Physical-biogeochemical dynamics of a surface flooded warm-core eddy off southeast Australia.

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Abstract

Warm core eddies (WCEs) formed from the East Australian Current (EAC) have a large impact on the heat, mass and biogeochemical budgets of the western Tasman Sea. The development and separation of an EAC WCE during October-December 2008 was observed using remotely sensed temperature, ocean colour and sea-level elevation, two Argo floats, a shipboard CTD, a shelf mooring array and a 15 day

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deployment of a Slocum glider. The eddy formed from an EAC meander during the first half of 2008. In the two months before separation in early December, fresher and warmer EAC water flooded the top of the eddy, submerging the winter mixed layer. The vertical transport due to submergence was $\sim 5$ Sv, about 1/6th of the mean southward flow of the EAC in October. The surface waters of the eddy were low in chlorophyll throughout the observations. Chlorophyll concentration at the top of the submerged nutrient-rich winter mixed layer ranged from 0.5-2 mg m$^{-3}$, with depth integrated chlorophyll concentration ranging between 25-75 mg m$^{-2}$. Elevated levels of coloured dissolved organic matter in the submerged layer correspond to oxygen depletion, suggesting respiration of sequestered production. Finally, a comparison is made with observations from WCEs in 1978 and 1997 in which surface flooding did not occur and radiative heating stratified the top 50 m. In these capped eddies surface chlorophyll concentrations were an order of magnitude higher than flooded eddies, but depth integrated chlorophyll was similar. EAC WCEs are more productive than surface chlorophyll concentrations would suggest, with the vertical position of primary productivity depending on whether, and to what depth, the winter mixed layer is submerged.

Key words: anti-cyclonic eddy, East Australian Current, Tasman Sea, deep chlorophyll maximum, autonomous underwater glider,

1 Introduction

The role of mesoscale processes (Ridgway and Dunn, 2003) and in particular warm core eddies (Hamon, 1965; Cresswell and Legeckis, 1986) in the East Australian Current system has long been recognised as both important and complex (Cresswell, 1983). Warm core eddies (WCEs) off southeast Australia are formed from low nutrient sub-tropical water that is advected south as part of the East Australian Current (EAC). Eddies separate from the main flow

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intermittently, with a mean interval of 90-180 days (Mata et al., 2006; Wilkin and Zhang, 2007), and generally migrate parallel to the Australian coast in a poleward direction. The path of an individual eddy can be complicated by reabsorption by the EAC (Nilsson and Cresswell, 1981), coalescence of one or more eddies (Cresswell, 1982), or injections of filaments of surrounding water into the eddy (Scott, 1981; Cresswell, 1983).

The biological processes within (Tranter et al., 1980; Griffiths and Wadley, 1986) and surrounding (Tranter et al., 1983, 1986) EAC WCEs were extensively studied between 1975-82. The differences in phytoplankton species from within and surrounding an EAC WCE have been found to be minor, although concentrations are elevated when compared to the tropical Coral Sea water from which they are formed (Jeffrey and Hallegraeff, 1987). In contrast, crustaceans inside an EAC WCE were a mixture of warm-water and cold-water species and were dissimilar to the crustacean fauna in either the Tasman Sea or the Coral Sea (Griffiths and Wadley, 1986). It has also been proposed that EAC WCEs result in enhanced fish catches (Young et al., 2001). EAC WCEs are not simply anti-clockwise rotating masses of Coral Sea water found anomalously south, but have dynamic physical processes that drive biological processes.

The biological processes in two eddies, Eddy F in 1978 and Eddy J in 1980, were carefully investigated for their contrasting biological response. In Eddy F there was anomalously high chlorophyll in the surface mixed layer (Tranter et al., 1980; Jeffrey and Hallegraeff, 1980). Tranter et al. (1980) explained the surface phytoplankton bloom by the easing of light limitation of phytoplankton growth in the eddy. Initially Eddy F had a > 200 m deep surface mixed layer. In late spring solar heating decreased the depth of the surface mixed layer to 50 m, and phytoplankton in the top 50 m were able to utilise the high nutrients remaining from winter. Surface chlorophyll concentrations were close to 1 mg m$^{-3}$ and depth integrated chlorophyll ranged between 22-55 mg m$^{-3}$ (Jeffrey
In the second eddy, Eddy J, fresher and warmer EAC water flooded onto the top of the deeply mixed over wintering eddy (Tranter et al., 1982). The over-winter nutrients were submerged and flooding waters that remained at the surface were nutrient deplete. Even though Eddy J was more stratified than Eddy F, surface chlorophyll concentration was lower.

