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Mesoscale eddies and the impact of coastal iron supply on primary production in the South Pacific Subtropical Front

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ABSTRACT

Subtropical and Subantarctic waters either side of the southern hemisphere Subtropical Front are considered iron-limited, suggesting production within the front is dependent on a supply of iron from atmospheric deposition, zonal advection of coastal water, or upwelling. We present the results from a one-day biogeochemical survey in Subtropical Water east of the North Island, New Zealand, in a region where mesoscale cyclonic and anticyclonic eddies entrain chlorophyll in filaments around the eddies. There was no significant relationship between upper-mixed layer chlorophyll and any physical or macronutrient quantity. However, chlorophyll was significantly positively correlated with dissolved iron. A simple model suggests that while vertical entrainment of iron into the upper mixed layer occurred, most of the dissolved iron in the eddy was due to entrainment of high-iron coastal water into low-iron offshore Subtropical Water, and that this iron supports primary production in otherwise iron-deficient water. We suggest that a significant component of the total primary production within the STF may be determined by mesoscale eddy induced lateral advection of iron.

Keywords: Iron supply; Neritic water; Subtropical water; Primary production; South-west Pacific Ocean; New Zealand

1. Introduction

Primary production at subtropical latitudes in the South Pacific Ocean is determined by the nutrient and physical characteristics of two main surface water masses in the region. To the south, Subantarctic Water (SAW) is recognised as high nitrate-low chlorophyll water (HNLC) where surface summer and winter nitrate concentrations are typically 7 and 15 mmol m^{-3}, respectively (Johnson et al., 2017), yet mean sea surface chlorophyll (SSC) is typically only 0.3 mg Chl m^{-3} (Banse and English, 1997). SAW is considered iron-limited (Abraham et al., 2000; Boyd et al., 1999; Boyd et al., 2000), and it is often concluded that overall production in SAW depends on the input of dissolved iron (dFe). To the north, Subtropical Water (STW) is regarded as being at least seasonally, if not year round, oligotrophic, where surface nitrate concentrations are typically <2.5 mmol m^{-3} year round (Chiswell et al., 2022; Ellwood et al., 2018). Although STW has not been as well studied as SAW, it is also likely to be seasonally depleted in dFe both in the western (Ellwood et al., 2008) and eastern (Blain et al., 2008) South Pacific Ocean.

These two water masses meet at the Subtropical Front (STF), which spans the globe near 40°S, and is a region of high primary production, where the mean SSC can exceed 0.75 mg Chl m^{-3} (Figure 1a). Since both STW and SAW are probably iron-limited, primary production in the STF is likely to be supported by an influx of dFe. Three commonly discussed candidates for iron enrichment in the ocean are atmospheric deposition (e.g. Jickells and Moore, 2015), zonal advection of high-iron neritic water (e.g. Boyd et al., 2012; Ellwood et al., 2014; Graham et al., 2015), and mesoscale-eddy driven vertical pumping (e.g. Uchida et al., 2020), and it seems that primary production along the STF must be controlled by some combination of these three mechanisms. Since mean SSC tends to show higher values closer to land (Figure 1a), the two most likely mechanisms would appear to be atmospheric deposition and zonal advection. However, there is no consensus on which mechanism dominates, and there is not yet a complete understanding of the role of iron in STF production.

The mean circulation east of the North Island is dominated by the warm-core anticyclonic Wairarapa Eddy, which recirculates STW from the northern edge of the STF so that there is a region of enhanced SSC extending from East Cape to the Chatham Islands (Figure 1b). Mean SSC is depressed at the centre of the eddy, and this depressed biomass has been related to the deeper mixed layer in the eddy centre limiting primary production (Bradford et al., 1982; Waite et al., 2007). However, this picture is only true in the mean, and at any given time, the region is dominated by a complex pattern of anticyclonic and cyclonic mesoscale eddies likely shed near the East Cape and propagating south (Chiswell, 2005).

During September-October 2012, a two-week cruise was made east of the North Island, New Zealand, to investigate the role of iron in the evolution of the spring bloom in STW. The spring bloom was not spatially heterogeneous, and a satellite image transmitted to the ship showed that SSC in the region was dominated by eddy mixing, with filaments of high SSC entrained around mesoscale eddies. As a result of this image, a one-day survey was designed to investigate an anticyclonic eddy and its impact on production.

