Local winds and encroaching currents drive summertime subsurface blooms over a narrow shelf

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Abstract

The ability to forecast the biological productivity of the coastal ocean relies on the quantification of the physical processes that deliver nutrients to the euphotic zone. Here we explore these pathways using observations of the coupled biological and physical variability of waters offshore of the east coast of Tasmania in the summertime. The observations include an array of moored autonomous profilers deployed over an 18-d period—providing continuous, full-depth measurements of turbulent microstructure, temperature, velocity, and chlorophyll a (Chl a) fluorescence, complemented by shipboard nutrient measurements. Local upwelling was driven by the encroaching East Australian Current (EAC) extension onto the shelf and to a lesser extent the local winds. The interaction of the local winds and the encroaching boundary current was reflected in the shelf nutrient budget and led to a rapid increase in subsurface Chl a. Diffusive vertical fluxes had minimal impact on subsurface Chl a in the mid-shelf and outer-shelf. Upwelling-favorable winds were too weak to drive significant vertical mixing, and mixing associated with the current-driven Ekman transport was too deep compared to the euphotic zone depth. The observed subsurface Chl a did not reflect the satellite estimates of productivity. Since the EAC extension transports warm, low-nutrient surface waters from the sub-tropics, satellite chlorophyll measurements decreased during the same period the depth-averaged Chl a increased. This seeming paradox illustrated how long duration, full water column sampling can elucidate the coupled biological and physical processes that aid our ongoing effort to forecast the biological state of the coastal ocean.

Coastal waters are biologically productive with diverse ecosystems that support a myriad of human uses, including the planet’s largest fisheries (Simpson and Sharples 2012). This elevated productivity is largely due to the physical dynamics that deliver nutrients into the sunlit surface waters for uptake by phytoplankton. For example, the ocean’s largest rates of carbon-fixation per area occur during the intense blooms formed during coastal upwelling, common in mid-latitude eastern boundary current regimes such as the Californian shelf (Antoine et al. 1996).

Winds, mesoscale and sub-mesoscale dynamics, terrestrial run-off, and internal waves all contribute to the coastal ocean’s elevated primary productivity (Simpson and Sharples 2012). In some regions, the system is dominated by a single physical process (Woodson et al. 2007). In others, several dynamical processes can be occurring simultaneously, and at similar time- or length-scales, challenging our ability to accurately model and forecast the state of the biogeochemistry in the coastal ocean (Lucas et al. 2011a; Prairie et al. 2012). The complexity in the physically mediated conditions for phytoplankton growth is only amplified by the nonlinear response of the biology, which is concurrently constrained by the complex biogeochemical (e.g., nutrient availability, Omand et al. 2012) and ecosystem dynamics at play (e.g., grazing, Prairie et al. 2012).

Observing the degree to which various physical processes influence the nutrient budget on the continental shelf is an important first step toward improved coastal ocean ecosystem forecasting. The primary physical mechanisms for nutrient input into coastal ocean ecosystems are from either from...
terrestrial sources via run-off, the transport of deep, nutrient-rich waters into the euphotic zone via advection, or from diapycnal (approximately vertical) and along-isopycnal (approximately horizontal) diffusive fluxes (Simpson and Sharpley 2012). When terrestrial sources can be ignored, the input of “new” nutrients into the continental shelf euphotic zone can be controlled by advection, for example in wind-forced upwelling regions (Lucas et al. 2014), or by mixing and diffusion, for example at tidal fronts (Moore 2003) or areas with energetic internal waves (Sharpley et al. 2001; Lucas et al. 2011b). At certain locations and times, advective and diffusive fluxes can be tightly coupled (Beaird et al. 2020), and combine to result in high levels of primary productivity (Omand et al. 2012; Lucas et al. 2014; Pitcher et al. 2014; Fearon et al. 2020).

In the regions downstream and offshore of strong western boundary currents, elevated eddy kinetic energy is associated with vertical transport that injects nutrients into the euphotic zone (Pascual et al. 2015). Domino isopycnals within cold-core eddies supports elevated chlorophyll concentrations and primary productivity (Nencioli et al. 2008). At smaller spatial scales, filaments, meanders, and strong fronts associated with submesoscale, ageostrophic dynamics play an important role in the productivity, structure, and function of the open-ocean ecosystem (Lévy et al. 2012; Mahadevan 2016).

Near the coast, mesoscale currents can drive vertical transport. Current-driven upwelling occurs when poleward currents encroach onto the continental shelf along a western boundary (Fig. 1). Onshore transport at depth results from Ekman veering of the poleward flow within the frictional bottom boundary layer (Oke and Middleton 2000; Roughan et al. 2003; Roughan and Middleton 2004). Like wind-forced upwelling (Fig. 1a), cool and nutrient-rich waters are uplifted near the coast, with the potential to impact the local primary productivity (Schaeffer et al. 2014). However, as the cross-shelf gradients in density increase, a resulting buoyancy force opposes the onshore Ekman transport, eventually arresting the flow (Oke and Middleton 2000; Roughan et al. 2003). Vertical mixing between uplifted and interior waters (Oke and Middleton 2000), or cross-shelf density gradients resulting from baroclinicity of the encroaching current, can extend the length of time before the onshore flow is blocked to on the order of days, perhaps sufficient for the local phytoplankton population to respond (Fig. 1b, inset). In general, the impact of current-driven upwelling on continental shelf ecosystems is less well studied than its wind-driven counterpart.

At times, both current- and wind-driven upwelling can occur, leading to an especially strong upwelling response over the shelf (Fig. 1c, Leber et al. 2017). In what follows, we examine such a case offshore of the east coast of Tasmania, using a combination of mooring and shipboard observations. The measurements quantified the temporal and spatial variability of upwelling and turbulent mixing forced by the local winds, onshore encroachment of the western boundary current extension, and internal waves. We show the response of phytoplankton concentration to these co-occurring physical processes on the shelf.

**Study area and methodology**

**The Tasmanian eastern continental shelf**

The eastern continental shelf of Tasmania, located southeast of mainland Australia (Fig. 2a), is subject to strong winds, tidal and internal wave forcing, and the eddy-rich southern extension of the East Australian Current (EAC, Ridgway and Godfrey 1997; Ridgway 2007; Suthers et al. 2011; Oliver et al. 2016; Oke et al. 2019). The EAC is a western boundary current that flows southward along the east coast of Australia, separating at the Australian mainland into the eddy-rich southern and eastern extensions between 30°S and 32°S (Oke et al. 2019). Typically the current is positioned offshore of the shelf-break, resulting in horizontally-sheared along-shelf currents and large vorticity gradients (Schlosser et al. 2019b; Malan et al. 2020), which can inhibit the exchange of deep ocean and coastal waters (Malan et al. 2020). In the coastal waters of mainland Australia, wind-driven upwelling is localized, short-lived, and inconsistent (Roughan and Middleton 2002; Schaeffer et al. 2014). Upwelling and nutrient enrichment is dominated by the encroachment of the EAC onto the shelf (Schaeffer et al. 2014), where current-driven onshore Ekman transport tends to persist for extended periods and is relatively common (Roughan et al. 2003; Schaeffer et al. 2013, 2014; Schaeffer and Roughan 2015). Compared to other western boundary currents like the Gulf Stream and Kuroshio Current, the EAC southern extension is a transient current with weaker transport but similar eddy kinetic energy and is the least barotropic (i.e., most baroclinic, Rykova et al. 2017; Oke et al. 2019).

