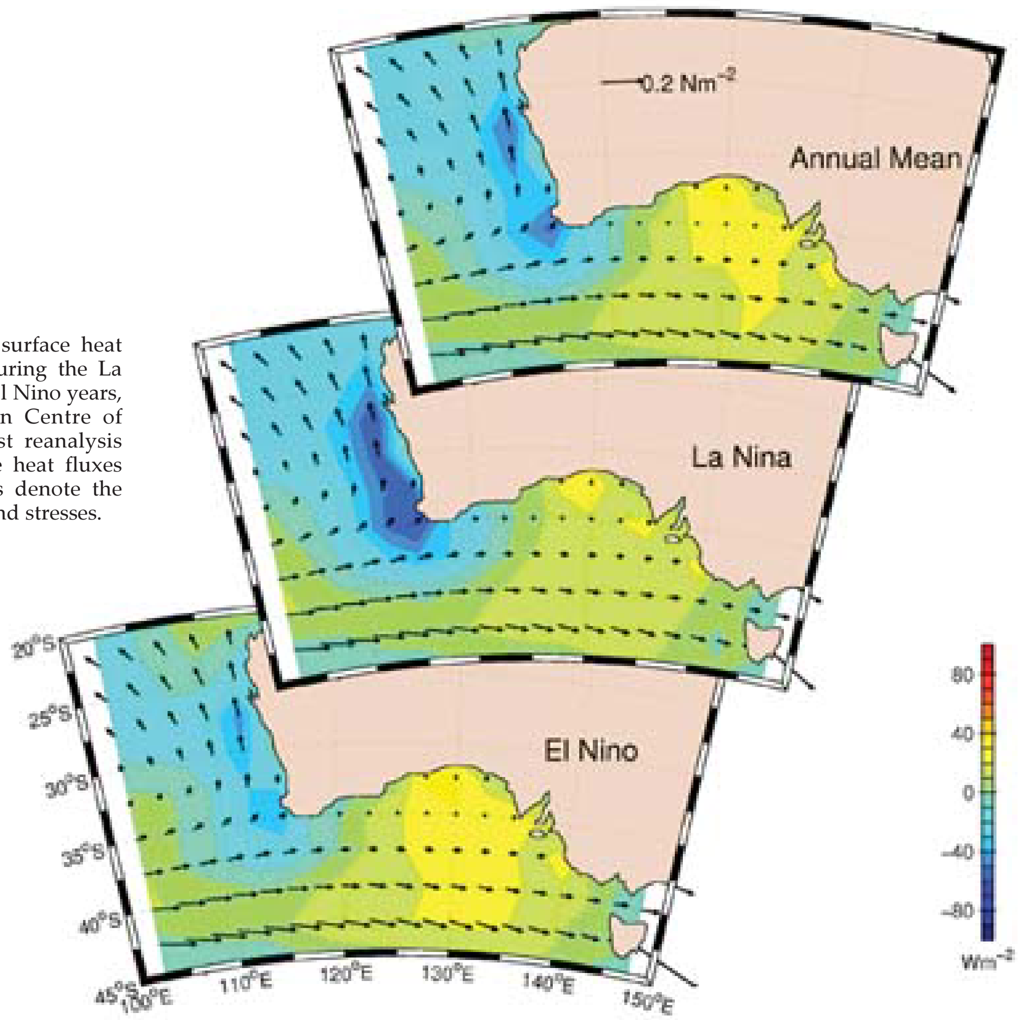

Figure 5. Average net sea surface heat fluxes, during 1957-2002, during the La Nina years, and during the El Nino years, derived from the European Centre of Mid-range Weather Forecast reanalysis product. Positive fluxes are heat fluxes into the ocean. The arrows denote the composites of the surface wind stresses.