Phytoplankton pigments, species and light climate were sampled in a third warm-core eddy, Eddy Mario, in April-May 1981 (Jeffrey and Hallegraeff, 1987). Eddy Mario’s life history was more complicated than either Eddy F or J (Cresswell and Legeckis, 1986), or indeed the presently studied eddy. The eddy contained two submerged isothermal layers as a result of the coalescence of two earlier eddies (Jeffrey and Hallegraeff, 1987). As a result, vertical profiles of light and pigments did not consistently show either a surface or sub-surface maximum, and, despite only a 4-6 day period, showed large horizontal variability in bio-optical properties. Depth integrated (0-150 m) chlorophyll varied between 26-60 mg m$^{-2}$, and was dominated by small phytoplankton (< 15 μm). In enriched areas diatoms concentrations were relatively high. Jeffrey and Hallegraeff (1987) found detrital pigments, identified from chromatography and spectral curve analysis, were dominant below 100 m depth, particularly in the top of the two isothermal layers.

All previous studies of the biological response in EAC WCEs were undertaken before the newly increased capacity in remotely-sensed observations and high-resolution vertical profiling made available through the Australian Integrated Marine Observing System (IMOS). In particular, autonomous underwater gliders with bio-optical sensors are providing new insights into both physical and biological properties of sub-mesoscale phenomena (Niewiadomska et al., 2008). This paper details observations of an EAC WCE using remotely sensed temperature, ocean colour and sea-level elevation, two Argo floats, a
The eddy that is the focus of this paper separated from an EAC meander in the first week of December, 2008. The eddy had low surface chlorophyll concentrations, similar to Eddy J described above, and typical of EAC WCEs. Like Eddy J, the low surface chlorophyll concentrations were due to the surface flooding of the deeply mixed eddy by more buoyant nutrient deplete waters. The high-resolution observations of sub-surface bio-optical properties obtained from the Slocum glider deployment provides new insights into biological processes in the large volumes of isolated nutrient-replete waters that are submerged in surface flooded eddies off the coast of southeast Australia.

2 Material and Methods

Remote sensed observations. Surface temperature and colour data was extracted from the NASA Ocean Color website. MODIS data used is processed to level 2 (temperature and chlorophyll algorithms applied on the orbital swath), while the SeaWiFS and CZCS data are further processed by spatial interpolation onto a regular grid (level 3). Satellite altimeter data were provided by NASA/CNES (Topex/Poseidon, Jason-1, -2) and ESA (ERS2, ENVISAT).

Glider mission. A Slocum glider was deployed off Port Stephens (32°44′S, 152°14′E) on the southeast Australian coast at 10:55 am on the 25 November 2008 and retrieved off Jervis Bay (35°3′S, 151°16′E) on the 11 December 2008 (Fig. 1). The glider generally undulated between the surface and 180 m, depth permitting. For a period between the 5-7 December the glider’s maximum depth was 100 m. The glider velocity was primarily determined by the ambient current, which initially advected it south in a filament of the EAC which was flooding the surface of the warm-core eddy. The glider velocity relative to the
flow was generally steered toward the centre of the eddy from deployment until the 3 December and then away from the centre until recovery. This produced a track that spiralled to close to centre of the eddy and then back to the perimeter while undertaking one rotation (Fig. 1).

The glider sensors included: a Seabird-CTD, a WETLabs BBFL2SLO 3 parameter optical sensor (measuring Chlorophyll-a, coloured dissolved organic matter (CDOM) and backscatter at 660 nm) and an Aanderaa oxygen optode. Descriptions of the correction of thermal lag of the conductivity cell and non-photochemical quenching of the fluorescence estimates are given in Appendix A and B respectively.

Shipboard CTD. During a RV Southern Surveyor cruise (SS200810) 47 CTD stations were undertaken primarily in shelf and slope waters between Broughton Island (32°30′S) and Sydney (34°S). In this paper data from CTD station 34 (33°17′S, 152°35′E on the 18 October 2008 at 10 am AEST) is used. The station was ~24 km east of the shoreward edge of the EAC with 22.7°C, 35.4 salinity surface waters. The surface current was flowing SSE at ~1.5 m s⁻¹. Some fraction of the EAC water at this time was flooding the surface of the WCE, while the remainder flowed east as the Tasman Front Jet. The ship underway CO₂ measurements were undertaken with a General Oceanics Inc. automated system (Model 8050) as part of the IMOS ships of opportunity (SOOP) program.