This article presents the results of this survey, where we found no discernible differences in the water mass properties, mixed-layer depths, or macronutrient concentrations between regions of high and low SSC in and around the eddy. The only measurable differences were that dFe was higher in regions of high SSC and where the upper mixed layer was deeper. A simple model suggests that while vertical entrainment of iron into the upper mixed layer occurred, most of the dissolved iron signal was due to the eddy entraining high-iron coastal water into low-iron offshore Subtropical Water, and that near-surface chlorophyll was a response to this iron. The resulting implication is that a significant component of the total primary production within the STF is determined by mesoscale-eddy induced zonal advection of iron.

2. Data and Methods

The cruise was made to the study site east of the North Island, nominally at 180°E, 39°S (Figure 2) from mid-September until early October 2012 to study the 2012 spring bloom. Details of the cruise and the results, including the bloom development, are given in Chiswell et al. (2019).

The one-day survey track was based on a MODIS image of 30 September (Figure 3) showing an anticyclonic eddy centred near 180°E, 39°S (labelled A_1 in Figure 2). The survey track started near the centre of the eddy as suggested by the surface chlorophyll pattern, and was then made about 50 km to the north/north-west before returning to the eddy centre. Data from a shipboard ADCP were used to determine near-surface ocean velocity and used to guide the survey in locating the eddy centre.

Both vertical profiles from CTD casts, and near-surface underway data from a towed-fish and the ship's sea chest instrumentation were collected.

Twelve CTD casts were made, from 03:30 on 2 October until 04:30 on 3 October (New Zealand Standard Time), 4 casts (2 through 5) were made during daylight, one cast (6) was made in twilight while the rest were made during the night. The CTD casts were made using a standard SeaBird 911 and carousel water sampler. Temperature, salinity, transmissivity, photosynthetic active radiation (PAR), and fluorescence profiles were measured to at least 350 m depth. Water samples for chlorophyll extractions and nutrient analyses were collected on up-casts using 24 10-L Niskin bottles mounted on the CTD rosette.

Chlorophyll derived from the CTD fluorometer was compromised by non-photochemical quenching (e.g., Carberry et al., 2019) during the daylight casts, so the CTD fluorometer data are not used in the analysis. Instead, following Bishop (1999), particulate organic matter (POM) was taken to be negatively proportional to the beam attenuation coefficient, *c*, derived from the CTD transmissometer as $c = \ln(T_r)/r$, where T_r is the transmissivity, and *r* is the path length. With no local calibration between POM and *c*, POM was normalised to have a maximum value of 1.0. A comparison of POM and CTD-derived chlorophyll (Chl) from night-time casts (to avoid quenching issues) for depths 5> 30 m over the entire 2-week cruise indicated a not quite linear relationship ($r^2 = 0.89$, Figure 4g). To the extent that Chl is an indicator of biomass, this supports the assumption that POM can be taken as a proxy for phytoplankton biomass.

The one-percent light level for each daylight cast was computed by fitting an exponential decay with depth function, $I = I_{sfc} \exp(-k \times z)$ to each PAR profile, and computing the one-percent light level as $Z_{100} = \ln(0.01)k$.

Macronutrients from the CTD up-cast samples were determined using an automated micro-segmented flow analyser with digital detection (Pickmere, 1998). The standard error in the automated analyses for each nutrient was estimated as the standard deviation of samples taken over a 6-hour night-time period earlier during the two-week cruise when the ship was on station. These overall standard deviations were 0.390 , 0.064, 0.23 and 0.093 μmol L^{-1} for nitrate, phosphate, silicate and ammonium, respectively. All indicators suggested variations seen in nutrient profiles were outside the error for individual measurements.

Near-surface dFe was measured from sea water pumped from a towed 'trace-metal fish', and was determined using flow injection analysis (Floor et al., 2015; Obata et al., 2002). Further analytical details are described in Ellwood et al. (2015).

Continuous near-surface measurements of temperature, salinity and fluorescence were made with a thermosalinograph (Seabird 38 and 21 sensors) and fluorometer (Wetlab ECO triplet) in the ship's sea chest. The sea chest also contained an inline Fast Repetition Rate Fluorometer (Chelsea Instruments FASTtracka FRRF) that provided measurements every 2 minutes of near-surface minimum (F_o) and maximum (F_m) fluorescence in the dark-acclimated state.

There was almost certainly some NPQ in the FRRF fluorescence (F_o and F_m), however, two extracted chlorophyll measurements from water samples during the survey agree well with F_o (see Figure 8), and here, F_o is used as a proxy for near-surface phytoplankton biomass (e.g., Ellwood et al., 2015).