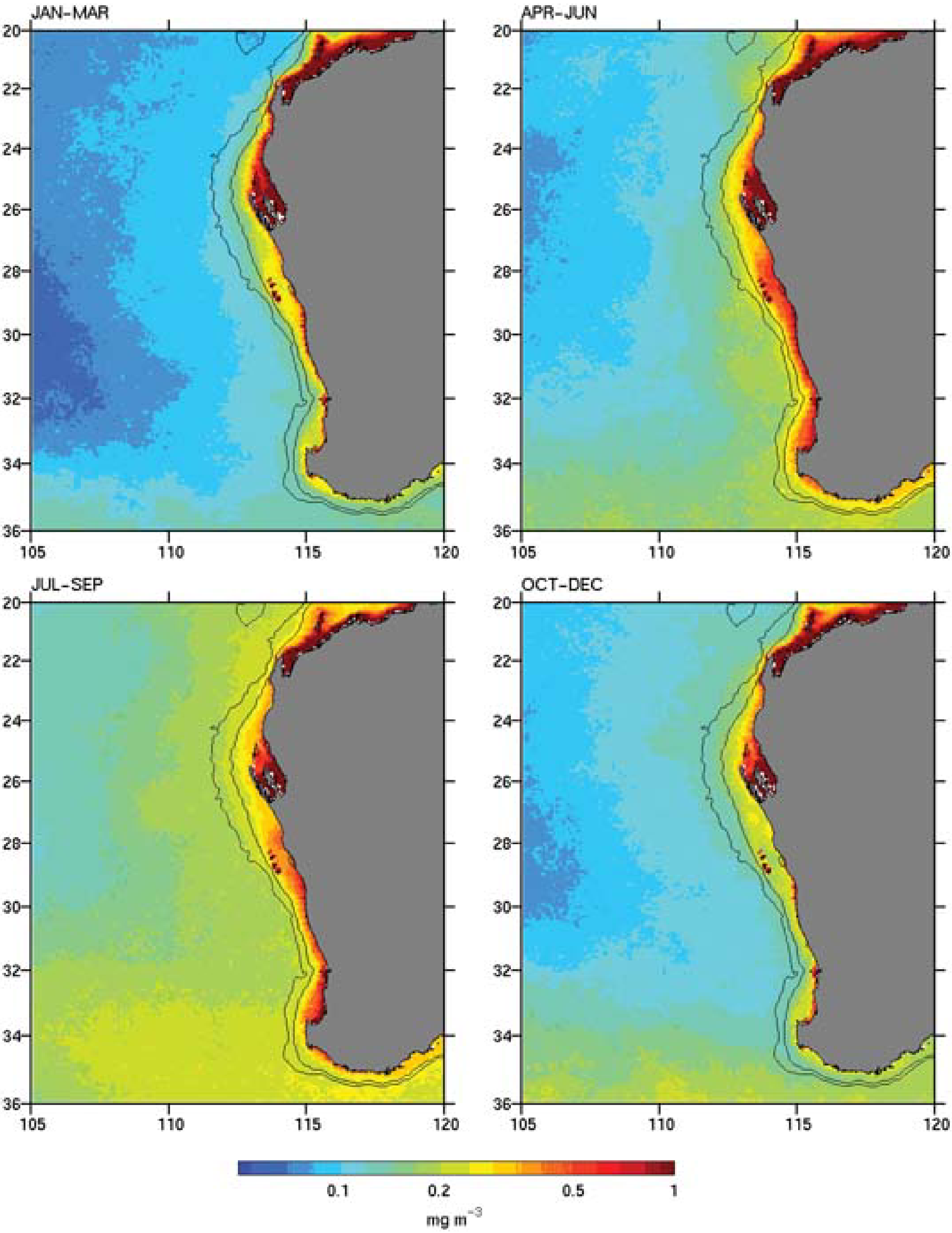

Figure 6. Seasonally-averaged sea surface chlorophyll *a* concentrations in the southeast Indian Ocean calculated from SeaWIFS monthly climatology data (1997–2007). The bottom bathymetry is denoted for the 50, 200, and 1000 m isobaths (adapted from Feng & Wild-Allen 2009).

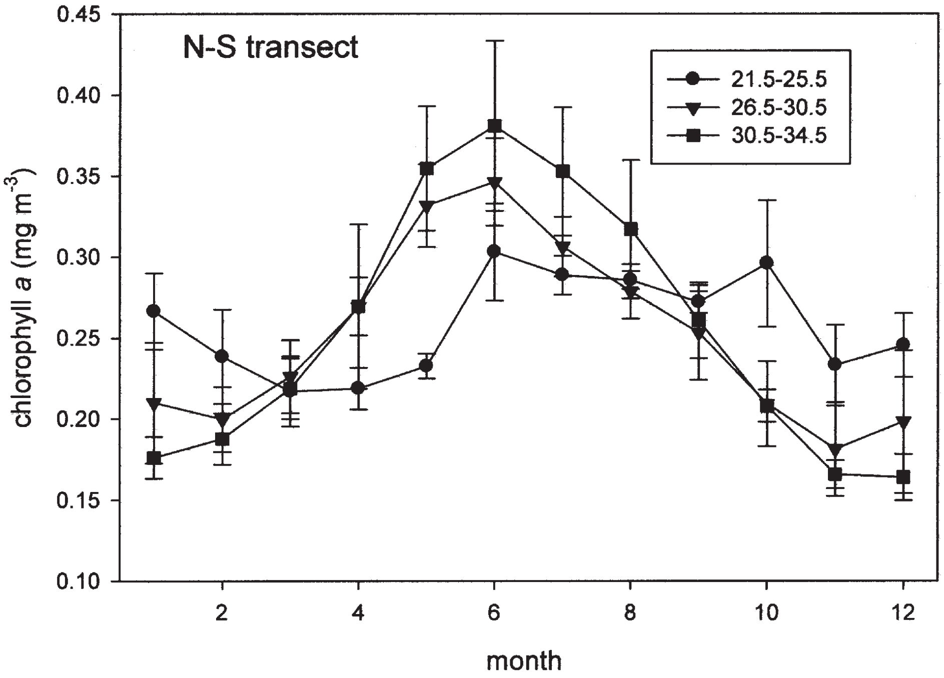

Figure 7. Monthly climatology of sea surface chlorophyll *a* concentrations averaged over three latitude bands on the shelf (approximately in the 100 and 1000 m depth range) off WA, derived from the 1997–2007 SeaWiFS ocean colour satellite data.

Ocean. The evaporative cooling is strong during austral winter, when the LC transport is strong. There are also consistent ENSO-related interannual variations in the surface heat loss – the heat loss is stronger during the La Nina years and weaker during the El Nino years (Figure 5). Surface cooling can induce strong vertical mixing which affects stratification in the water column and thus is important in nutrient cycling in the mixed layer of the LC (Greenwood *et al*. 2007).

Satellite measurement of chlorophyll *a* concentration

The launch of the Sea-viewing Wide Field-of-view Sensor (SeaWiFS) satellite in the late 1990s has provided measurements of upper ocean chlorophyll *a* concentration with 30% accuracy (Bailey & Werdell 2006). The SeaWiFS data have been used to quantify the seasonal variations of surface phytoplankton biomass in the LC system (Figure 6 Feng *et al*. 2007; Moore *et al*. 2007). The accuracy of the satellite data also allows us to explore the interannual variability.

In the oligotrophic marine environment off the west coast of WA, there are low upper ocean chlorophyll *a* concentrations compared to other eastern boundary current systems (with typical value of ~1 mg m^{-3}; http://oceancolor.gsfc.nasa.gov/cgi/climatologies.pl). There has been a recent discovery of late-autumn to early-winter chlorophyll *a* enhancement in the LC system south of Abrolhos from *in situ* observations (Koslow *et al*. 2008), which is also captured by the satellite measurements (Lourey *et al*. 2006; Feng *et al*. 2007; Moore *et al*. 2007). Averaged over the shelf (approximately in the 50–1000 m depth range), consistent late-autumn to early-winter (May–June) peaks of upper ocean chlorophyll *a* concentrations are observed south of Abrolhos, with peak values of about 0.4 mg m^{-3} (Figures 6 and 7). Offshore of the LC and within about 500 km off the coast, late-winter (July–September) phytoplankton enhancements are generally observed (Figure 6), with winter values of about 0.2–0.3 mg m^{-3}. Further offshore in the oligotrophic, subtropical open-ocean, the chlorophyll *a* concentration is very low and there is a winter peak of less than 0.2 mg m^{-3}. The September–October chlorophyll *a* peaks south of 32°S are likely due to the northward migration of the subtropical front, with peak values of about 0.3 mg m^{-3} (Figure 6). On the continental shelf north of Shark Bay, there is the summer (January) chlorophyll *a* peak (Figure 6), which is driven by upwelling favourable winds (Hanson *et al*. 2005a). The broad winter (June–October) peak in Figure 7 is likely due to that the box for average is relatively large so that winter peak in the offshore region may dominate the average. Most of the seasonal cycles are significant compared with the 30% accuracy range.