Argo floats. Two Argo floats sampled waters in the eddy before separation. After separation no Argo floats sampled the eddy. The Argo profiles used are float 5900871 profile numbers 111 and 112 on the 14 and 24 October 2008; and float 5900562 profiles 147, 148 and 149 on the 20 and 30 October and the 10 November.

Sydney mooring array. The Sydney mooring array consists of the Sydney Water Ocean Reference Station (ORS065) that has been maintained since 1991,
and the IMOS moorings SYD100 and SYD140 that form a shore normal line extending from Bondi, Sydney (33°55′S). Data has been used from the 11 November until the 8 December 2008. The ADCP on SYD140 failed between the 11 November and the 5 December. For more technical details of the mooring array see Roughan et al. (this issue).

3 Eddy formation

In mid January 2008 a WCE was spawned off the EAC as a result of its southward migration, and the simultaneous westward propagation of a cold core eddy at 33°30′S. After this separation, a meander in the EAC began to grow. Calculated geostrophic surface currents (not shown) indicate that for the first few months of 2008 this meander received EAC waters from the north, and lost water to the south. The water lost to the south formed mesoscale features in the Tasman Sea south of 34°S.

By May 2008 the meander, centred at 33°S, 155°E, was directing flow primarily to the east as part of the Tasman Front Jet. This state in the eddy formation was still evident at the beginning of September (Fig. 2A). The water at the core of the meander was significantly cooler than the EAC that was flowing around the edge of the meander. Over time the less dense edge water flooded the centre of the eddy and submerged the existing deeply mixed surface layer, as has been previously observed (Cresswell, 1983; Tranter et al., 1982). This creates a submerged layer, ~200 m thick. Out of contact with surface heat and freshwater fluxes, the layer maintains a relatively constant temperature and salinity over time, and is referred to as an isothermal layer.

The rate of submergence of the isothermal layer can be estimated from ARGO vertical profiles leading up to the separation of the eddy (Fig. 3). From the salinity and temperature profiles 50 and 23 days before the separation of the
eddy, the isothermal layer was submerged 80 m (marked by blue intervals on the vertical profiles of temperature and salinity in Fig. 3). The surface mixed layer also deepens before separation, demonstrating that its depth is being determined by the interface between the lighter flooding EAC water and the denser submerged layer.

The glider salinity observations of the separated eddy (Fig. 4) show that this vertical movement was relatively uniform across the eddy. The spring EAC water that flooded the eddy forms a surface layer between 60 and 100 m deep with the isothermal layer lying beneath. Assuming that the 80 m vertical movement over 27 days occurred evenly across the eddy surface area of $143 \times 10^9$ m$^2$ implies a vertical transport of $\sim$5 Sv, a significant fraction of the average October southward flow of the EAC of 30 Sv (Ridgway and Godfrey, 1997). The southward movement of the eddy from

The Slocum glider sampled a filament that was the last water to become entrained in the WCE before separation from the EAC (Fig. 2B & 4). The final injection of EAC water into the eddy ceased about the 3 December, as the eddy became fully separated. The mooring array on the continental shelf off Sydney (Fig. 5) measured strong southward velocities (up to 1.25 m s$^{-1}$ at the surface) of 19-22°C water from the 25 November until the 3 December (Fig. 5B). At the moorings array on the 100 m (Fig. 5B) and 140 m (Fig. 5C) isobaths, the 20+°C waters that entered the eddy are evident in the top 40-60 m. The inshore mooring (Fig. 5A) has cooler surface temperatures and only moderate southward velocities that suggest the water flowing past it was not entrained into the eddy.

The glider observations of the final EAC filament to flood the eddy (with $S < 35.55$) give a filament depth of 60-80 m, which approximately corresponds to the depth of the -0.5 m s$^{-1}$ contour at the SYD100 mooring (Fig. 5). The width of the $S < 35.55$ filament perpendicular to the flow is $\sim$25 km. Assuming
a depth averaged velocity in the filament of 1 m s$^{-1}$ over 70 m gives a transport of 1.75 Sv. In order to achieve the 5 Sv of flow that submerged the eddy in a 70 m deep filament flowing at 1 m s$^{-1}$ requires a width of 71 km. Such a broad filament is visible in Fig. 2A and B.