The Moderate Resolution Imaging Spectroradiometer (MODIS, Esaias et al., 1998), launched in 2002 provides satellite-derived estimates of sea surface chlorophyll. Data used here were downloaded on land in near real time from the Ocean Color website (https://oceancolor.gsfc.nasa.gov/) and transmitted to the ship once per day.

Daily estimates of surface currents were obtained from the AVISO (Archiving, Validation and Interpretation of Satellite Oceanographic data).

Least-squares regressions between various quantities were performed using Matlab's fitlm.m routine, which provides p-values, assuming that the number of degrees of freedom equals the data length minus 2. Linear regression slopes are considered significant if $p<0.05$.

3. Results

3.1 The eddy field off the North Island

The satellite-derived surface velocity and chlorophyll fields taken during the cruise on 2 October 2012, just after the spring bloom had started, show several mesoscale (~100 km length scales) eddies, filaments, and other features influencing the SSC field (Figure 2a). Instead of a single Wairarapa eddy, there were at least three distinct anticyclonic eddies (labelled A_1 to A_3) and two distinct cyclonic eddies (C_1, C_2) in the region of the mean Wairarapa eddy.

The Rossby number (vorticity/planetary vorticity, $R_o = \zeta / f$) in these eddies typically had a maximum amplitude of 0.15 (Figure 2b), confirming that they were mesoscale, and approximately geostrophic (Mahadevan, 2016). Histograms of R_o for the entire region are slightly skewed positive (positive R_o indicates cyclonic flow in the southern hemisphere), but they become more positively skewed for regions of high SSC (Figure 2c). Moderately high SSC (>1 mg Chl m^{-3}) mostly occurred when the absolute value of R_o was less than 0.1, with a preference for cyclonic eddies (Figure 2e). Since the centres of the eddies had R_o about 0.15, this demonstrates that elevated values of SSC typically occurred over the flanks of the eddies rather than at their centres.

Conventional theory would suggest that primary production is higher in cyclonic than anticyclonic eddies because mixed layers are shallower in cyclonic eddies (e.g. Doblin et al., 2016), but the satellite image supports the idea that it is the circulation at the edges of the eddies that is important, although not in a simple way.

3.2 Physics of Eddy A1

The MODIS image of 30 September (Figure 3) showed clear evidence of anticyclonic eddy A1 centred near 180°E, 39°S. Unfortunately, the considerable cloud cover (typical for New Zealand) meant the next clear image was not until 2 October, i.e. after the survey had been started. SSC had evolved into a quite different looking feature by 2 October, and little information can be gleaned on how this change evolved – whether, for example, it was due to the eddy fragmenting and disappearing or whether growth in SSC obscured the eddy.

All twelve CTD casts made in the survey had a shallow upper mixed layer between 15 and 40 m thick sitting over a near-isopycnal layer that extended to between 250 and 350 m depth (Figure 4). For all casts, except cast 7 (which was the most inshore station) this deep mixed layer had density, σ_t = 26.499±0.001 kg m^{-3} (there was a 0.055 °C and 0.01 range in temperature and salinity in this layer).

This near-isopycnal deep layer was almost certainly a remnant mixed layer from the previous winter mixing, whereas the upper mixed layer reflects the emergence of new stratification in the spring. Density differences between the upper and remnant mixed layers ranged between 0.07 and 0.17 kg m^{-3} (Figure 4c). Thus using the common 0.125 kg m^{-3} density difference criterion for mixed layer depth (e.g., Kara et al., 2000) would sometimes place the mixed layer as the upper mixed layer and sometimes as the remnant mixed layer. Here, the upper mixed layer depth (uMLD) was defined as the depth where the density exceeded the surface value by 0.05 kg m^{-3}, and the remnant mixed layer depth (rMLD) was defined as the depth where the density exceeded the 200m value by 0.05 kg m^{-3}.

The upper mixed layer was on average 0.36°C warmer and 0.017 fresher than the remnant mixed layer. Temperature in the upper mixed layer ranged from 13.58 to 13.93°C, and while the warmest temperature was from noon on 2 October, any effects of diurnal heating were aliased by spatial variations in the temperature field since the second-warmest temperature was from 03:30 on 2 October.