There are detectable interannual variations in the upper ocean chlorophyll *a* concentrations on the continental shelf off the west coast (Figure 8a). North of 28°S (Abrolhos), positive chlorophyll *a* concentration anomalies are observed during 2000–2001, and 2004–2006, while there are negative anomalies during 1999 and 2002. The variations south of 28°S tend to be correlated with Southern Oscillation Index, with positive anomalies in 1999–2000 and 2006 and negative anomalies in 1997 (not shown), and 2002–2004 (Figure 8b). These north-south differences suggest that more than one physical factor influences the interannual variations in chlorophyll *a* biomass and ocean production along the west coast. Note that averaged over the spatial domain of the LC the

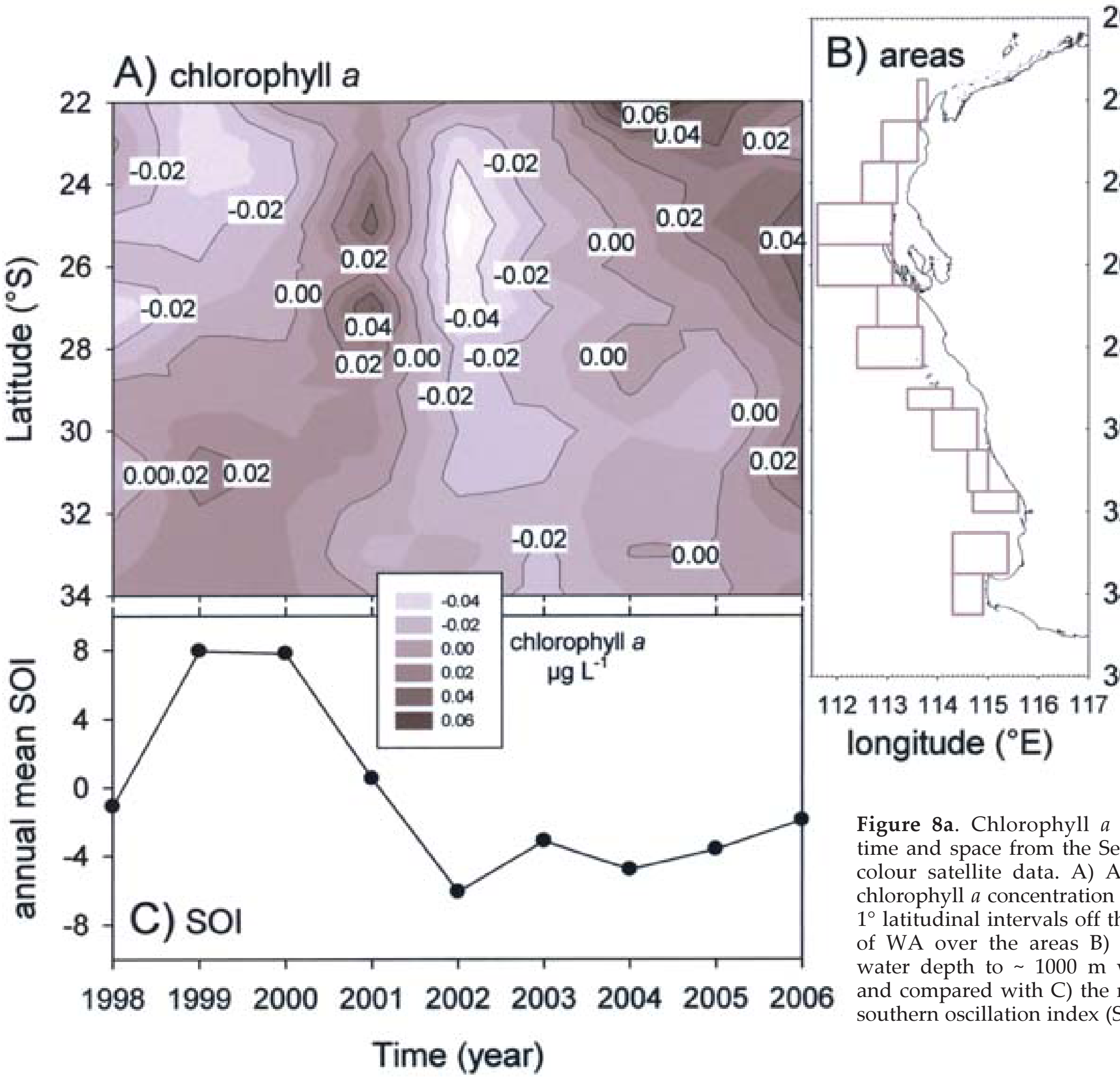

Figure 8a. Chlorophyll *a* variation in time and space from the SeaWiFs ocean colour satellite data. A) Annual mean chlorophyll *a* concentration anomalies at 1° latitudinal intervals off the west coast of WA over the areas B) from ~ 50m water depth to ~ 1000 m water depth, and compared with C) the mean annual southern oscillation index (SOI).