The EAC southern extension transports warm, saline, and nutrient-poor waters from the subtropics to the Tasman shelf in the summer, displacing the cold and nutrient-rich northward-flowing current that reaches a maximum latitudinal extent during winter months (Ridgway and Condie 2004; Oliver et al. 2016). The lack of silicate in EAC waters benefits dinoflagellates over diatoms, and diatoms are currently decreasing in abundance on the shelf (Thompson et al. 2009). Historically, the largest phytoplankton blooms over the shelf were observed in spring months, but the surface signature of this bloom has weakened by approximately 50% from 1997 to 2007 (Thompson et al. 2009).

The waters offshore of the east coast of Tasmania are also subject to strong surface forcing conditions, leading to a complex physical environment with highly variable biological states. It is thus a valuable laboratory for the study of the biological response to physical dynamics that interact across a range of scales. Strong local winds from passing atmospheric fronts can drive coastal upwelling or downwelling and set off energetic near-inertial motions (Schlosser et al. 2003; Roughan and Middleton 2004). Like wind-forced
et al. 2019b). With an average latitude of 42°S, the diurnal tide is sub-inertial on the 28 km wide shelf, leading to the presence of energetic diurnal coastal-trapped waves over the outer-shelf and shelf-break (Schlosser et al. 2019a). There is an energetic internal wave field, even though the incoming semidiurnal internal tide is predominantly reflected offshore by the steep continental slope (Klymak et al. 2016; Marques et al. 2021).

Poleward winds and the resulting coastal upwelling can drive primary production, as found from the Maria Island monitoring station that has been operating since 1944 (Thompson et al. 2009). Harris et al. (1988) hypothesized that wind mixing events can erode the pycnocline and replenish nutrients in the surface waters, although mixing was not directly observed. These “local” responses to wind forcing may be acting concurrently with changes in the state of the coastal ocean mediated by the large-scale behavior of the eddy-rich EAC and fluxes of nutrients due to near-inertial motions, internal tides, and internal waves. Separating these physical drivers of coastal ocean productivity is a difficult observational challenge.

Field experiment
In February 2015, we deployed five moorings on the 28 km wide shelf (Fig. 2), as previously described in Schlosser et al. (2019a,b). These moorings consisted of three traditional moorings and two autonomous profilers (Wirewalkers; Pinkel et al. 2011). All moorings measured temperature and all but the most inshore profiler measured velocity (due to an Acoustic Doppler Current Profiler [ADCP] failure; Table 1). We also present 18 d of microstructure temperature measurements collected by χ pods (Moum and Nash 2009; Perlin and Moum 2012), from which we compute vertical mixing over the full water column (~100 m) at the two profilers deployed at the mid- and outer-shelf, 10 km (W10) and 21 km (W21) offshore, respectively. Mixing estimates were collected simultaneously with measurements of temperature, salinity, currents, and in situ fluorescence (chlorophyll a [Chl a]), a proxy for Chl a concentration, via the onboard moored autonomous “Wirewalker” profilers (Rainville and Pinkel 2001; Pinkel et al. 2011). Here the Wirewalkers achieved 0.25 m vertical resolution from the near-bottom to the near-surface with a profile repeat rate of every 10 min or less.
Our naming convention includes a letter to designate the type of mooring (M for traditional mooring or W for profiler) and a number to designate its distance from the shoreline in kilometers. Moorings were deployed for 18 d from the 5 February 2015 to 24 February 2015 (yeardays 38–55). Over the majority of this period, clouds were persistent but patchy,

![Diagram](image)

**Fig. 2.** (a) Sitemap of mainland Australia and the island of Tasmania to the south, with Bass Strait separating the land masses and Maria Island positioned to the southeast of Tasmania. Average streamlines over the field campaign (blue, HYCOM), the region shown in panel (b) (black patch), and the limits of the ship location during the mooring deployment (red). (b) The mooring locations with bathymetry contoured every 25 m to 175 m (gray, thin) and from 200 m every 100 m (thick). (c–e) Cross-shelf sections of the mesoscale along-shore velocity, $v_M$, with mesoscale temperature, $T_M$, contoured (black, °C) on day (c) 39, (d) 46, and (e) 50.

**Table 1.** The position of profilers and moorings deployed in this experiment and their design and sampling (reproduced from Schlosser et al. 2019a). Traditional moorings are denoted with “M” and profilers are denoted with “W” and their distance from the shore in kilometers is included in their name. a = seabird 39, b = seabird 56, c = seabird 37, d = seabird 39 with pressure sensor.

| Mooring name and location | Total depth (m) | ADCP location (m ASB) | ADCP sampling (pings per ensemble, vertical bins) | Thermistor locations (m ASB) | Thermistor sampling |
|---------------------------|-----------------|-----------------------|-----------------------------------------------|----------------------------|---------------------|
| W10 (148.3995°E, 41.4134°S) | 80              | NA*                   | NA                                            | Entire water column, sampled every 0.25 m | 0.17 s per profile, 6 min between profiles |
| M17 (148.4884°E, 41.4005°S) | 105             | 3.36                  | 6 pings/10 s, 2 m bins                       | 1.86a, 3.17a, 6.5a, 9a, 11.5a, 14a, 16.5d, 19.5a, 25a, 30a, 35a, 40a, 45d, 55a, 65a, 75a, 85a | a, c, d = 10 s       |
| W21 (148.5357°E, 41.3854°S) | 114             | 0.2                   | 6 pings/10 s, 2 m bins                       | Entire water column, sampled every 0.25 m | 0.17 s per profile, 8 min between profiles |
| M25 (148.5760°E, 41.3916°S) | 115             | 3.07                  | 6 pings/10 s, 2 m bins                       | 1.86a, 2.86a, 4.5c, 6.5a, 9a, 14a, 16.5c, 19.5a, 25a, 35a, 45d, 55a, 85b, 95b | a, c, d = 10 s       |
| M29 (148.6318°E, 41.3839°S) | 184.8           | 4.28                  | 12 pings/15 s, 2 m bins                      | 2.4a, 4.0a, 10b, 15c, 20b, 30b, 40b, 55b, 70d, 85b, 100b, 120d, 140b, 160d | a, c, d = 10 s, b = 0.5 s |

*ADCP was deployed on a 30° angle and so data are unusable.
impacting the daily ocean color satellite observations so instead we present the 8-d average of the daily observed Visible Infrared Imaging Radiometer Suite (VIIRS) product with 4 km resolution (NASA 2017). We present surface velocities from the HYCOM and NCODA Global 1/12° Analysis with daily temporal resolution (Chassignet et al. 2007). To resolve a daily sea surface temperature (SST) unimpaired by clouds, we present the NOAA high-resolution (1/4°) estimate with daily temporal resolution (Reynolds et al. 2007). Throughout the manuscript, time is in GMT + 0, where the local time was GMT + 11. In addition to moorings, we collected biological data during 20 h of ship-based sampling from the R/V Revelle near the shelf-break with a conductivity, temperature, and depth (CTD) sensor and concurrent water sampling.