Gridded upper and remnant mixed layer depths are shown in Figure 5, along with the velocities at 10-m and 100-m derived from the ADCP. The 100-m velocities are from near the top of the remnant mixed layer, but are at the depth limit of the ADCP and are noisier than the 10-m velocities. Velocities from both levels show anticyclonic circulation. The remnant mixed layer was deeper at the centre of the eddy (Figure 5b), consistent with anticyclonic eddies, reflecting the fact that these eddies extend to near the sea floor (Chiswell, 2003). The uMLD was negatively but not significantly correlated with rMLD (slope = -0.1, r^2 = 0.24, p-value =0.1).

The ADCP velocities indicate the anticyclonic eddy was at least ~30 km in radius with speeds at 10 m depth being ~0.4 m s^{-1} at the northern edge of the survey (Figure 5). The vorticity ζ calculated from the 10-m velocities was 1.4×10^{-4} s^{-1} leading to a Rossby number $R_o = \zeta / f$ of -0.16, which puts this eddy into the mesoscale class (i.e. $|R_o| << 1$), thus, geostrophy approximately holds, the flow is two-dimensional and the magnitude of the vertical velocity is several orders of magnitude smaller than the horizontal velocities (Mahadevan, 2016).

A 3-D view (Figure 6) shows the eddy from a different perspective, and shows the relatively thin warm fresher upper mixed layer sitting over the deeper remnant mixed layer that is warmer and more saline near the centre of the eddy. The remnant mixed layer was deepest near Casts 2 and 10 with shallowest values near the edge of the survey (Casts 6 and 8), whereas the upper mixed layer was shallowest near Cast 2 and deepest near Casts 6 and 8.

3.3 Chlorophyll distribution within Eddy A1

Chlorophyll profiles from the CTD fluorometer (Figure 4d) show NPQ above about 30 m depth in casts made during the day. However, POM was well-mixed in the upper mixed layer in all casts (Figure 4e), likely due to wind-mixing from strong winds on 1 October (see Chiswell et al., 2019). This suggests that all biological quantities (i.e. phytoplankton, dFe, and nutrients) were well mixed in the upper mixed layer, and that near-surface values (in particular, the towed-fish and sea-chest data) well represent upper mixed layer values.

The mean one-percent light level during the survey was 60 m, indicating that all upper mixed layers were shallower than the euphotic zone. In every cast, POM was elevated in the upper mixed layer and then decayed with depth to reach values 0.05 to 0.09 in the remnant mixed layer. POM in the upper mixed layer (hereafter, POM_u) ranged from 0.57 to 1.0, and visually separates into two groups, which we term high- and low-POM groups, having mean POM_u values of 0.88 and 0.63, respectively (Figure 4e).

There is little correlation between POM_u and any physical variable. Inspection of Figure 4 shows there is no immediately obvious relationship between POM group and temperature, salinity, or density. The two warmest, freshest, and lightest upper mixed layers were high-POM, but so too was the coolest, most saline, and densest upper mixed layer. There is also no clear relationship between POM_u and uMLD, for example, two high-POM casts had the shallow uMLD, but other high-POM casts had deep uMLD.

Least-squares regressions between POM_u and upper mixed layer temperature or salinity return non-significant results with low correlation-squared ($r^2<0.1$, $p>.1$, not shown). There is also no statistically significant relationship between POM_u and either uMLD or rMLD ($r^2 \sim 0$, $p>0.1$, not shown).

Much of the reason for these low correlations is because (as is illustrated in the 3-D view of the eddy, Figure 6d), POM was high both near the centre of the eddy (casts 2 and 3 with high upper layer T and S, and shallow uMLD) as well as on the flanks of the eddy (casts 6 to 8 with low upper layer T and S, and deeper uMLD).

Figure 7 shows the macronutrient concentrations (NH_4, NO_3+NO_2, PO_4, SiO_4) as a function of depth, colour-coded by POM group. Unfortunately, the depth levels for the nutrient sampling were pre-determined before the survey, and the 30-m level falls near the base of the upper mixed layer in many casts (see Figure 4), which complicates interpretation (i.e. not all these 30-m nutrient values were unambiguously from the upper mixed layer). There is considerable scatter, but most macronutrients increased with depth. At 10 m depth (which is the only level consistently in the upper mixed layer), NO_3+NO_2 and PO_4 showed lower values for the casts that had high POM_u, which would be consistent with the uptake of these nutrients by phytoplankton, but SiO_4 and NH_4 show no evidence of uptake.