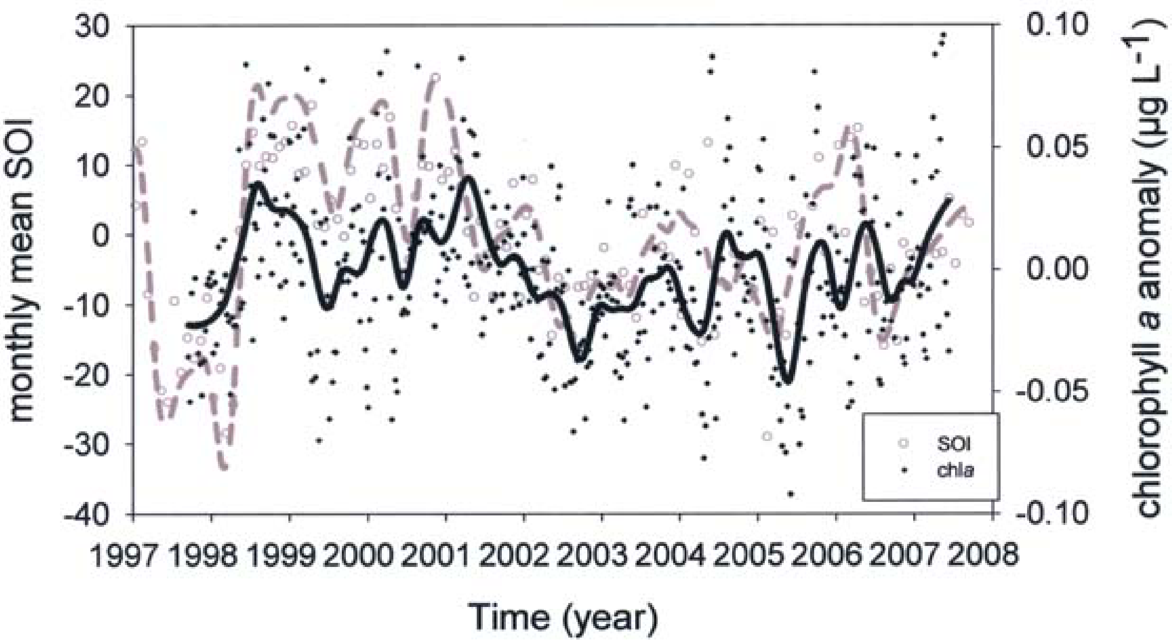

Figure 8b. Time series of 8-day chlorophyll *a* anomalies from SeaWiFs ocean colour satellite data over the area 29 to 31°S and 113.5 to 114.5 °E off the west coast of WA. The monthly chlorophyll *a* anomaly (solid line) could be approximated by a linear function of monthly SOI (dashed line); $r^2 = 0.135$, $P < 0.001$.

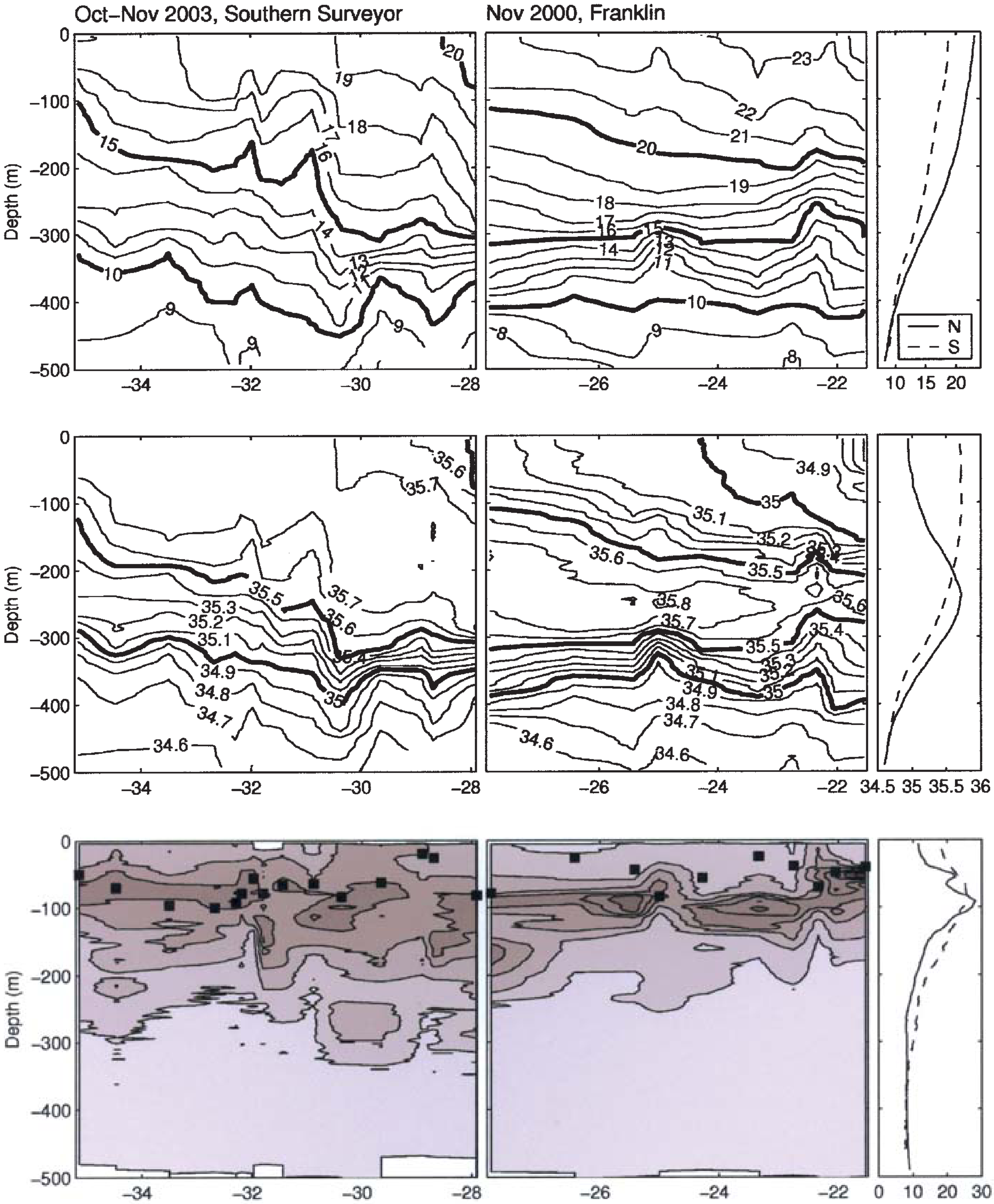

Figure 9a. (top panel) Temperature, (middle panel) salinity and (lower panel) fluorescence readings collected with CTD (conductivity-temperature-depth) sensors along the 500 isobath during October–November 2000 and 2003. The unit for temperature is °C and the darker shadings denote higher fluorescence readings. The solid squares in the fluorescence panels denote the CTD station and the mixed layer depth. The averaged profiles north (solid) and south (dashed) of 28°S are displayed to the right of each panel.