Fluorescence

The Wirewalker autonomous profilers measured Chl a fluorescence, a proxy for phytoplankton biomass, over > 90% of the water column. We did not calibrate our fluorescence instruments in the field and we instead report the relative fluorescence units by normalizing the raw intensities (in volts) by the average intensity to allow comparison between the profilers. Nonphotochemical fluorescence quenching was corrected using the empirical correction method from Todd et al. (2009) and changed the fluorescence observations in the upper water column by < 10%. In steady-state conditions with constant carbon-to-chlorophyll ratios, Chl a can be used as a proxy for primary productivity. Our primary productivity estimates throughout this manuscript are based on this simplification.

Water samples

The CTD rosette collected water samples eight times at four to six depths distributed over the water column near the shelf-break (M29) over 20 h (a total of 48 samples) on day 48. Samples were not collected at consistent depths but were collected over a depth range of 20–180 m and conditions were representative of the time series. We obtained dissolved inorganic nitrate and silicate concentrations by filtering 30 mL of water through 0.45 μm filters into sterile vials that remained frozen until analysis by the Commonwealth Scientific and Industrial Research Organisation laboratory after 2 years. Every water sample had one duplicate from the same Niskin bottle, which was used to evaluate the efficacy of concentrations. The detection limit was 0.02 μM and the reported accuracy was 1%.

In Schlosser et al. (2019b), we demonstrated that the observed water mass had a linear temperature-density relationship. From the nitrate concentrations (NO₃, μM) and the concurrently measured temperature (T, °C) from the CTD, we determined a good correlation \( r^2 = 0.96 \) and \( p \)-value < 1 × 10⁻⁵ with the linear \( T = -NO₃ \) relationship \( NO₃(z,t) = -1.9T(z,t) + 35.1 \) for temperatures below 17.5°C by least-squares fitting over this temperature range (see Supporting Information). We applied the same method for silicate concentrations (SiO₂) and temperature, resulting in \( SiO₂(z,t) = -0.67T(z,t) + 11.1 \) for temperatures below 17.5°C (\( r^2 = 0.96 \) and \( p \)-value < 1 × 10⁻⁵). The relationship between nitrate and density surfaces (i.e., isopycnals) is stable in most water bodies (Omand and Mahadevan 2013, 2015), but in turbulent regions with enhanced vertical diffusive nutrient fluxes, nitrate varies along isopycnals (Hales et al. 2005; Beaird et al. 2020). Nitrate can also be remineralized (i.e., the microbial conversion to nitrate) within the euphotic zone resulting in recycled rather than new production (Dore and Karl 1996), which we did not consider. Although these are major assumptions that must be examined specifically in future work, there is good evidence that this approach is effective to first order (Lucas et al. 2011a,b, 2014; Green et al. 2019).

Surface mixed layer

We estimated the surface mixed layer depth from the local stratification, which we found was dominated by temperature variations (Schlosser et al. 2019a). We converted temperature to density (ρ, kg m⁻³) using our linear relationship \( ρ(z,t) = 1028.7 - 0.167T(z,t) \) and then estimated the buoyancy frequency (\( N_s \), s⁻¹) from \( N(z,t)^2 = -\frac{2\rho_0}{\rho} \frac{\partial\rho(z,t)}{\partial z} \), where \( g \) is the gravity constant (ms⁻²) and \( \rho_0 \) is the average \( \rho \). To estimate the mixed layer depth, we vertically binned the temperature to 2-m bins and applied a 6-m and 1-h moving-average before estimating \( N \). We defined the surface mixed layer as the region with \( N < 0.004 \)s⁻¹, and then finally applied a moving-average over 2-h at W10 and 3.3-h at W21. Note this method of estimating the surface mixed layer depth was only possible for the autonomous profilers with their high vertical resolution in upper ocean stratification.

Turbulent mixing

We estimated mixing from high-resolution profiles of temperature variance measured by the \( χ \) pods (Moum and Nash 2009; Perlin and Moum 2012), an FP07 fast-response temperature sensor and three-component motion package, sampled at 100 Hz and deployed on the W10 and W21 profilers. The Wirewalker profiler ratcheted downward and rose freely with the thermistor mounted above the top of the profiler (Lucas et al. 2016). We used data from the \( χ \) pods only during the upward movement of the positively buoyant profiler, during which the median upward speed was 0.35 ms⁻¹ (5th and 95th percentile upward speed of, respectively). We estimated the dissipation of temperature variance, \( χ (K^2 s^{-1}) \), following the methods of Moum and Nash (2009) and Perlin and Moum (2012), where the required methodology for a profiling microstructure measurement were outlined by Perlin and 206 Moum (2012). We then followed Osborn and Cox (1972) to estimate the eddy diffusivity for heat, \( K_T (m^2 s^{-1}) \), via:

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\[ K_T(z, t) = \frac{1}{dT/dz} \left( \frac{\partial T}{\partial z} \right)^2, \]

where \( dT/dz \) is the background vertical gradient in temperature and \( \langle \cdot \rangle \) represents a 1-m vertical average (the same range over which the dissipation of temperature variance was computed). We only estimated \( K_T \) in regions of sufficient stratification, where \( dT/dz > 0.001 \text{Km}^{-1} \) and \( N > 0.004 \text{s}^{-1} \).

**Vertical fluxes**

We estimated the vertical diffusive flux of proxy nitrate \( (q_N, \text{g N m}^{-2} \text{s}^{-1}) \), per unit horizontal area, at the W10 and W21 profilers with \( \chi \) pods using:

\[ q_N(z, t) = -K_T(z, t) \frac{dc(z, t)}{dz} = -K_T(z, t) \left( \frac{d[1.9T(z, t) + 35.1]}{dz} \right), \]

where \( c \) is the nitrate concentration (g m\(^{-3}\)) derived from the temperature observations. We assume the flow is sufficiently turbulent that the diffusivity for the tracer is the same as the diffusivity of heat \( (K_T) \). To avoid erroneously large fluxes where nitrate reduces from 2 \( \mu \text{M L}^{-1} \) to zero, we do not estimate nitrate fluxes where \( T \geq 17.5^\circ \text{C} \).

For the computation of \( K_T \) relevant to \( q_N, \chi \) and \( dT/dz \) were averaged vertically in 2-m bins and in time (at 1-, 3-, or 6-h intervals) before computing the relevant \( K_T \). Estimates were only computed if > 50\% of data points were retained after applying the stratification criteria. We then recalculated \( dT/dz \) and \( K_T \) from the binned quantities, following the methodology of Moum and Nash (2009) and Perlin and Moum (2012). The resulting \( K_T \) distribution was log-normal but slightly skewed to the right with more spread in \( K_T \) larger than the median than below the median, but with a similar median and mean (<2\% difference).

We only show nitrate fluxes here; however, the shelf was also limited in silicate due to the presence of silicate-depleted EAC waters (Thompson et al. 2009). Our observed ratio of silicate to nitrogen (Si : N) was 0.28, as compared to the estimated composition ratio of marine diatoms of 1.05 and nanoplancton (<20 \( \mu \text{m} \)) of 0.80 ± 0.35 of Brzezinski (1985), where a lower ratio indicates lower requirements for silicate. Therefore, our observed ratio will favor phytoplankton species with lower silicate requirements, and primary productivity may have been dependent on the flux of silicate rather than nitrate.

