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

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**ABSTRACT**

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

39

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

42

## 1. Introduction

43 Primary production at subtropical latitudes in the South Pacific Ocean is  
44 determined by the nutrient and physical characteristics of two main surface water  
45 masses in the region. To the south, Subantarctic Water (SAW) is recognised as high  
46 nitrate-low chlorophyll water (HNLC) where surface summer and winter nitrate  
47 concentrations are typically 7 and 15  $\text{mmol m}^{-3}$ , respectively (Johnson et al., 2017),  
48 yet mean sea surface chlorophyll (SSC) is typically only 0.3  $\text{mg Chl m}^{-3}$  (Banse and  
49 English, 1997). SAW is considered iron-limited (Abraham et al., 2000; Boyd et al.,  
50 1999; Boyd et al., 2000), and it is often concluded that overall production in SAW  
51 depends on the input of dissolved iron (dFe). To the north, Subtropical Water (STW)  
52 is regarded as being at least seasonally, if not year round, oligotrophic, where surface  
53 nitrate concentrations are typically  $<2.5 \text{ mmol m}^{-3}$  year round (Chiswell et al., 2022;  
54 Ellwood et al., 2018). Although STW has not been as well studied as SAW, it is also  
55 likely to be seasonally depleted in dFe both in the western (Ellwood et al., 2008) and  
56 eastern (Blain et al., 2008) South Pacific Ocean.

57 These two water masses meet at the Subtropical Front (STF), which spans the  
58 globe near  $40^{\circ}\text{S}$ , and is a region of high primary production, where the mean SSC can  
59 exceed 0.75  $\text{mg Chl m}^{-3}$  (Figure 1a). Since both STW and SAW are probably iron-  
60 limited, primary production in the STF is likely to be supported by an influx of dFe.  
61 Three commonly discussed candidates for iron enrichment in the ocean are  
62 atmospheric deposition (e.g. Jickells and Moore, 2015), zonal advection of high-iron  
63 neritic water (e.g. Boyd et al., 2012; Ellwood et al., 2014; Graham et al., 2015), and  
64 mesoscale-eddy driven vertical pumping (e.g. Uchida et al., 2020), and it seems that  
65 primary production along the STF must be controlled by some combination of these  
66 three mechanisms. Since mean SSC tends to show higher values closer to land (Figure  
67 1a), the two most likely mechanisms would appear to be atmospheric deposition and  
68 zonal advection. However, there is no consensus on which mechanism dominates, and  
69 there is not yet a complete understanding of the role of iron in STF production.

70 The mean circulation east of the North Island is dominated by the warm-core  
71 anticyclonic Wairarapa Eddy, which recirculates STW from the northern edge of the  
72 STF so that there is a region of enhanced SSC extending from East Cape to the  
73 Chatham Islands (Figure 1b). Mean SSC is depressed at the centre of the eddy, and  
74 this depressed biomass has been related to the deeper mixed layer in the eddy centre

75 limiting primary production (Bradford et al., 1982; Waite et al., 2007). However, this  
76 picture is only true in the mean, and at any given time, the region is dominated by a  
77 complex pattern of anticyclonic and cyclonic mesoscale eddies likely shed near the  
78 East Cape and propagating south (Chiswell, 2005).

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

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

## 2. Data and Methods

96 The cruise was made to the study site east of the North Island, nominally at  
97 180°E, 39°S (Figure 2) from mid-September until early October 2012 to study the  
98 2012 spring bloom. Details of the cruise and the results, including the bloom  
99 development, are given in Chiswell et al. (2019).

100 The one-day survey track was based on a MODIS image of 30 September  
101 (Figure 3) showing an anticyclonic eddy centred near 180°E, 39°S (labelled A<sub>1</sub> in  
102 Figure 2). The survey track started near the centre of the eddy as suggested by the  
103 surface chlorophyll pattern, and was then made about 50 km to the north/north-west  
104 before returning to the eddy centre. Data from a shipboard ADCP were used to

105 determine near-surface ocean velocity and used to guide the survey in locating the  
106 eddy centre.