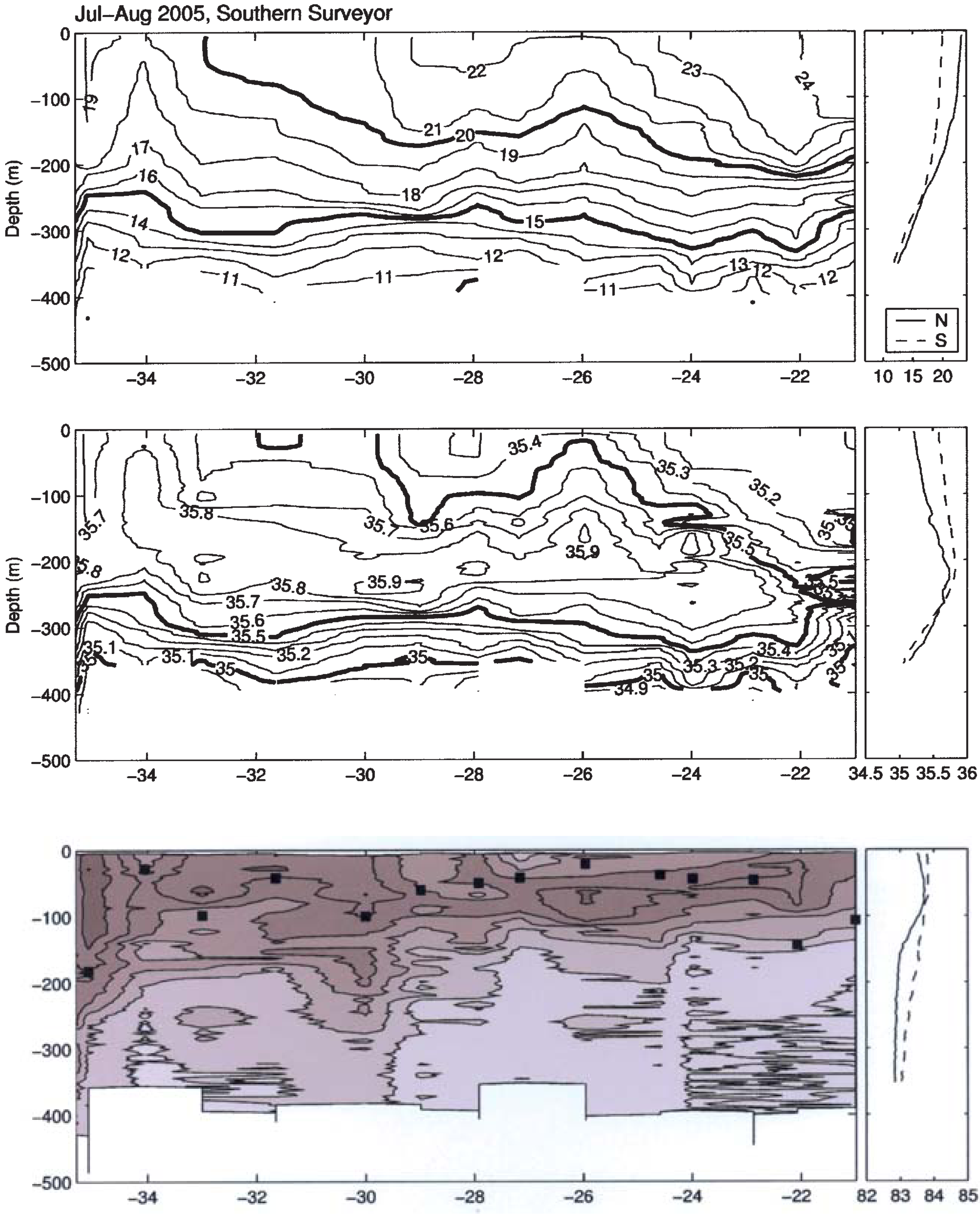

Figure 9b. same as Figure 9a except for the July-August 2005 Southern Surveyor cruise.

annual chlorophyll *a* anomalies is a small fraction (less than 30%) of the total standing stocks (Figure 8a), so that precautions should be taken in interpreting the interannual variations.

In situ observations

In recent years, a number multi-disciplinary research cruises have been carried in the LC system off the west coast of WA. In combination with satellite observations, the cruise data have been used to understand the nutrient dynamics off the coast, especially the coastal upwelling off the NW Cape and the role of mesoscale eddies (Morrow *et al*. 2003; Hanson *et al*. 2005a, 2005b; Waite *et al*. 2007; Thompson *et al.* 2007; Koslow *et al*. 2008). In this section, we use two north-south hydrographic sections along the continental shelf break during the two opposite seasons (spring/summer and autumn/winter) to highlight the north-south and seasonal variations of the biophysical properties along the west coast of WA.

From alongshore (500 m isobath) sections during the austral spring (a composite from the R/V *Franklin* cruise in November 2000 and the R/V *Southern Surveyor* cruise in late October–November 2003), there are general decreasing trend of temperature and increasing trend of salinity in the upper 200 m (the LC) from north to south (Figure 9a). Air-sea fluxes and exchanges with subtropical water masses modify the LC waters when the current flows southward. North of Abrolhos, both temperature and salinity stratifications contribute to produce a relatively strong and shallow thermocline (<100 m). Salinity maximum (>35.7 psu) subtropical surface waters formed in the mid-latitude due to excessive evaporation are subducted to about 200 m depth north of Abrolhos (Figure 9a), which further strengthens the density stratification in the north. The subducted waters are partly entrained in the LC and partly entrained in the northward flowing Leeuwin Undercurrent (Domingues *et al*. 2007). In the south, the thermocline is relatively deep, likely due to the convergence of the LC (Figure 9a). The salinity stratification partly compensates the temperature stratification so that the density stratification in the upper ocean is relatively weak. The diffused thermocline is also likely due to the stronger LC eddy activity in the south. Subduction of the South Indian Ocean Central Water may also contribute (*e.g.*, Woo & Pattiaratchi 2008).

Inferred from fluorescence data, there are well-defined deep chlorophyll (biomass) maximum (DCM) layers north of Abrolhos Islands in late spring (Figure 9a). The DCM is at about 50 m depth off Ningaloo (22°S), where the wind-driven upwelling may be important (Hanson *et al*. 2005a), and at about 100 m depth in other shelf areas in the north. South of Abrolhos, the DCM layers still exist but are more diffused than those in the north, as shown in the average profiles north and south of Abrolhos.

There are similar decreasing trend in temperature and increasing trend in salinity from north to south in the upper 200 m along the shelf break (400 m) during the July–August 2005 (Figure 9b), except that the salinity maximum waters are no longer exposed to the sea surface, likely due to the stronger LC in winter that brings less salty tropical waters residing on the top of the local water mass. There also tends to be stronger mesoscale variability along the pathway of the LC, compared to spring, such as there are upwelling centres near 26°S and 34°S and downwelling centres near 22°S and 29°S. The upper ocean water column is less stratified compared to summer, likely due to the combined actions of the mesoscale eddies and mixing driven by the surface heat loss.

In July–August 2005, the DCM south of Abrolhos no longer exists and there is a near surface chlorophyll *a* maximum (Figure 9b). In the north, the DCM layers are slightly shallower than 100 m and are more closely linked to the mixed layer depths.