**Data analysis**

To separate the low-frequency component of the currents and temperature, we applied a 40-h low-pass filter using the PL66 filter (Beardsley and Rosenfeld 1983). We label these low-pass filtered, mesoscale quantities with a subscript \( M \). Unless otherwise specified, we smoothed all quantities with a 1-h moving average in figures to improve readability. We calculated the depth-averaged along-shore vertical shear via \( S(t) = \frac{\langle \partial v(z, t) \rangle}{\partial z} >_H \), using 2-m vertical bins in the \( z \) direction and 120-s segments in time \( t \), where \( < \cdot >_H \) indicates an average over the full water column with a seabed depth \( H \). Note that the average orientation of isobaths over the continental shelf was approximately north/south and so \( u (v) \) was directed east/west (north/south).

To confirm a linear relationship between two signals, we apply cross-correlation analysis by calculating the Pearson cross-correlation coefficient. To assess significance we compare to the 95\% significance limit, \( L_{95\%} = 1.96/\sqrt{n} \), where \( n \) is the number of samples. We consider the correlation coefficient to be significant if it exceeds this \( L_{95\%} \) value. We also cross-correlate signals where one signal is lagged in time to another, and in these cases, we must consider the potential impact that auto-correlation has on our cross-correlation. If we cross-correlate two signals that are both significantly auto-correlated to itself when lagged in time, the resulting cross-correlation will indicate a spurious relationship (Dean and Dunsmuir 2016). To address this, we estimate the “corrected” significance limit, \( CL_{95\%} = L_{95\%} \times F \), where \( F = \frac{\sqrt{(1+ab)}}{(1-ab)} \) and \( a \) and \( b \) are the auto-correlation of the two compared signals at the cross-correlated time lag (Dean and Dunsmuir 2016). When we then estimate the Pearson cross-correlation coefficient, we compare the signals with length \( n \) minus the time lag.

**Results**

**Regional Sea surface variability**

In the late austral summer of 2015, the Tasman shelf had warming SST, increasing surface along-shore velocities, and generally decreasing surface Chl \( a \) concentrations (Fig. 3), as estimated from 8-d averaged satellite (NASA 2017) and HYCOM products (Chassignet et al. 2007). Throughout the field campaign, the HYCOM surface currents flowed southward from mainland Australia to Tasmania (Fig. 2a), and the temperature and salinity at the outer-shelf (W21) were consistent with waters offshore from the mainland (not shown). The HYCOM surface currents resolved two counter-rotating eddies located offshore of the shelf during the period of the experiment (Figs. 2a). The hindcast model indicated that the clockwise rotating eddy of the pair moved gradually southwest-ward throughout the field campaign, with the counter-clockwise eddy to the west progressively squeezed between the clockwise eddy and the shelf (Fig. 3a–c). The offshore current also moved on-shore, with HYCOM surface currents indicating intensifying southward currents on the shelf from yearday 38 onwards (Fig. 3a–c). As the current encroached, SST warmed over the shelf and sea surface Chl \( a \) concentrations generally decreased (Fig. 3d–f), particularly over the cross-shelf region where we deployed the moorings (water depths ranging from 80 to 185 m).

**Subsurface variability over the continental shelf**

At both the mid-shelf (W10) and outer-shelf (W21) profilers (Fig. 2; Table 1), the observed subsurface Chl \( a \) was
highly variable in terms of both time and depth (Fig. 4g,h). A subsurface Chl \(a\) maximum was persistent at both profilers but was always shallower at the mid-shelf than the outer-shelf. We observed several periods of increasing Chl \(a\) at both profilers. At times, Chl \(a\) increased simultaneously, but at other times Chl \(a\) increased at only the outer-shelf.

Multiple physical processes were active during the 18-d observational period, including strong, variable wind stress (>0.1Nm\(^{-2}\), Fig. 4a), energetic southerly baroclinic currents (>0.2ms\(^{-1}\), Fig. 4b), and a shoaling and evolving thermocline (Fig. 4i,j). For convenience, we approximate the thermocline, where vertical temperature gradients are largest, as the depth of the 16°C isotherm (black line, Fig. 4i,j). We explore the interplay of these dynamics and their impact on Chl \(a\) variability in the following sections.

Sub-inertial processes

Our mid-shelf (W10) observations of subsurface temperature were generally consistent with an idealized two-layer water column responding to up- and down-welling winds, with isotherms shoaling (deepening) following upwelling winds (downwelling, Fig. 5a,b). Further offshore, however, the low-frequency temperature and velocities at the outer-shelf and shelf-break moorings (W21 and M29, Fig. 2c–e) reflected the encroachment of the surface intensified (i.e., baroclinic) EAC extension rather than a direct response to local winds. Consistent with the satellite observations and HYCOM model, the encroaching current increased the poleward velocities over the shelf with the largest velocities found at the shelf-break (M29) before day 44 but at the outer-shelf (M25 and M29) for the next 6 d (Figs. 2c–e, 5c). Then, consistent with current-driven upwelling, near-bottom temperatures cooled following southward near-bottom velocities (Fig. 5e). The combination of increasing poleward and vertically sheared flow and near-bottom onshore transport yielded an increase in the isotherm cross-shelf tilt and decreased subsurface temperatures on the shelf (Figs. 2c–e, 5c–d).

Onshore winds preceded the current’s onshore movement by O(1 d), for example, winds were sustained towards shore from...
days 43 to 46 as the vertical shear in the along-shelf mesoscale velocity increased (Fig. 5a–c). The mesoscale flow was also oriented more onshore. Although these weak onshore winds might marginally increase along-shelf transport, the HYCOM simulations indicated a nearby eddy-dipole was concurrently moving southwest with one eddy moving onto the shelf (Figs. 2a, 3c). We did not observe the relationship between the regional wind forcing and the evolving offshore eddy field, but it was apparent that the change in the currents over the continental shelf was inconsistent with wind-forcing alone.

The along-shelf depth-averaged vertical shear ($S_M$) and cross-shelf density gradients varied together (Fig. 5d), suggesting the baroclinic current was in thermal wind balance with the background density field (Fig. 5d, Gill 1982). However, even though we observed large cross-shelf tilts in the 16°C isotherm across the shelf (e.g., 63 m on day 46 in Fig. 5b), the computed theoretical vertical shear underestimated the observed shear by around 50%. Excluding the days 43 to 46 period when winds were both relatively strong and directed onshore, which likely led to ageostrophic conditions, the thermal wind balance predicted the trend in outer-shelf (W21) vertical shear.

In general, upwelling winds and near-bottom currents co-occurred, so the upwelling response on the shelf was a net effect of the two processes. The correlation between the along-shelf wind direction and the thermocline depth was strongest at the mid-shelf (W10) and weakened with distance offshore as the influence of the baroclinic current increased (Fig. 5a,c). For example, during upwelling winds, the thermocline shoaled at the mid-shelf (W10, day 41), but remained at a similar depth at the outer-shelf (W21), and deepened at the shelf-break (M29, Fig. 5c,d). The combined influence of the wind forcing and the encroaching baroclinic current led to thermocline and nutricline distributions that varied in time and across the shelf, with direct implications for the availability of nutrients over the shelf.