While Chl and POM_u from the CTD casts, and F_o from the FRRF are all proxies for phytoplankton biomass, F_o is the more useful proxy to compare with dFe because of its more frequent sampling, and it is worth demonstrating that F_o correlates well with the other measures of phytoplankton biomass. Surface chlorophyll from the night-time CTD casts and POM_u from all casts align well with F_o (r^2 values of 0.74 and 0.38, and p-values of 0.007 and 0.033, respectively, Figure 8a). Similarly, F_o and POM_u both compare well with SSC from the 2 October MODIS image (which was taken about the time cast 4 was made). There are some discrepancies, but except for the last two casts (by which time the satellite image was 12 to 16 hours old), high F_o and POM_u generally overlie regions of high SSC, and low F_o and POM_u overlie regions of low SSC (Figure 8b).

Figure 9a presents F_o along with dFe from the tow fish. The ship passed through a region of elevated phytoplankton biomass during the morning of 2 October, followed by low biomass in the afternoon and then elevated biomass during the night (these are the regions near the centre and on the flanks of the eddy seen in Figure 6). A least-square regression (with F_o values interpolated to the dFe sample times) suggests that dFe is positively correlated with F_o ($r^2 = 0.27$; $p = 0.006$, Figure 9d). That dFe is correlated with phytoplankton biomass is also illustrated when dFe is superimposed on satellite SSC where high values of dFe coincide with high SSC (Figure 8b).

Dissolved iron, dFe, is correlated with uMLD ($r^2 = 0.18$; $p = 0.03$, Figure 9b, e), although since the operational procedure was to deploy the trace-metal tow fish between the CTD casts, the dFe values were taken in slightly different locations, and there is the possibility that small-scale patchiness confounds this relationship. Because dFe increases with depth (Chiswell et al., 2019, Figure 4), this suggests that part of the dFe signal could be due to entrainment of higher-dFe water as the upper mixed level deepened.

The upper mixed layer depth and F_o were uncorrelated (r^2= 0.01; $p = 0.175$, not shown), so they can be used as independent variables in a linear predictor to estimate dFe,

$$\mathrm{dFe}_{est} = a \times \mathrm{uMLD} + b \times \mathrm{F}_0 + c\,.$$

A multiple least-squares linear regression using the data shown in Figure 9, returns $a = 0.003 \pm 0.0018$ nmol L^{-1} m^{-1} and $b = 0.37 \pm 0.14$ nmol L^{-1}, and $c = -0.18 \pm 0.10$ nmol L^{-1}, with an overall p-value of 0.005. A fit with uMLD and F_o de-meaned and normalised by their respective standard deviations, returns $a = 0.019 \pm 0.01$, $b = 0.025 \pm 0.1$, and $c = 0.17 \pm 0.1$, indicating that the contribution to iron uptake/loss by normalised fluorescence variations is about 1.3 times that of normalised mixed layer depth variations. Whether computed dimensionally or non-dimensionally, the fit, dFe_{est}, accounts for 36% of the variance in dFe (Figure 9f). The implications of this fit are discussed in detail later, but for now we simply comment that positive a is consistent with entrainment of high-dFe water from depth, whereas positive b is unexpected since one might expect uptake of iron to lead to negative b.

4. Discussion

The remnant mixed layer in eleven of the twelve casts had near-identical density (σ_t = 26.499±0.001 kg m^{-3}), suggesting that the deeper water within Eddy A_1 had a common origin. This eddy was likely formed near East Cape where one to two eddies are shed per year that propagate south-west, but are topographically blocked by the Chatham Rise so that in the mean, the circulation is as shown in Figure 1 (Chiswell, 2005). The CTD casts were made only to 500 m at most, so that the full water-column baroclinic structure of eddy A_1 cannot be determined from this survey. However, the mean Wairarapa Eddy extends to at least 1000 m depth (Roemmich and Sutton, 1998), and individual mesoscale eddies are as deep (Chiswell, 2003), hence the observation that the remnant mixed layer structure is consistent with anticyclonic rotation (i.e. deeper mixed layer in the centre of the eddy, Figure 5). The shallow upper mixed layer (mean depth = 30 m) had negligible impact on the circulation.

The upper mixed layer during the survey was on average 0.37°C warmer and 0.019 fresher than the remnant mixed layer. This temperature difference is similar to that seen earlier in the cruise as the ocean became warmer and began to stratify (Chiswell et al., 2019). The decrease in salinity in the upper mixed layer can be accounted for by local precipitation exceeding evaporation (0.01m of rain would lead to about 0.01 salinity decrease if mixed over a 30-m upper mixed layer). Thus, the T-S properties of the upper mixed layer were likely to have been heavily modified by recent heating and rainfall, and so cannot be used to determine if regions of high- and low- POM were of coastal vs offshore origin (or vice-versa).