The glider path shows three water masses distinguishable by salinity (Fig. 4) - the submerged layer ($S = 35.64 - 35.66$), a surface mixed layer ($S = 35.56 - 35.62$), and the last filament of EAC water discussed above ($S = 35.46 - 35.54$). In October, the climatological salinity of surface shelfbreak water at 30$^\circ$S, 32$^\circ$S and 34$^\circ$S is 35.51, 35.57 and 35.58 respectively (CSIRO Atlas of Regional Seas (CARS) climatology, Ridgway and Dunn (2003), Baird et al., this issue). The final filament entering the eddy at 34$^\circ$S has low salinity, characteristic in October of EAC water that is located anomalously south.

4 Physical characteristics of the separated eddy

The eddy separated on approximately the 3 December and by the 13 December was an isolated water mass that could be defined by surface temperatures above 20$^\circ$C, positive sea level elevation anomaly and surface chlorophyll concentration < 0.25 mg m$^{-3}$ (Fig. 2C & Fig. 6C). The surface area of the eddy can be approximated assuming an elliptical shape. On the 13 December, ~10 days after the final injection of EAC water, the eddy had major and minor axes of lengths of ~232 and ~196 km (Fig. 6C), giving a surface area of 143 $\times$ 10$^9$ m$^2$.

The rotation rate of the eddy can be estimated from the curl of the velocity field, from Lagrangian paths or from radial speed. Angular velocity, $\theta$, is given by half the vorticity (the curl of the velocity field) in radians. The time to complete one solid body rotation is given by $2\pi/\theta$. On 30 November, the curl of the surface velocity field calculated from geostrophic flow of the observed
sea level elevation is \( \sim 1.6 \text{ rad d}^{-1} \), so the time to complete a revolution is 7.85 days. Lagrangian paths calculated using a stationary velocity field suggest that close to the centre of the eddy it takes 6 days to return to the same geographical location. Thirdly, a regression of radial speed versus radial distance gives a period of 8.4 days. For the glider moving along an 0-200 m depth undulating path with relative glider velocity primarily in a radial direction, the time to make one rotation (return to the same geographical location) was 7.01 days.

5 Biogeochemical characteristics of the separated eddy

Surface chlorophyll concentrations of the eddy throughout the study period are low (Fig. 6). During formation, the eddy centre surface chlorophyll concentration was 0.2 mg chl m\(^{-3}\) (Fig. 6A). Outside the eddy to the east concentrations in the Tasman Sea waters exceed 1 mg chl m\(^{-3}\). A climatology of surface chlorophyll concentrations for the Australian region (Condie and Dunn, 2006) has spring Tasman Sea values varying between 0.2 and 0.5 mg chl m\(^{-3}\). Surface filaments of up to 0.4 mg chl m\(^{-3}\) are found on the perimeter of the eddy, but calculated geostrophic currents (Fig. 2A) suggest are likely to have been advected northeast out of the eddy rather than retained.

Shortly before separation (Fig. 6B), surface chlorophyll concentration varies between 0.1 - 0.5 mg chl m\(^{-3}\). To the northeast of the centre (35\(^{\circ}\)20\('\)S, 153\(^{\circ}\)00\('\)E) and along the shelfbreak to the west (34\(^{\circ}\)40\('\)S, 151\(^{\circ}\)15\('\)E) concentrations of up to 0.5 mg chl m\(^{-3}\) can be seen along flow streamlines, suggesting injections of EAC water during formation have affected the later biological response within the eddy. In particular, the filament on the shelf break was sampled by the glider on the 28 November (Fig. 7), with a 10-20 km wide, 60 m deep filament of 1.5 mg chl m\(^{-3}\).

After separation surface chlorophyll concentration clearly distinguishes eddy
water from that outside the eddy (Fig. 6C). At the eddy centre the chlorophyll concentrations are at their lowest (0.1 mg m$^{-3}$). On the edge, particularly to the east on the edge of a cold-core eddy, concentrations reached 0.8 mg m$^{-3}$. By early 2009 the eddy surface chlorophyll concentrations were similar to the surrounding water (Fig. 6D). The eddy itself is still a distinct feature with sea level elevation and temperature anomalies of 0.5 m and 2°C relative to the surrounding waters (Fig. 2D). Both eddy and surrounding waters have low surface chlorophyll concentrations due to a mid-summer minimum typical for the Tasman Sea. The surface characteristics of the eddy from October 2008-January 2009 conform to the view that EAC warm core eddies are unproductive islands of tropical waters advected south into the Tasman Sea.