There are no suitable *in situ* observations that can well resolve the interannual variations of the biological properties in the LC system. Long term *in situ* monitoring of the biophysical properties in the LC is planned as part of the national Integrated Marine Observing System.

Nutrient dynamics

In the context of climate-driven ocean variability, we next consider the primary physical factors likely to influence nutrient fluxes, and thus primary production, in the LC. The main questions to be addressed are: what are the key processes that cause differences in the seasonal cycle of chlorophyll *a* concentrations north and south of the Abrolhos Islands? What are the drivers of the interannual variability of biogeochemical cycling along this coast?

Seasonal cycle

While north of Abrolhos enhanced nutrient concentrations are observed during the episodic upwelling events in summer (Hanson *et al*. 2005a), off the lower west coast, enhanced concentrations of dissolved nitrate and silicate have been observed during the austral autumn-winter, when the LC and its eddy field are strong and the continental shelf off the lower west coast of WA is flooded with the LC waters (Johannes *et al*. 1994; Thompson & Waite 2003; Hanson *et al*. 2005b; Koslow *et al*. 2008). This highlights strong seasonal differences in the LC influence north and south of Abrolhos, and therefore potentially the physical mechanisms controlling nutrient supply and primary production.

Mechanism 1 – Meridional erosion of seasonal thermocline drives nutrient injection into surface layers.

In the ITF region and north of Australia, the thermocline and nitrocline are relatively shallow, due to both excessive heat and freshwater inputs at the sea surface. As the tropical waters move south to join the LC and enter regions with a progressively deeper seasonal thermocline (Domingues 2006), the thermocline is eroded by vertical mixing and nutrients immediately below the thermocline could be mixed into the surface layer (Figure 10). South of Abrolhos, the DCM tends to be shallower than the nitrocline depth during the austral winter (Figures 9b and 11), likely due to less stratification and stronger vertical mixing.

This mechanism would clearly have a seasonal component, and be most likely to occur during periods of

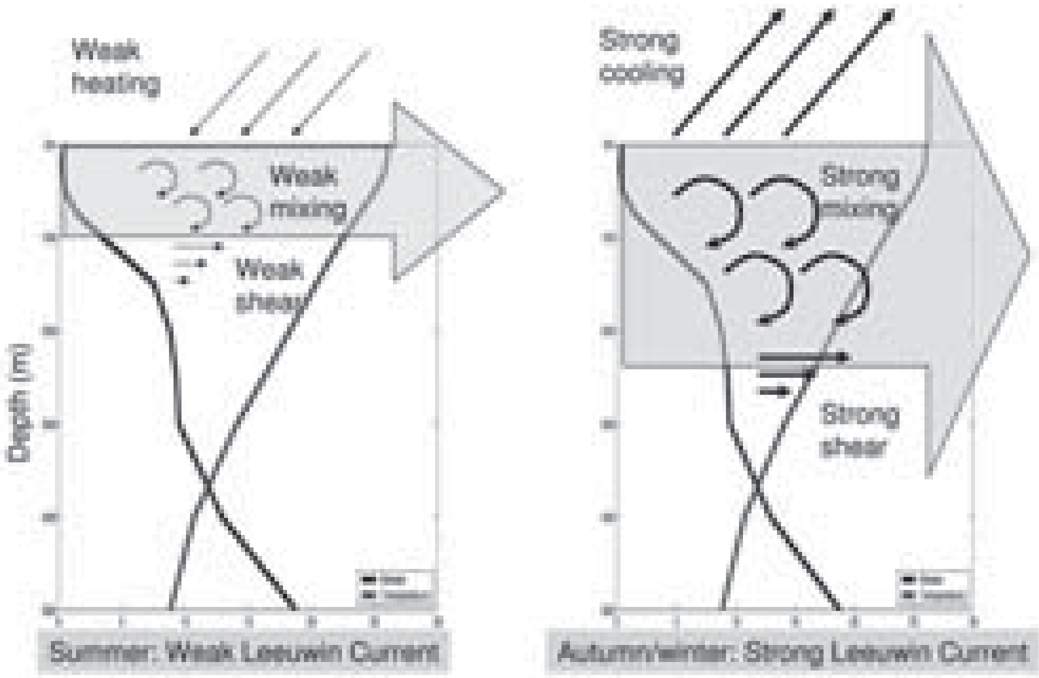

Figure 10. Schematic on how the erosion of the thermocline could bring nitrate into the euphotic layer in the LC system. The left panel is the summer condition when the LC is weak and the right panel is for the autumn/winter condition when the LC is strong. The schematic may also be applied to the difference between the El Niño and La Niña years. The vertical profiles of temperature (light line, unit: °C) and nitrate (dark line, unit: mg m^{-3}) are for the typical tropical waters near the source region of the LC.

accelerating LC (*i.e.*, March – June) and/or significant overall surface cooling, which could continue well into winter (Figure 10). Once this nutrient injection was exhausted, the vertical structure would likely revert to a more typical "tropical structure" (Cullen 1982) where diffusion of nutrients from below balances light from above, resulting in the formation of DCM (Figure 9a). This mechanism is considered to be one of the key processes for nutrient injection for the late autumn to early winter chlorophyll enhancement in the LC.

Mechanism 2 – The lifting/dropping of the nitrocline can increase/decrease primary production.

North of Abrolhos, the DCM is below the mixed layer depth (Figure 9), indicating that the DCM develops due to typical tropical conditions (Cullen 1982). The DCM depth is strongly correlated with the nitrocline depth (Figure 11), and there tends to be greater production when or where the DCM and nitrocline depths are shallower (Hanson *et al.* 2005a), indicating the sensitivity of upper ocean production to vertical movement of the thermocline (nitrocline) depth. This mechanism is responsible for the summer chlorophyll peaks north of Abrolhos.

Mechanism 3 – Both the horizontal and vertical advection related the eddy activities are important in enhancing ocean production in the LC system.

The presence of eddies generally contributes to enhance the biological productivity. Jenkins (1988), and later McGillicuddy & Robinson (1997) propose that in the oligotrophic Sargasso Sea, the time-varying eddy field may supply the required nutrients to sustain the observed primary production in the nutrient depleted gyre, due to eddy upwelling mechanism.