Local Chl a response to advection

The shelf nutrient budget is a function of wind- and current-driven upwelling, and we can investigate the biological impact of these motions by analyzing the statistical link between nitrate and subsurface Chl a concentrations. We compared the depth-averaged Chl a, $\langle \text{Chl a}\rangle_H$, to the depth-integrated proxy nitrate concentration, $\int_0^H \text{NO}_3^- dz$, where $\text{NO}_3^-$ is the concentration of proxy nitrate (Fig. 6a,b). Following an influx of nitrate, Chl a should increase after some time lag, depending on the growth rate of the local phytoplankton species present. We estimated the lag between nitrate and Chl a
that resulted in the largest significant cross-correlation and smallest mean absolute error (MAE = \((\|\text{Chl}_{a\text{fit}} - \text{Chl}_{a\text{obs}}\|)\)), where \(\text{Chl}_{a\text{fit}}\) is the fitted Chl \(a\) estimate and \(\text{Chl}_{a\text{obs}}\) is the observed Chl \(a\).

At the mid-shelf (W10), the fit between nitrate and Chl \(a\) was well correlated and significant for a lag of 2 d \((r^2 = 0.83, \text{CI}_{95\%} = 0.46)\), indicating a local Chl \(a\) response 2 d after the nitracline shoaled across the shelf (Fig. 6c). Note that we excluded days when the water column was relatively well mixed \((N < 8 \times 10^{-3} \text{ s}^{-1})\), but the fit was significant regardless. However, at the outer-shelf (W21) and for the same time lag the linear fit was insignificant \((r^2 = 0.14, \text{Fig. 6d})\). The cross- and along-shelf advection of Chl \(a\) increased in magnitude with distance offshore as velocities also increased (Fig. 2c-e), which may explain the weaker correlation between nitrate and Chl \(a\) at the outer-shelf. Note that if we excluded days from the fit when cross-shelf velocities were larger \((|\nu_M| > 0.1 \text{ ms}^{-1})\) the fit largely improved but remained insignificant due to the remaining auto-correlation of the signals \((r^2 = 0.41, \text{CI}_{95\%} = 0.54, L_{95\%} = 0.35)\). However, without additional cross-shelf measurements of Chl \(a\), we cannot estimate the lateral extent of the bloom or better estimate the cross-shelf flux from cross-shelf gradients in Chl \(a\).

**Mixing and internal waves**

At the mid- and outer-shelf, moored profiling measurements of temperature microstructure showed enhanced turbulent diffusivities \((K_T)\) below the surface mixed layer and near
the seabed, with especially large rates estimated following intense winds (Figs. 4k,l, 7a,b). During these intense winds, regions of enhanced diffusivities extended into the thermocline at W10 (e.g., on day 43 in Fig. 7a). However, diffusivities near the thermocline at the outer-shelf were not related to wind events and were occasionally elevated during periods with weak winds (Fig. 7b).

Below the surface mixed layer, the diffusive vertical flux of nitrate \(\dot{q}_{\text{N}}\) was sporadic with enhanced fluxes (i.e., \(\dot{q}_{\text{N}} > 2 \times 10^{-7} \text{ g N m}^{-2} \text{ s}^{-1}\)) sustained for only short periods (\(<1\text{ d}\), Fig. 7c–f). At the 16°C isotherm, our proxy for the nitracline, we observed enhanced diffusive fluxes following intense downwelling winds (\(>0.2\text{ N m}^{-2}\)) when the mixed layer was deepening inshore (e.g., days 39, 43, and 54). At the outershelf, however, waters with enhanced turbulent diffusivities had near-zero local nitrate concentrations, and hence diffusive nitrate fluxes were smaller than those inshore where the wind-driven mixing penetrated to the estimated depth of the nitracline.

The periods of elevated diffusive nitrate fluxes were short-lived (\(<12\text{ h}\)), primarily due to the deepening of the nutricline following downwelling winds, wind- and current-forced downwelling of isotherms across the entire shelf, and a flattening of the nutricline that subsequently deepened the nutricline on the mid-shelf. Although turbulent diffusivities and diffusive nitrate fluxes were relatively small near the thermocline at the outer-shelf, following the intense winds on day 42 and the generation of near-inertial waves on the shelf (Schlosser et al. 2019b), we resolved a near-inertial periodicity to the fluxes (18.1 h) when near-inertial velocities were enhanced. This indicates our measurements were sensitive to internal wave-driven turbulence in the highly stratified thermocline. Tidal internal waves were also present but these waves had weakly vertically sheared velocities as compared to the near-inertial waves and thereby resulted in minimal mixing (Schlosser et al. 2019a).

**Relative contributions to the nutrient budget**

The estimated vertical diffusive flux of nitrate at the 16°C isotherm was only \(O(1 \times 10^{-7} \text{ g N m}^{-2} \text{ s}^{-1})\) as compared to the estimated advective flux in nitrate during upwelling winds of \(O(1 \times 10^{-5} \text{ g N m}^{-2} \text{ s}^{-1})\). This advective flux estimates the net change in nitrate and includes both vertical and horizontal advection. The relative influence of advection and diffusion can be quantified via the Peclet number, \(\text{Pe}\):

\[
\text{Pe} = \frac{w Z_{\text{eu}}}{K_T} \sim \frac{w \frac{dz}{dt}}{K_T \frac{d^2 c}{dz^2}}
\]

where a representative estimate of the vertical velocity \(w\) during upwelling is \(1 \times 10^{-4} \text{ m s}^{-1}\), the euphotic zone depth \(Z_{\text{eu}}\) was estimated at 40 m for the eastern Tasman shelf (Harris et al. 1987), and during wind-mixing events \(K_T\) reached a maximum of \(O(1 \times 10^{-4} \text{ m}^2 \text{ s}^{-1})\)(Fig. 7a,b). The resulting minimum \(\text{Pe}\) estimate was 40, so even when turbulent diffusivities were at maximum values, by this measure, mixing would need...
to be at least 10 times larger to match the magnitude of the advective contribution to the nutrient budget.

**Discussion and conclusions**

**Current-driven upwelling on the Tasman shelf**

The strong horizontal shear of western boundary currents inhibits the cross-shelf exchange of nutrient-rich offshore waters with nutrient-depleted coastal waters (Malan et al. 2020). Long-term changes in western boundary current strength have been correlated to variability in coastal ecosystem productivity. For example, inter-annual strengthening of the EAC extension over the region was associated with a 50% decrease in surface Chl $a$ between 1997 and 2007 (Thompson et al. 2009). EAC strengthening has also been linked to changes in the habitats, plankton abundance, and fishery production of the region (Harris et al. 1992; Ridgway 2007; Johnson et al. 2011).