In every cast, the upper mixed layer showed more POM than in the remnant mixed layer (Figure 4e), indicating more phytoplankton biomass in the upper mixed layer. A complete analysis of this production is beyond the scope of this article, however, higher phytoplankton biomass in the upper mixed layer is broadly consistent with the Onset of Stratification model for spring blooms (Chiswell, 2011), but the lack of correlation between POM_u or F_o with either depth or temperature of the upper mixed layer suggests that the amount of phytoplankton biomass in the upper mixed layer cannot be ascribed to the age (assuming that a cooler upper mixed layer is formed more recently than a warmer upper mixed layer), or depth of the upper mixed layer. In other words, it seems unlikely that regions of low-biomass had stratified more recently or were earlier in the spring bloom cycle than regions of high-biomass.

Apart from a suggestion of nitrate and phosphate uptake, there was no statistically significant relationship between phytoplankton biomass in the upper mixed layer and macronutrients, although this lack of relationship may well have been due more to the lack of vertical nutrient sampling.

There was also no statistically significant relationship between upper mixed layer chlorophyll or POM and the remnant mixed layer depth. At first glance, this may not be surprising since it is unlikely that a mixed layer some 200-300 m below the upper mixed layer has a role in determining the upper mixed layer production. However, it has been suggested (Uchida et al., 2020) that vertical velocities associated with mesoscale eddies can bring iron-rich water from depth into the mixed layer where it can be consumed by phytoplankton, and their model simulations show much higher dFe in eddy-rich areas. Crucially, however, within their eddies they found a negative relationship between dFe and phytoplankton biomass, which we did not see, so while we cannot exclude such a mechanism from playing a role, it seems unlikely that it is the dominant mechanism driving the production within the eddy field in our study.

The only significant relationships we could find here were between dFe and chlorophyll or POM, and between dFe and upper mixed layer depth, with a simple linear model accounting for 36% of the variance in dFe.

While dFe was not measured within the remnant mixed layer during this survey, three days earlier in the cruise it was about 0.35 nmol L^{-1} (see Figure 4 in Chiswell et al., 2019). About 18% of the near-surface dFe variance can be explained by deepening of the upper mixed layer, and to the extent that the uML deepens through vertical mixing, this suggests entrainment of high-dFe water from the remnant mixed layer into the upper mixed layer (Figure 9e).

About 27% of the variance in near-surface dFe is explained by increased F_o, (Figure 9d). This positive relationship between dFe and phytoplankton biomass is a little surprising because one would expect uptake during a bloom to lead to a negative relationship. When combined with the satellite images showing apparent lateral entrainment of neritic water, we suggest that the most likely explanation for this result is that the eddies transport high-dFe coastal water offshore. This conclusion is supported by results from the same region in 2008, where Ellwood et al. (2014) found that mixed layer Mn:Fe and Mn:Al ratios were elevated compared to crustal values, and based on these ratios and particle-backtracking simulations, suggested the most likely source of iron was continental water.

Even so, if the coastal water had uniform iron concentration, uptake would lead to a negative relationship between dFe and phytoplankton biomass. There are two potential reasons for a positive relationship. One is that our results are confounded by atmospheric deposition, the other is that coastal water has such a large range in dissolved iron that uptake is obscured by this variability.

If phytoplankton are iron-limited, patchy atmospheric deposition of iron could lead to patches of phytoplankton, and thus to a positive relationship between dFe and phytoplankton, although it is by no means certain that deposition leads to increased dFe since lithogenic particles associated with atmospheric deposition can remove soluble iron in the water column via scavenging (Tagliabue et al., 2017).