A sub-surface maximum for chlorophyll concentration with values up to 2 mg m$^{-3}$ occurs along the transect (Fig. 7). Within the eddy the sub-surface maximum is found between 50 m and 100 m, with a thickness of $\sim$20 m and is located at the top of the submerged layer. The closest the glider came to the centre was on the 2-3 December (Fig. 1) during which time the sub-surface maximum lifted from 85 to 60 m. It is difficult to determine whether this relatively small depth change is a product of the formation of the eddy, a local (in time and/or space) instability such as internal waves, or vertical velocities at the centre. In any case, it is no bigger than variations in surface mixed layer depth seen throughout the rest of the eddy (Fig. 7).

Coloured dissolved organic matter (CDOM) inferred from fluorescence is low in the eddy surface mixed layer (Fig. 8A) due to photo-oxidation of CDOM to optically inactive forms of DOC (Oubelkheir et al., 2005; Niewiadomska et al., 2008). In the submerged isothermal layer which was at the surface $\sim$2 months earlier, CDOM content is increased due to bacterial production/remineralization of organic matter during the time submerged below the euphotic zone. Significantly, CDOM concentration is no higher in the region of the sub-surface chlorophyll maximum than in the rest of the submerged
layer. In another surface flooded EAC WCE, Jeffrey and Hallegraeff (1987) also found high values of detrital pigments in the submerged isothermal layer.

The concentration of dissolved oxygen is highest in the surface mixed layer (Fig. 8B), averaging ~93 % of saturation. The surface layer is well mixed and in contact with the atmosphere. Oxygen in the submerged layer is between 80 and 90 % saturation, with regions of high CDOM having the lowest oxygen. Assuming that the submerged layer was close to equilibrium before it was submerged, ~0.5 ml L⁻¹ of oxygen has been consumed. This corresponds to the respiration of 0.5 ml L⁻¹ × 1000 L m⁻³ × 1.42 mg O₂ ml⁻¹ × 1/32 mmol O₂ (mg O₂)⁻¹ × 16/138 mmol N (mmol O₂)⁻¹ × 1.59 mg Chl (mmol N)⁻¹ ~ 4 mg Chl m⁻³ phytoplankton biomass (although this respiration could be of either phytoplankton or zooplankton) [Chl:N from Fasham (1993); N:O₂ from Kirk (1994)] The surface flooding has sequestered a significant quantity of organic matter that has been respired.

At the top of the isothermal layer in the portion with elevated chlorophyll concentration, oxygen is reduced, but less than in the rest of the isothermal layer. This could be either because (1) this section of the isothermal layer retained some contact with the atmosphere during the flooding, (2) vertical diffusion resulted in an intermediate concentration between the surface layer and the bulk of the isothermal layer or (3) production associated with the sub-surface chlorophyll maximum offset reductions in oxygen due to respiration. With such a sharp gradient of properties such as salinity and oxygen, it is likely the increased oxygen in the top of the isothermal layer is mostly due to local production.
6 Discussion

The phenomena of less dense water flowing over the top of a WCE has been described for a number of boundary current systems, and been referred to using terms such as submerging (Jeffrey and Hallegraeff, 1987) and flooding (Tranter et al., 1982) in the Tasman Sea, overwashing (Hitchcock et al., 1985) in the North Atlantic, overwashing (Tomosada, 1978) in the North Pacific and surface injection (Dietze et al., 2009) in the eastern Indian Ocean. In each of these systems, the vertical transport associated with surface flooding represents, for a few weeks at a time, a significant sink of water for major boundary currents.

The high-resolution bio-optical sampling of the surface flooded WCE provides new insights into their productivity. The sub-surface chlorophyll maximum at the top of the isothermal layer demonstrates that flooded WCEs are more productive than their surface mixed layer chlorophyll concentrations would suggest. The chlorophyll maximum is at the top of the isothermal layer, suggesting production is consuming nutrients from the top of the deeply mixed winter layer rather than upwardly diffused nutrient. As such, only the unused nutrients within the euphotic zone drive sub-surface production, and vertically-integrated primary production in the WCE is light limited. The perception from surface colour that WCEs are unproductive and nutrient limited may be incorrect for the surface flooded WCEs that are common in the Tasman Sea.