The observations analyzed here emphasized the disconnect between the surface and subsurface variability on the narrow Tasman shelf. As sea surface temperatures warmed over time (Figs. 3a–c, 4c,d), the depth-averaged temperatures cooled (Fig. 4c,d). Concurrently, surface Chl $a$ concentrations decreased over time (Fig. 3d–f), while subsurface Chl $a$ fluorescence increased by twofold (Fig. 6a,b). The biophysical changes corresponded with intensifying baroclinic currents over the shelf, suggesting that the shore-ward movement of the EAC southern extension eddy field led to a shoaling thermocline and nutricline over the shelf, especially during concurrent periods of upwelling-favorable winds (Fig. 2c–e). This coupling yielded increasing depth-averaged Chl $a$ even as surface Chl $a$ dropped and SST increased.

Despite the complex biophysical variability inherent in this system, a simple linear relationship can be formed to relate the NOAA SST product (Reynolds et al. 2007), which was highly correlated to the observed in situ near-surface temperature, to both the mid-shelf depth of the thermocline and the depth-averaged Chl $a$ for the period sampled by the mooring measurements (Fig. 8). This empirical relationship showed a linear and positive relationship between SST and subsurface Chl $a$. Although the fit might not be expected to be generally valid for the region, it indicates that, at times, current-driven upwelling impacted subsurface Chl $a$ concentrations, and could be tracked by satellite observations of the sea surface.

In addition to the physical drivers of biological variability, the observations also suggested non-nutrient constraints on productivity. The mid-shelf Chl $a$ was twice as sensitive to changes in proxy nitrate concentration compared to the outer-shelf (Fig. 6c,d). Assuming nutrient uptake by...
phytoplankton was limited to the euphotic zone, the cross-shelf difference in the Chl $a$ sensitivity to proxy nitrate, combined with the deeper depths at which proxy nitrate was greater than zero at the outer-shelf, may suggest co-limitation by light and nutrients. There was likely ecological variability, such as variable compositions of phytoplankton populations or variability in grazing. Despite these uncertainties, the significant correlation between proxy nitrate concentrations and Chl $a$ at the mid-shelf showed that productivity was in general nutrient-limited, and that current-driven upwelling could increase the productivity of the Tasman Shelf.

The role of mixing

This study was augmented by the direct measurements of turbulent microstructure collected at the moorings, allowing for the separation of mixing-driven and advective fluxes. In other strongly forced shelf regions, the coupling of wind-driven upwelling and vertical mixing can be multiplicative, acting to amplify the nutrient transport into the euphotic zone by upwelling (Lucas et al. 2014; Beaird et al. 2020). In addition, Beaird et al. (2020) shows that wind-driven upwelling can cause baroclinic and symmetric instabilities that lead to irreversible mixing, and hence even larger nutrient transport and productivity. In general, diffusive fluxes contribute an estimated 5% to 50% of total nutrient transport during wind-driven upwelling (Hales et al. 2005; Omand et al. 2012; Doubell et al. 2018; Beaird et al. 2020). Here, our observations indicate that the advective fluxes dominated diffusive fluxes due to winds, currents, or internal waves, even during strong forcing conditions.

The relatively small diffusive fluxes we observed are likely an additional indication that current-driven upwelling dominated the upwelling signal. Vertical mixing also results from current-driven upwelling, but at the upper edge of the bottom boundary layer (Oke and Middleton 2000), rather than the base of the surface mixed layer, as would be expected from wind-driven upwelling (Fig. 1). We observed enhanced vertical mixing when near-bottom currents increased at the outer-shelf (e.g., day 54, Figs. 5e, 7b), but at 100 m depth, likely below the euphotic zone. The resulting mixing-driven nutrient fluxes would be expected to have little impact on primary productivity at these depths. Inshore, there may have been some change in Chl $a$ due to enhanced diffusive nitrate fluxes where the bottom boundary layer is near or within the euphotic zone, but we do not have observations of that region. At the mid- and outer-shelf, it appears that mixing associated with current-driven upwelling had minimal impact on the observed Chl $a$.

Conclusions

The observed summertime Chl $a$ on the Tasman shelf was dependent primarily on the cross-shelf position and velocities of the eddy-rich EAC extension, and secondarily on the local winds. By contrast, contributions of diffusive nitrate fluxes were small. While the nutrient-poor waters of the EAC extension suppressed surface Chl $a$ in the region, due to the combined influence of wind- and current-driven upwelling, subsurface Chl $a$ increased, disconnecting surface and subsurface Chl $a$ variability. Although we observed small vertical diffusive fluxes relative to the total nitrate fluxes, we predict the biophysical dependence of this system could at times be even more complex, with internal waves enhancing vertical mixing and nutrient fluxes, and contributing to increased Chl $a$. Our observations highlight how depth-resolved measurements of biophysical variability can be used to understand multiscale physical control of coastal ecosystem primary productivity in varying boundary current regimes.

Conflicts of interest

None declared.
Data availability statement
The Tasmania field data are available at https://doi.org/10.26182/tx7m-5076 (Jones 2022). The HYCOM output is publicly available at https://hycom.org (Chassignet et al. 2007). The VIIRS ocean color products are available from NASA’s Ocean Color database at https://oceancolor.gsfc.nasa.gov (NASA 2017). The NOAA Daily Optimum Interpolation Sea Surface Temperature is publicly available from https://www.psl.noaa.gov/data/gridded/data.noaa.oisst.v2.highres.html (Reynolds et al. 2007). Matlab codes used for the data analysis can be made available upon request to T. L. Schlosser.