In the LC region, a Southern Surveyor section, which transverses an anticyclonic eddy and a cyclonic eddy, captures enhanced chlorophyll *a* signatures in both types of eddies (Figure 12). There tended to be enhanced primary production in mesoscale eddies, especially in the warm-core eddies associated with the LC (Waite *et al*. 2007; Feng *et al*. 2007). The nitrocline depth far exceeds the DCM depth in the eddy field compared with those in the LC (Figure 11), indicating that eddy activity may have enhanced the mixing of water masses. Another important role of eddy activity to regional biogeochemistry is their ability to transport properties horizontally across streamlines of the mean flow (Feng *et al*. 2007). In addition, the entrainment of productive waters from the continental shelf during the eddy formation (Greenwood *et al.* 2007; Paterson *et al*. 2008) results in the transport of particulate nutrients offshore.

Mechanism 4 – In situ nitrification can increase localized production.

There is possibly anecdotal evidence that nitrification may be occurring below the mixed layer depth in the anticyclonic LC eddies (Paterson *et al.* 2008).

Mechanism 5 – Benthic-pelagic coupling may be an important process on the seasonal cycle

From a simple nitrogen budget for the west coast of WA (Feng & Wild-Allen 2009), the benthic input may

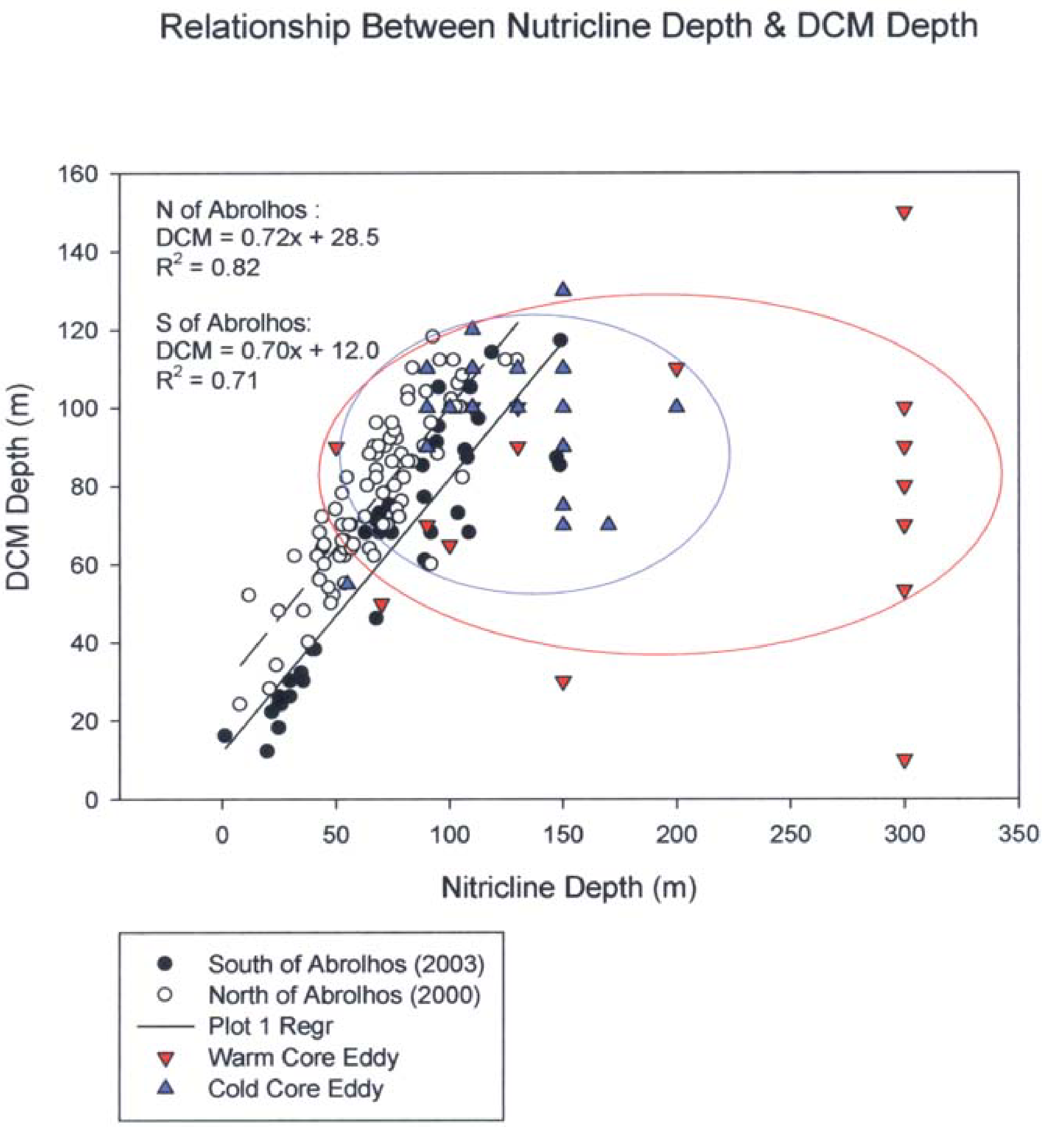

Figure 11. Relationships between the nitrocline depth and the deep chlorophyll maximum depth north and south of Abrolhos Islands along the LC, observed in November 2000 and October-November 2003, respectively. The straight lines are linear regressions for the two regions. The relationships in two offshore mesoscale eddies in October 2003 are also shown as triangles and are highlighted with elliptical shapes.

provide a significant portion of the nitrogen required to support the annual primary production on the continental shelf. Note that the benthic re-supply of nutrient to the water column is driven by storms and swells, which is coincident with strong onshore winds during the austral autumn-winter.

In summary, there are a few important physical and biological factors that potentially influence the seasonal cycle of the nutrient dynamics in the LC and on the shelf. Some factors have large-scale impacts and can affect the whole LC region, such as the air-sea heat flux (Figure 5). Some factors have different impacts on the nutrient dynamics north and south of Abrolhos, such as wind-driven upwelling being more effective north of Abrolhos due the shallow nitrocline, while the LC and eddy advection being more effective south of Abrolhos.

Interannual variations

During the La Niña years, the stronger LC is linked with a deeper thermocline (nitrocline) depth off the shelf edge so that the wind-driven upwelling is less effective in bringing deep nutrient onto the shelf, likely resulting in reduced summer production north of Abrolhos (Furnas 2007). The opposite would occur during the El Niño years.