6.1 Eddy mass and carbon budget

As a result of their large size and occurrence a few times a year, EAC WCEs play a significant role in both the mass and carbon budgets of the Tasman Sea. For an isothermal layer $\sim 200$ m thick, the volume of water submerged from the surface to below the mixed layer during the formation of the eddy is
∼28 × 10^{12} \text{ m}^3. Using the calculation of respired oxygen above with a C:O_2 ratio of 106/138 (Kirk, 1994), ∼0.2 g C m\(^{-3}\) of organic carbon was sequestered in the layer, for a total of sequestered organic carbon in the isothermal layer of 200 m × 143 × 10^9 \text{ m}^2 × 0.2 g C m\(^{-3}\), or 5.7 × 10^{12} g C. For comparison, Australia’s carbon emissions in 2004 were 104 × 10^{12} g C yr\(^{-1}\). The mass of dissolved inorganic carbon (DIC) in the submerged layer, based on a surface DIC concentration of 2 mol C m\(^{-3}\) (Macdonald et al., 2009), is 686 × 10^{12} g C.

The downward transport within the eddy must be accompanied by upward transport outside the eddy. Low salinity (Fig. 4) and oxygen (Fig. 8B) water to the north of the eddy (900 km along the transect) suggests that the submergence drives upwelling around the edges of the eddy. The high CDOM and presumably DIC concentrations of the upwelled water illustrate that the net fluxes associated with the submergence may be offset by upwelling. In any case, the Tasman Sea generally absorbs atmospheric CO_2 (Takahashi et al., 2002; Macdonald et al., 2009), and the surface waters at the CTD station were significantly below atmospheric (pCO_2 of 340 µatm). Furthermore, model simulations suggest upwelled waters in the Tasman Sea rarely outgas (Macdonald et al., 2009). So efficient vertical transport associated with submerged isothermal layers is likely to increase regional oceanic carbon absorption.

6.2 Capped warm-core eddies.

As mentioned in the introduction, EAC WCEs can have high surface productivity. Tranter et al. (1980) sampled a WCE off the NSW coast in mid September and mid November 1978 (Fig. 9). In September, the mixed layer depth was ∼250 m (shown to 150 m in Fig. 9C), and the surface chlorophyll concentration was low. Two months later, a cap of ∼0.5°C warmer surface water formed due to solar heating, with decreased nitrate concentration and
elevated phytoplankton biomass. The elevated biomass is clear in the CZCS image of the 21 November 1978 (Fig. 9E).

An even more dynamic event occurred at the same time of the year in 1997 (Fig. 9D, F). Fortuitously a geomagnetic field study in the region (Hitchman et al., 2000) sampled the eddy in September when surface chlorophyll concentration was low (Fig. 9D). Two months later surface chlorophyll increased by an order of magnitude (Fig. 9F). The November 1997 event produced the most dramatic contrast between surface chlorophyll concentration inside and outside a EAC WCE off southeast Australia in the satellite record (1978-1986, 1996-97, 1997-2009).

Warm core eddies migrating from tropical to subtropical regions are potentially productive because surface stratification in the tropics allows nutrients to build-up at intermediate depths. If the waters are mixed vertically, then nutrients can be brought to the surface under light limiting conditions. Subsequent stratification eases light limitation, and utilisation of the available nutrients in the mixed layer produces a surface bloom. This dynamic could not occur in 2008, and in fact rarely does, because the eddy remained attached to the EAC through until the end of November. The EAC water in spring is both fresher and warmer than in winter, and so submerges the winter mixed layer.

6.3 Summary

A suite of observations of a surface flooded warm-core eddy off southeast Australia in late 2008 reveal that: (1) the vertical transport in surface flooded WCEs off southeast Australia can account for a sizeable fraction of the EAC transport during eddy formation; (2) phytoplankton within EAC WCEs with submerged winter mixed layers can utilise some of the over-wintering nutrients
and therefore produce depth integrated chlorophyll equal to capped eddies; (3) vertical transport of organic matter in the submerged isothermal layer represents an important carbon flux in the Tasman Sea. Warm core eddies can have drastically different biological responses depending on whether, and to what depth, the winter mixed layer is submerged. Future observational and theoretical work should focus on the factors leading up to separation that determine whether a surface flooded or capped eddy separates.

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A Appendix A. Correction for thermal lag effects in salinity estimate

The effect of thermal lag on calculations of salinity from conductivity have been correct using the approach of Morison et al. (1994) with coefficients $\tau = 14$ s and $\alpha = 0.13$.