References
Antoine, D., J.-M. André, and A. Morel. 1996. Oceanic primary production 2. Estimation at global scale from satellite (coastal zone color scanner) chlorophyll. Global Biogeochem. Cycl. 10: 57–60. doi:10.1029/2003GL016889
Beaird, N. L., E. L. Shroyer, L. W. Juranek, B. Hales, and M. A. Goñi. 2020. Nutrient-rich gravity current formed by upwelling in Barrow Canyon: High resolution observations. J. Geophys. Res. Ocean 125(7): 1–19. doi:10.1029/2020jc016160
Beardsley, R. C., and L. K. Rosenfeld. 1983. Introduction to the CODE-1 moored array and large-scale data report, p. 1–16. In L. K. Rosenfeld [ed.], Tech. Rep. WHOI-83-23. CODE Technical Report 21CODE-1: Moored array and large-scale data report. Woods Hole Oceanography Institute.
Brzezinski, M. A. 1985. The Si:C:N ratio of marine diatoms: Interspecific variability and the effect of some environmental variables. J. Phycol. 21: 347–357. doi:10.1111/j.0022-3646.1985.00347.x
Chassignet, E. P., H. E. Hurlburt, O. M. Smedstad, G. R. Halliwell, P. J. Hogan, A. J. Wallcraft, R. Baraille, and R. Bleck. 2007. The HYCOM (Hybrid Coordinate Ocean Model) data assimilative system. J. Mar. Syst. 65: 60–83. doi:10.1016/j.jmarsys.2005.09.016
Dean, R. T., and W. T. M. Dunsmuir. 2016. Dangers and uses of cross-correlation in analyzing time series in perception, performance, movement, and neuroscience: The importance of constructing transfer function autoregressive models. Behav. Res. Methods 48: 783–802. doi:10.3758/s13428-015-0611-2
Dore, J. E., and D. M. Karl. 1996. Nitrification in the euphotic zone as a source for nitrite, nitrate, and nitrous oxide at Station ALOHA. Limnol. Oceanogr. 41: 1619–1628. doi:10.4319/lo.1996.41.8.1619
Doubell, M. J., D. Spencer, P. D. van Ruth, C. Lemckert, and J. F. Middleton. 2018. Observations of vertical turbulent nitrate flux during summer in the Great Australian Bight. Deep Sea Res. Part II Top. Stud. Oceanogr. 157–158: 27–35. doi:10.1016/j.dsr2.2018.08.007
Fearon, G., S. Herbette, J. Veitch, G. Cambon, A. J. Lucas, F. Lemarié, and M. Vichi. 2020. Enhanced vertical mixing in coastal upwelling systems driven by diurnal-inertial resonance: Numerical experiments. J. Geophys. Res. Ocean 125: 1–23. doi:10.1029/2020JC016208
Gill, A. E. 1982. Atmosphere-ocean dynamics. Elsevier.
Hales, B., J. N. Moun, P. Covert, and A. Perlin. 2005. Irreversible nitrate fluxes due to turbulent mixing in a coastal upwelling system. J. Geophys. Res. C Ocean 110: 1–19. doi:10.1029/2004JC002685
Green, R. H., N. L. Jones, M. D. Rayson, R. J. Lowe, C. E. Bluteau, and G. N. Ivey. 2019. Nutrient fluxes into an isolated coral reef atoll by tidally driven internal bores. Limnol. Oceanogr. 64: 461–473. doi:10.1002/lio.11051
Harris, G., C. Nilsson, L. Clementson, and D. Thomas. 1987. The water masses of the east coast of Tasmania: Seasonal and interannual variability and the influence on phytoplankton biomass and productivity. Mar. Freshw. 38: 569–590. doi:10.1071/MF9870569
Harris, G. P., P. Davies, M. Nunez, and G. Meyers. 1988. Interannual variability in climate and fisheries in Tasmania. Nature 333: 754–757. doi:10.1038/333754a0
Harris, G. P., F. B. Griffiths, and L. A. Clementson. 1992. Climate and the fisheries off Tasmania—Interactions of physics, food chains and fish. Benguela Trophic Funct. 12: 585–597. doi:10.2989/02577619209504726
Johnson, C. R., and others. 2011. Climate change cascades: Shifts in oceanography, species’ ranges and subtidal marine community dynamics in eastern Tasmania. J. Exp. Mar. Bio. Ecol. 400: 17–32. doi:10.1016/j.jebem.2011.02.032
Jones, N. (Creator). 2022. Tasmanian Eastern Continental Shelf 2015. The University of Western Australia. TECSdata.(zip). doi:10.26182/tx7m-5076
Klymak, J. M., H. L. Simmons, D. Braznikov, S. Kelly, J. A. MacKinnon, M. H. Alford, R. Pinkel, and J. D. Nash. 2016. Reflection of linear internal tides from realistic topography: The Tasman continental slope. J. Phys. Oceanogr. 46: 3337–3347. doi:10.1175/JPO-D-16-0061.1
Leber, G. M., L. M. Beal, and S. Elipot. 2017. Wind and current forcing combine to drive strong upwelling in the Agulhas Current. J. Phys. Oceanogr. 47: 123–134. doi:10.1175/JPO-D-16-0079.1
Lévy, M., R. Ferrari, P. J. S. Franks, A. P. Martin, and P. Rivière. 2012. Bringing physics to life at the submesoscale. Geophys. Res. Lett. 39: 1–13. doi:10.1029/2012GL052756
Lucas, A. J., C. L. Dupont, V. Tai, J. L. Largier, B. Palenik, and P. J. S. Franks. 2011a. The green ribbon: Multiscale physical control of phytoplankton productivity and community structure over a narrow continental shelf. Limnol. Oceanogr. 56: 611–626. doi:10.4319/lo.2011.56.2.0611
Lucas, A. J., P. J. S. Franks, and C. L. Dupont. 2011b. Horizontal internal-tide fluxes support elevated phytoplankton productivity over the inner continental shelf. Limnol. Oceanogr. Fluids Environ. 1: 56–74. doi:10.1215/21573698-1258185

Schlosser et al.
Local Winds and Encroaching Currents
Lucas, A. J., G. C. Pitcher, T. A. Probyn, and R. M. Kudela. 2014. The influence of diurnal winds on phytoplankton dynamics in a coastal upwelling system off southwestern Africa. Deep Sea Res. Part II Top. Stud. Oceanogr. 101: 50–62. doi:10.1016/j.dsr2.2013.01.016

Lucas, A., and others. 2016. Adrift upon a salinity-stratified sea: A view of Upper-Ocean processes in the Bay of Bengal during the southwest monsoon. Oceanography 29: 134–145. doi:10.5670/oceanog.2016.46

Mahadevan, A. 2016. The impact of submesoscale physics on primary productivity of plankton. Ann. Rev. Mar. Sci. 8: 161–184. doi:10.1146/annurev-marine-010814-015912

Malan, N., and others. 2020. Eddy-driven cross-shelf transport in the East Australian Current separation zone. J. Geophys. Res. Ocean 125: 1–15. doi:10.1029/2019JC015613

Marques, O. B., and others. 2021. Internal tide structure and temporal variability on the reflective continental slope of Southeastern Tasmania. J. Phys. Oceanogr. 51: 611–631. doi:10.1175/JPO-D-20-0044.1

Moore, C. M., and others. 2003. Physical controls on phytoplankton physiology and production at a shelf sea front: A fast repetition-rate fluorometer based field study. Mar. Ecol. Prog. Ser. 259: 29–45. doi:10.3354/meps259029

Moum, J. N., and J. D. Nash. 2009. Mixing measurements on an Equatorial Ocean mooring. J. Atmos. Oceanic Tech. 26: 317–336. doi:10.1175/2008JTECHO167.1

NASA Ocean Biology Processing Group. 2017. VIIRS-SNPP Level 3 mapped chlorophyll data version R2018.0. NASA Ocean Biology DAAC; [accessed 2021 February 03]. Available from: doi:10.5067/NPP/VIIRS/L3M/CHL/2018

Nencioli, F., V. S. Kuwahara, T. D. Dickey, Y. M. Rii, and R. R. Bidigare. 2008. Physical dynamics and biological implications of a mesoscale eddy in the lee of Hawai‘i: Cyclone Opal observations during E-Flux III. Deep. Res. Part II Top. Stud. Oceanogr. 55: 1252–1274. doi:10.1016/j.dsr2.2008.02.003

Oke, P. R., and J. H. Middleton. 2000. Topographically induced upwelling off eastern Australia. J. Phys. Oceanogr. 30: 512–531. doi:10.1175/1520-0485(2000)030<0512:TIUOE>2.0.CO;2

Oke, P. R., and others. 2019. Revisiting the circulation of the East Australian Current: Its path, separation, and eddy field. Prog. Oceanogr. 176: 102139. doi:10.1016/j.pocean.2019.102139

Oliver, E. C. J., M. Herzfeld, and N. J. Holbrook. 2016. Modeling the shelf circulation off eastern Tasmania. Cont. Shelf Res. 130: 14–33. doi:10.1016/j.csr.2016.10.005