Atmospheric deposition of dust in the New Zealand region has long been ascribed to Australian dust storms (Boyd et al., 2004; Kidson, 1930; Mahowald et al., 2009), but there are not sufficient data to determine the temporal and spatial scales of the deposition in the region. Ellwood et al. (2018; their Figure 5) present a map of annual mean Fe deposition based on modelling of global dust distributions (Albani et al., 2014) that shows broad area of deposition extending east of Australia and covering New Zealand. This modelled deposition has little of the spatial structure of surface chlorophyll seen in Figure 1, in particular there is no evidence of high deposition along the STF, and since the STF is considered to be region of high dFe (Banse and English, 1997; Boyd et al., 2004) this argues that deposition is not the major source of iron to the STF. Such featureless deposition might be expected in the annual mean, but various attempts to investigate the role of atmospheric deposition from satellite imagery have not provided more precise information on the temporal and spatial scales of deposition at shorter time scales. Boyd et al. (2004) inferred oceanic supply of dFe from episodic increases in chlorophyll concentrations in SAW seen in ocean colour images between 1997 and 2001, and found no evidence that these events were mediated by atmospheric iron supply, although they also could not explain these events from lateral advection or vertical mixing, and did comment that dust storms during this time sent plumes over both STW and SAW. They concluded that more data are needed, including on rainfall patterns in relation to dust plume trajectories to distinguish wet from dry deposition.

With little to no measurements of atmospheric deposition of iron in the region, the role of deposition in this mesoscale eddy system must remain an open question.

This then raises the question of whether our implied finding that dFe is high and variable in coastal water is consistent with previous work. There is surprisingly little information on dFe in coastal North Island water, however, it is reasonable to assume that coastal water, in general is higher in dFe than offshore water. For example, Hutchins et al. (1998) found dFe in California coastal water to vary between <0.1 and >8.0 nmol L^{-1}, and explained this high variability was due to uneven distributions of sources of iron such as rivers and resuspension of shelf sediments. More locally, Sander et al. (2015) found dFe on average to be 1.0 ± 0.4 nmol L^{-1} in neritic water on the Otago shelf, compared to 0.2 ± 0.1 nmol L^{-1} offshore in SAW. Croot and Hunter (1998) reported dFe as high as 6 nmol L^{-1} on the Otago shelf, which they suggested was due to wind-induced upwelling rather than fluvial input. Upwelling occurs around the North Island east coast (Sharples and Greig, 1998) and may also be a significant source of dFe. On balance, it seems that high and variable dFe in coastal water is a reasonable finding.

Evidence that dFe is limiting in south-west Pacific Ocean STW is more scarce, perhaps because STW is generally considered oligotrophic in macronutrients and there has been relatively little interest in iron in these waters. However, Ellwood et al. (2008) found winter dFe levels ~0.1 nmol L^{-1} within and across the Subtropical Front in the Tasman Sea. Near 30°S, Ellwood et al. (2018) found surface dFe less than 0.2 nmol L^{-1} in the central Tasman Sea. Furthermore, phytoplankton located north of the Tasman Front (i.e. in STW, but to the west of our study region) were considered to be near the threshold for iron limitation (Ellwood et al., 2013).

In summary, the main results of this study are that the mesoscale eddies off the east coast of New Zealand are deep, with their circulation driven by the deep baroclinic structure, whereas (at least during the spring bloom) the surface production is largely constrained to a shallow upper mixed layer. From a biological perspective, the main role of the mesoscale circulation is to mix high-dFe neritic water with low-dFe offshore STW. Since extensive mixing of the two water masses extends several hundred km offshore, we conclude that a significant component of production within the STF is likely determined by mesoscale eddy induced zonal advection of iron. This conclusion supports those made by Graham et al. (2015), who suggested that bioavailable iron from the continental shelves is entrained into western boundary currents and then advected along the STF.

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References

Figures

Figure 1. Mean sea surface chlorophyll (SSC) from 2002 - 2016 derived from the Moderate Resolution Imaging Spectroradiometer (MODIS) on NASA's Aqua satellite. **a)** Mean SSC for the southern hemisphere; and **b)** Mean SSC for the New Zealand region. Black vectors are mean sea surface velocity derived from AVISO ocean altimeter data. Red lines are 0.5 and 0.75 mg Chl m^{-3} contour levels, dash-dot black line is 1000 m isobath, showing the Chatham Rise extending east of New Zealand. EC, WE, and CI indicate East Cape, Wairarapa Eddy, and Chatham Islands.

Figure 2. **a)** Sea surface chlorophyll (SSC) from 2 October 2012 derived from the Moderate Resolution Imaging Spectroradiometer (MODIS) on NASA's Aqua satellite. Vectors are surface currents derived from AVISO ocean altimeter data. Labels A, and C indicate anti-cyclonic and cyclonic eddies, numbered eddies are referred to in the text. The blue line near eddy A_1 indicates the cruise track during the one-day survey; **b)** Rossby number (vorticity divided by planetary vorticity, $R_0 = \zeta / f$) for 2 October 2012. Dashed contours indicate $R_0 = \pm 0.1$; and **c-f)** Histograms of Rossby number for regions where SSC exceeds various levels. Dash-dot line in **a** and **b** indicates the 250 m isobath, and data are only shown where the water depth is greater than 250 m to exclude the coastal zone.