B Appendix B. Correction for non-photochemical quenching of fluorescence

The WETlabs ECOpuck deployed on the Slocum glider determines fluorescence from excitation/emission wavelengths of 470/695 nm. Factory calibrations are used to convert fluorescence to chlorophyll. In situ measured fluorescence intensity is reduced in the presence of ambient light, a process called non-photochemical fluorescence quenching (Falkowski and Kolber, 1995; Sackmann et al., 2008). To correct for the reduced estimate of chlorophyll during daylight hours, a light-dependent correction of chlorophyll for the deployment has been developed:

$$chl_{cor} = chl_{uncor} \left(1 + \frac{I_z}{300}\right)$$

where $chl_{cor}$ is the corrected estimate of chlorophyll, $chl_{uncor}$ is the estimate of chlorophyll without correction, and $I_z$ is the light at depth $z$ of the observation. The value of 300 W m$^{-2}$ used in Eq. B.1 is based on minimising the vertical gradient of chlorophyll in the top 40 m on the 1 December when the glider was close to the centre of the eddy (Fig. 10). This time was chosen as (1)
low chlorophyll concentrations improves the accuracy of Eq. B.2, (2) strong stratification below the level of quenching reduces the risk of changes in vertical profiles with time and (3) the glider remained close to the eddy centre between 1-2 December and showed little change in its physical or bio-optical properties. The effect of the correction term throughout the eddy at 20 m depth is shown in the insert of Fig. 10.

The vertical light field is calculated from:

\[ I_z = I_0 e^{\frac{k_w + k_p P}{\cos \theta} dz} \]  \hspace{1cm} (B.2)

where \( I_0 \) [W m\(^{-2}\)] is the surface irradiance at the local mean (or sundial) time based on orbital cycles (Brock, 1981), \( k_w = 0.04 \text{ m}^{-1} \) is the attenuation of light due to non-phytoplankton components, \( k_p = 0.03 \text{ m}^{-1} \) (mmol N m\(^{-3}\))\(^{-1} \) is the nitrogen specific attenuation rate due to phytoplankton (Oschlies and Schartau, 2005), \( P \) is the phytoplankton biomass [mmol N m\(^{-3}\)] and \( \theta \) is the azimuth angle of the unscattered light path through the water.

For the calculation of \( I_0 \), surface albedo is a function of azimuth angle according to Fresnel’s equation (Kirk, 1994), and Snell’s law is used to account for the refraction of light at the air/water interface altering the pathlength of light through the water (Kirk, 1994). A 6 hourly averaged reduction in surface irradiance from the NCEP reanalysis is used to account for clouds (Kalnay et al., 1996).

For the calculation of \( I_z \), observed fluorescence is used to estimate \( P \). Within the surface mixed layer chlorophyll is relatively constant, so fluorescence-estimated chlorophyll will be a reasonable approximation of the overlying phytoplankton biomass. Using observed fluorescence in the sub-surface chlorophyll maximum will overestimate overlying phytoplankton biomass, and therefore underestimate the light. However, since the quenching effect has asymptoted to zero by the depth of the sub-surface maximum, this does not adversely
affect the correction of quenching.

Conversion from \( chl \) to \( P \) is based on 1.59 mg chl (mmol N)\(^{-1} \) (Fasham, 1993). The use of a chlorophyll to calculate phytoplankton biomass is not ideal due to variations the Chl:N ratio with size (Baird et al., 2007). In the centre of the eddy this will not be significant as non-phytoplankton components dominate the attenuation, but may be important on the shelf.

To estimate \( P \) requires the corrected chlorophyll which is not initially known. First the light level is determined from the uncorrected chlorophyll using Eq. B.2. The initial estimate of the ambient light, \( I_z \), is then used to calculated an improved chlorophyll estimate based on Eq. B.1. The improved chlorophyll estimate is used to provide a further improved ambient light estimate, which then provides a better estimate of chlorophyll from the original uncorrected fluorescence. Two further iterations of this procedure produces an error in the estimate of ambient light of less than 1 %.