Omand, M. M., F. Feddersen, R. T. Guza, and P. J. S. Franks. 2012. Episodic vertical nutrient fluxes and nearshore phytoplankton blooms in Southern California. Limnol. Oceanogr. 57: 1673–1688. doi:10.4319/lo.2012.57.6.1673

Omand, M. M., and A. Mahadevan. 2013. Large-scale alignment of oceanic nitrate and density. J. Geophys. Res. Ocean. 118: 5322–5332. doi:10.1002/jgrc.20379

Omand, M. M., and A. Mahadevan. 2015. The shape of the oceanic nitracline. Biogeosciences 12: 3273–3287. doi:10.5194/bg-12-3273-2015

Osborn, T. R., and C. S. Cox. 1972. Oceanic fine structure. Geophys. Fluid Dyn. 3: 321–345. doi:10.1080/03091927208236085

Pascual, A., S. Ruiz, B. Buongiorno Nardelli, S. Guinehut, D. Ludicone, and J. Tintoré. 2015. Net primary production in the Gulf Stream sustained by quasi-geostrophic vertical exchanges. Geophys. Res. Lett. 42: 441–449. doi:10.1002/2014GL062569

Perlin, A., and J. N. Moum. 2012. Comparison of thermal variance dissipation rates from moored and profiling instruments at the equator. J. Atmos. Oceanic Tech. 29: 1347–1362. doi:10.1175/JTECH-D-12-00019.1

Pinkel, R., M. A. Goldin, J. A. Smith, O. M. Sun, A. A. Aja, M. N. Bui, and T. Hughen. 2011. The Wirewalker: A vertically profiling instrument carrier powered by ocean waves. J. Atmos. Oceanic Tech. 28: 426–435. doi:10.1175/2010JTECHO805.1

Pitcher, G. C., T. A. Probyn, A. du Randt, A. J. Lucas, S. Bernard, H. Evers-King, T. Lamont, and L. Hutchings. 2014. Dynamics of oxygen depletion in the nearshore of a coastal embayment of the southern Benguela upwelling system. J. Geophys. Res. Ocean. 119: 2183–2200. doi:10.1002/2013JC009443

Prairie, J. C., K. R. Sutherland, K. J. Nichols, and A. M. Kaltenberg. 2012. Biophysical interactions in the plankton: A cross-scale review. Limnol. Oceanogr. Fluids Environ. 2: 121–145. doi:10.1215/21573689-1964713

Rainville, L., and R. Pinkel. 2001. Wirewalker: An autonomous wave-powered vertical profiler. J. Atmos. Oceanic Tech. 18: 1048–1051.

Reynolds, R. W., T. M. Smith, C. Liu, D. B. Chelton, K. S. Casey, and M. G. Schlax. 2007. Daily high-resolution-blended analysis of sea surface temperature. J. Climate 20: 5473–5496.

Ridgway, K. R. 2007. Long-term trend and decadal variability of the downward penetration of the East Australian Current. Geophys. Res. Lett. 34: 1–5. doi:10.1029/2007GL030393

Ridgway, K. R., and S. A. Condle. 2004. The 5500-km-long boundary flow off western and southern Australia. J. Geophys. Res. C Ocean. 109: 1–18. doi:10.1029/2003JC001921

Ridgway, K. R., and J. S. Godfrey. 1997. Seasonal cycle of the East Australian Current. J. Geophys. Res. 102: 22921–22936. doi:10.1029/97JC00227

Roughan, M., and J. H. Middleton. 2002. A comparison of observed upwelling mechanisms off the east coast of Australia. Cont. Shelf Res. 22: 2551–2572. doi:10.1016/S0278-4343(02)00101-2

Roughan, M., P. Oke, and J. H. Middleton. 2003. A modeling study of the climatological current field and the trajectories of upwelled particles in the East Australian Current. J. Phys. Oceanogr. 33: 2551–2564. doi:10.1175/1520-0485(2003)033<2551:AMSOTC>2.0.CO;2
Roughan, M., and J. H. Middleton. 2004. On the East Australian Current: Variability, encroachment, and upwelling. J. Phys. Oceanogr. 109: C07003. doi: 10.1029/2003JC001833

Rykova, T., P. R. Oke, and D. A. Griffin. 2017. A comparison of the structure, properties, and water mass composition of quasi-isotropic eddies in western boundary currents in an eddy-resolving ocean model. Ocean Model. 114: 1–13. doi: 10.1016/j.ocemod.2017.03.013

Schaeffer, A., M. Roughan, and B. Morris. 2013. Cross-shelf dynamics in a Western Boundary Current Regime: Implications for upwelling. J. Phys. Oceanogr. 43: 1042–1059. doi: 10.1175/JPO-D-12-0177.1

Schaeffer, A., M. Roughan, and J. Wood. 2014. Observed bottom boundary layer transport and uplift on the continental shelf adjacent to a western boundary current. J. Geophys. Res. Ocean. 119: 4922–4939. doi: 10.1002/2013JC009735

Schaeffer, A., and M. Roughan. 2015. Influence of a western boundary current on shelf dynamics and upwelling from repeat glider deployments. Geophys. Res. Lett. 42: 121–128. doi: 10.1002/2014GL062260

Schlosser, T. L., N. L. Jones, R. C. Musgrave, C. E. Bluteau, G. N. Ivey, and A. J. Lucas. 2019a. Observations of diurnal coastal-trapped waves with a thermocline-intensified velocity field. J. Phys. Oceanogr. 49: 1973–1994. doi: 10.1175/JPO-D-18-0194.1

Schlosser, T. L., N. Jones, C. E. Bluteau, M. H. Alford, G. N. Ivey, and A. J. Lucas. 2019b. Generation and propagation of near-inertial waves in a Baroclinic Current on the Tasmanian shelf. J. Phys. Oceanogr. 49: 2653–2667. doi: 10.1175/JPO-D-18-0208.1

Sharples, J., C. M. Moore, and E. R. Abraham. 2001. Internal tide dissipation, mixing, and vertical nitrate flux at the shelf edge of NE New Zealand. J. Geophys. Res. Ocean 106: 69–81. doi: 10.1029/2000JC000604

Simpson, J. H., and J. Sharples. 2012. Introduction to the physical and biological oceanography of shelf seas. Cambridge Univ. Press.

Suthers, I. M., and others. 2011. The strengthening East Australian Current, its eddies and biological effects—An introduction and overview. Deep. Res. Part II Top. Stud. Oceanogr. 58: 538–546. doi: 10.1016/j.dsr2.2010.09.029

Thompson, P. A., M. E. Baird, T. Ingleton, and M. A. Doblin. 2009. Long-term changes in temperate Australian coastal waters: Implications for phytoplankton. Mar. Ecol. Prog. Ser. 394: 1–19. doi: 10.3354/meps08297

Todd, R. E., D. L. Rudnick, and R. E. Davis. 2009. Monitoring the greater San Pedro Bay region using autonomous underwater gliders during fall of 2006. J. Geophys. Res. Ocean 114: 1–13. doi: 10.1029/2008JC005086

Woodson, C. B., and others. 2007. Local diurnal upwelling driven by sea breezes in northern Monterey Bay. Cont. Shelf Res. 27: 2289–2302. doi: 10.1016/j.csr.2007.05.014

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