Figure 3. Moderate Resolution Imaging Spectroradiometer (MODIS) images of sea surface chlorophyll (SSC) from 30 September to 4 October, 2012. **a)** 30 September; **b)** 2 October; **c)** 4 October; and **d)** enlarged view of 2 October, along with upper mixed-layer velocity from the ship's ADCP (red vectors). Numbered circles show the CTD cast locations from the one-day survey made in the anticyclonic eddy A1 shown in Figure 2. Colours of the circles in **d)** indicate whether the casts were considered high-POM (green) or low-POM (blue, see text). Black vectors are surface currents derived from AVISO satellite altimetry.

Figure 4. Profiles from the 12 CTD casts made during the one-day survey. **a)** Temperature (T); **b)** Salinity (S); **c)** Density (σ_t); **d)** Chlorophyll (Chl); **e)** Particulate organic matter (POM); **f)** Temperature vs. salinity (T-S); and **g)** POM vs Chl for the upper 5 to 30 m from CTD casts made over the entire cruise. Casts have been colour-coded by two groups, high-POM (green) and low-POM (blue). Horizontal blue line shows the mean one-percent light level (Z_{100}) from the daylight casts. Horizontal red dashed lines show the nutrient sample depths. Note the change of depth scale between upper and lower panels. Magenta lines in plots are from cast 7, which was closest inshore and had lowest temperature and salinity in the remnant mixed layer.

Figure 5. Gridded mixed-layer depths derived from CTD casts shown in Figure 4. **a)** Upper mixed layer depth (uMLD); and **b)** Remnant mixed layer depth (rMLD). Red and blue vectors are ADCP-derived velocities at 10 m and 100 depths, respectively, showing anticyclonic circulation.

Figure 6. Three-dimensional views of **a**) Tempearture (T); **b**) Salinity (S); **c**) Density (σ_t); and **d**) Particulate organic matter (POM) from the one-day CTD survey. Red vectors are ADCP-derived velocities at 10 m showing anticyclonic circulation. Vertical lines indicate upper mixed-layer depths and are colour-coded as high POM_u (green) and low POM_u (blue). Black and white dashed lines indicate the depths of the remnant and upper mixed layers, respectively. Numbers in **d** are the cast number. Note change of vertical (depth) scale between **a**, **b** and **c**, **d**.

Figure 7. Macronutrient profiles, from the CTD survey. **a)** Ammonium (NH_4); **b)** Nitrate plus nitrite (NO_3+NO_2); **c)** Phosphate (PO_4); and **d)** Silicate (SiO_4). Values have been colour-coded as high-POM (green) and low-POM (blue) profiles (see Figure 4). Errror bars are one standard deviation of analytical errors and do not include CTD sampling errors.

Figure 8. **a)** Time series of fluorescence (F_o) from Fast Repetition Rate Fluorometer (green line, and 1-hour smoothed, dark-green line), surface chlorophyll from night-time CTD casts (dark green squares), extracted chlorophyll from water samples at 10 m (red diamonds), and upper-mixed layer particulate organic matter (POM_u) from CTD casts (circles). POM_u has been colour-coded as high- POM_u (green) and low- POM_u (blue); and **b)** Scatter plots of F_o, POM_u, and dFe superimposed over sea surface chlorophyll (SSC) derived from Moderate Resolution Imaging Spectroradiometer for 2 October 2012 (see Figure 3). Note the colour scale for SSC is as shown in Figure 3. Numbers in circles indicate CTD cast number. Casts 2 to 5 were made during daylight.

Figure 9. Dissolved iron from underway tow fish (dFe), fluorescence from Fast Repetition Rate fluorometer (F_o), upper mixed layer depth from CTD casts (uMLD), and fit of dFe from linear model ($dFe_{est} = a \times uMLD + b \times F_0 + c$). **a)** dFe and F_o; **b)** dFe and uMLD; **c)** dFe and dFe_{est}; and **d-f)** linear regressions of F_o, uMLD, and dFe_{est} against dFe, where F_o and uMLD, has been interpolated to dFe sample times. All slopes are considered significant ($p<.05$). In **a)** and **b)** scales for F_o and uMLD are shown to the right of the axes.

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