In the south, with the assistance of the stronger eddy field and greater surface heat loss, the stronger LC during the La Niña years likely leads to stronger thermocline erosion and higher nutrient flux into the upper ocean, so that it is likely that the primary production is enhanced due to the increased LC strength. The opposite would likely occur during an El Niño year. Thus, in this region the primary production may not

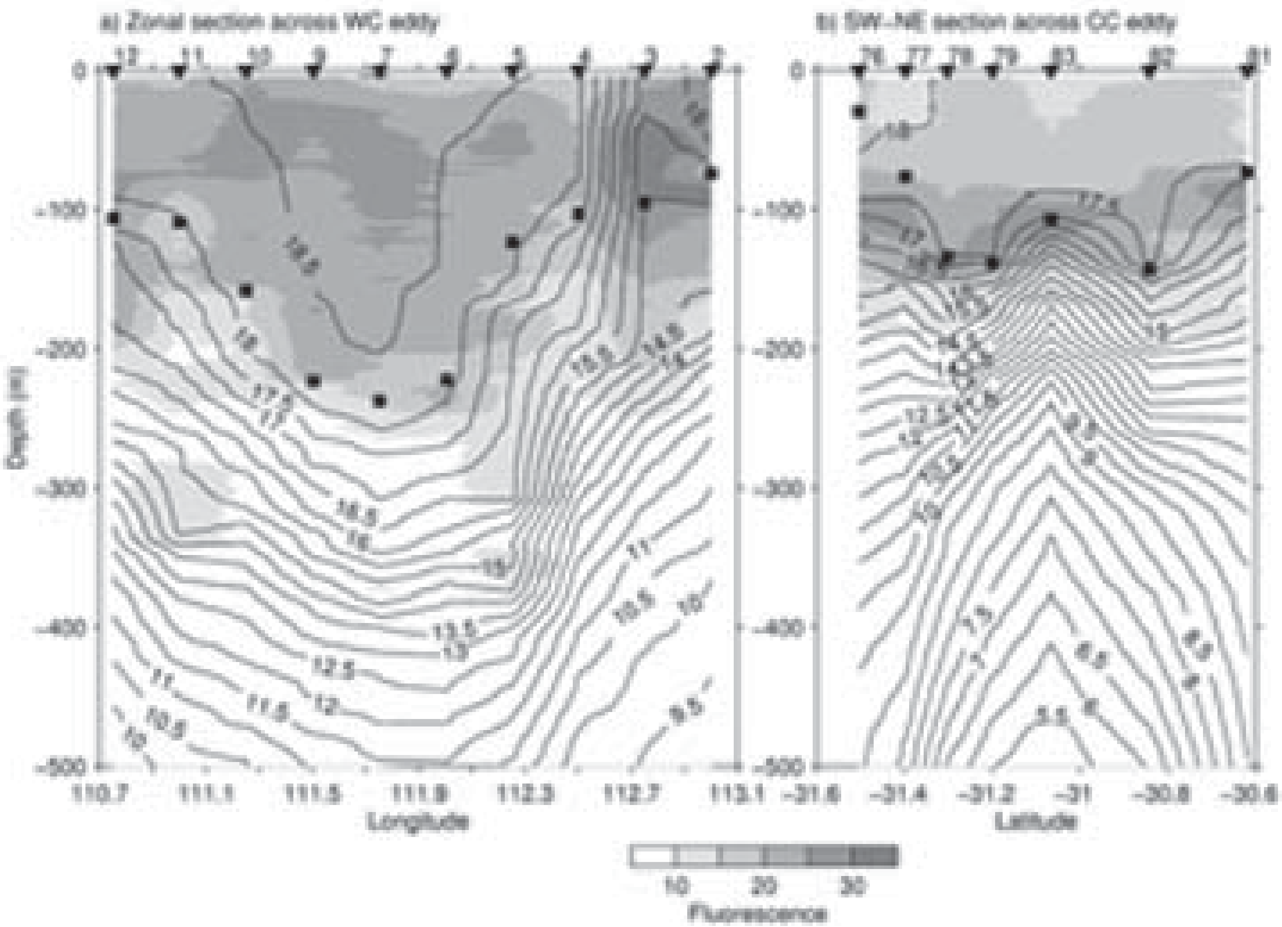

Figure 12. Vertical temperature structures along two transects across a warm-core (WC) eddy (a) and a cold-core (CC) eddy (b) from Feng *et al*. (2007).The shadings denote the fluorescence measurements and the solid squares denote the depth of the mixed layer. The station numbers of the CTD casts are denoted on the tops of the panels.

follow the global trend in response to ENSO (Behrenfeld *et al*. 2006), *e.g.*, there is a cooling trend of sea surface temperature (SST) during 1999–2004 off the lower west coast of WA (Pearce & Feng 2007), while there is a decreasing trend in chlorophyll *a* biomass over the same period (Figure 8).

Climate change and potential impacts

In the past 50 years, SST off the west coast, especially the lower west coast, of Australia, is warmed up faster than the average trend of the global ocean (Pearce & Feng 2007). From climate model simulations, it is suggested that in a warming climate, mean state in the tropical Pacific evolves towards a more El Niño-like condition, with reduced trade winds and reduced thermocline tilting (IPCC, 2007). The reduced thermocline tilting has produced a shallow thermocline anomaly in the Eastern Indian Ocean as well as along the west coast of WA (Wainwright *et al*. 2008). The observed trends in the tropical Pacific and Eastern Indian Ocean have shown that the Indonesian Throughflow may have reduced its strength by 20% in terms of volume transport. The shallowing of thermocline in the Eastern Indian Ocean implies a reduction of the pressure gradient that drives the LC, and hence a reduction in the LC transport (Feng *et al*. manuscript in preparation). In regional climate, there has been a significant decrease of rainfall in the southwest WA region during the recent decades, due to fewer winter storms arriving at the coast (IOCI 2002). The reductions of the LC and the winter storms are expected to reduce nutrient inputs into the upper ocean south of Abrolhos. North of Abrolhos, where the ocean productivity is more sensitive to the depth of the nitrocline, the ocean production is likely to increase with a weakening LC. In addition, with the global warming the surface layer of the ocean becomes lighter and less likely to mix with the colder, denser water below, so that in principle, primary productivity would be reduced in a general sense (Behrenfeld *et al.* 2006).

To accurately predict these effects, we need to have a combined assessment of the effects of the global warming, changes in regional circulation, and changes in the regional air-sea heat fluxes. This will be achieved through long-term monitoring of the shelf environment and climate model downscaling for the LC system. Further research on the biogeochemical responses of the LC system to climate variability and climate change is necessary to underpin the long term management of the marine resources in this region.

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