References


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Fig. 1. Glider path and locally calculated depth-averaged velocity relative to the (translating) eddy centre. Glider location is marked at midday Australian East Standard Time. The IMOS Sydney Mooring array depth averaged velocity for the 28 November is also shown. Shading indicates whether the local water is in the filament of EAC waters filling the eddy (grey) or within the eddy (black). The eddy centre calculated using altimetry on the 1 and 5 of December are marked with a bold S and F. The black points are plotted in the insert for distance from the centre of the eddy against relative velocity. The fit of the line (forced through the origin) gives one estimate of the period of the eddy.
Fig. 2. Surface temperature (Aqua MODIS, Level 2) with tidal-residual, isostatically-adjusted sea level anomaly using the Jason-1, Jason-2 and Envisat altimetry. The 200 m isobath is shown as a dashed black line. The Panels are (A) 93 days and (B) 3 days before eddy separation, and (C) 10 days and (D) 47 days after.
Fig. 3. Vertical profiles of waters entering and within the eddy. Colours show temperature (grey, °C), salinity (purple), nitrate (red, mmol m⁻³) and fluorescence/chlorophyll (green, -). Thin solid lines are ARGO float profiles, with the line label giving time in days before eddy separation. Dashed lines are EAC waters upstream of the eddy at 33°17'S, 152°35'E on the 18 October 2008 at 10 am AEST observed by the RV Southern Surveyor (SS200810). The thick solid lines are from close to the eddy centre on 3 December 2008, the day of separation (35°18'S, 152°28'E). The blue intervals show sinking of vertical profiles of temperature and salt from 50 to 23 days before separation.
Fig. 4. Salinity calculated from conductivity with a correction for the temperature lag of the conductivity sensor (see Appendix A). Clear sky irradiance is shown as a black line at the surface with the local date. Potential density contours are shown every 0.25 kg m$^{-3}$, with 0.5 increments labelled and thickened.
Fig. 5. Vertical profiles of temperature (shading) and north-south velocity (contours) from (A) ORS065 mooring at 33°54.40′S, 151°18.97′E; (B) SYD100 mooring at 33°56.63′S, 151°23.03′E; and (C) SYD140 mooring at 34°00.08′S, 151°27.92′E (see locations on Fig. 1). Velocity contours every 0.25 m s⁻¹, north being positive, with 0, 0.5 and 1 labelled and bold (and -0.25 in C.). ADCP failed at SYD140 before 5 December.
Fig. 6. Surface chlorophyll concentration (Aqua MODIS, Level 2) with geostrophic velocity calculated from sea level anomaly in Fig. 2. Arrows represent Lagrangian paths for the 24 hours leading up to the midday satellite image. The 200 m isobath is shown as a grey line. The Panels are (A) 93 days and (B) 3 days before eddy separation, and (C) 10 days and (D) 47 days after.
Fig. 7. Chlorophyll concentration along the glider transect calculated from fluorescence, corrected for non-photochemical quenching. Clear sky irradiance is shown as a black line at the surface with the local date (Nov/Dec 2008). Temperature contours are shown every 0.25°C, with thickened integer contours that are labelled. The effect of non-photochemical quenching has been removed (see Appendix B). The grey line gives depth-integrated chlorophyll concentration (for the depth of the glider track).
Fig. 8. Vertical profile along the glider transect of (A) coloured dissolved organic matter (ppb) determined from fluorescence with potential density contours (kg m$^{-3}$) and (B) dissolved oxygen (ml L$^{-1}$) with oxygen saturation (percentage) contours. For more details see Fig. 7.
Fig. 9. Increasing productivity in two EAC warm core eddies as a result of the shallowing of the surface mixed layer. In situ vertical profiles of temperature (black) and nitrate (red) from A. 1-3 October 1997, RV Franklin 199708 (Hitchman et al., 2000); B. Eddy J, 19-20 November 1978, RV Sprightly 197815 (Tranter et al., 1980); and C. Eddy J, 17-20 September 1978, RV Sprightly 197812 (Tranter et al., 1980). Symbols shown in Panel E refer to the location of temperature and nitrate profiles shown in Panels A, B and C. E. 8 day mean chl a concentration centred 21 November 1978 from the Coastal Zone Color Scanner (primarily a 18 November pass). SeaWiFS Level-3 daily chlorophyll on the 27 September (D) and 22 November 1997 (F). Arrows give 1 day Lagrangian geostrophic velocity paths for the 27 September and 20 November 1997 determined from the Topex/Poseidon and ERS2 altimeters.
Correction for non-photochemical quenching of fluorescence

Fig. 10. Chlorophyll estimates from uncorrected (grey) and corrected (black) fluorescence in the centre of the eddy on 1 December 2008 (main panel) and for the whole eddy at 20 m depth (insert). Correction of non-photochemical quenching is described in Appendix B. BB06 - (Behrenfeld and Boss, 2006).