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

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

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

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

132 Macronutrients from the CTD up-cast samples were determined using an  
133 automated micro-segmented flow analyser with digital detection (Pickmere, 1998).  
134 The standard error in the automated analyses for each nutrient was estimated as the  
135 standard deviation of samples taken over a 6-hour night-time period earlier during the

136 two-week cruise when the ship was on station. These overall standard deviations were  
137 0.390 , 0.064, 0.23 and 0.093  $\mu\text{mol L}^{-1}$  for nitrate, phosphate, silicate and ammonium,  
138 respectively. All indicators suggested variations seen in nutrient profiles were outside  
139 the error for individual measurements.

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

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

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

153 The Moderate Resolution Imaging Spectroradiometer (MODIS, Esaias et al.,  
154 1998), launched in 2002 provides satellite-derived estimates of sea surface  
155 chlorophyll. Data used here were downloaded on land in near real time from the  
156 Ocean Color website (<https://oceancolor.gsfc.nasa.gov/>) and transmitted to the ship  
157 once per day.

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

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

### 3. Results

#### 164 3.1 The eddy field off the North Island

165 The satellite-derived surface velocity and chlorophyll fields taken during the  
166 cruise on 2 October 2012, just after the spring bloom had started, show several  
167 mesoscale (~100 km length scales) eddies, filaments, and other features influencing  
168 the SSC field (Figure 2a). Instead of a single Wairarapa eddy, there were at least three  
169 distinct anticyclonic eddies (labelled A<sub>1</sub> to A<sub>3</sub>) and two distinct cyclonic eddies (C<sub>1</sub>,  
170 C<sub>2</sub>) in the region of the mean Wairarapa eddy.

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

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

#### 185 3.2 Physics of Eddy A1

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



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

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

208 The upper mixed layer was on average 0.36°C warmer and 0.017 fresher than  
 209 the remnant mixed layer. Temperature in the upper mixed layer ranged from 13.58 to  
 210 13.93°C, and while the warmest temperature was from noon on 2 October, any effects  
 211 of diurnal heating were aliased by spatial variations in the temperature field since the  
 212 second-warmest temperature was from 03:30 on 2 October.

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

221 The ADCP velocities indicate the anticyclonic eddy was at least ~30 km in  
 222 radius with speeds at 10 m depth being  $\sim 0.4 \text{ m s}^{-1}$  at the northern edge of the survey  
 223 (Figure 5). The vorticity  $\zeta$  calculated from the 10-m velocities was  $1.4 \times 10^{-4} \text{ s}^{-1}$   
 224 leading to a Rossby number  $R_o = \zeta / f$  of -0.16, which puts this eddy into the

225 mesoscale class (i.e.  $|R_o| \ll 1$ ), thus, geostrophy approximately holds, the flow is two-  
226 dimensional and the magnitude of the vertical velocity is several orders of magnitude  
227 smaller than the horizontal velocities (Mahadevan, 2016).

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

### 234 3.3 Chlorophyll distribution within Eddy A1

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

242 The mean one-percent light level during the survey was 60 m, indicating that all  
243 upper mixed layers were shallower than the euphotic zone. In every cast, POM was  
244 elevated in the upper mixed layer and then decayed with depth to reach values 0.05 to  
245 0.09 in the remnant mixed layer. POM in the upper mixed layer (hereafter,  $POM_u$ )  
246 ranged from 0.57 to 1.0, and visually separates into two groups, which we term high-  
247 and low-POM groups, having mean  $POM_u$  values of 0.88 and 0.63, respectively  
248 (Figure 4e).

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

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

260 Much of the reason for these low correlations is because (as is illustrated in the  
261 3-D view of the eddy, Figure 6d), POM was high both near the centre of the eddy  
262 (casts 2 and 3 with high upper layer T and S, and shallow uMLD) as well as on the  
263 flanks of the eddy (casts 6 to 8 with low upper layer T and S, and deeper uMLD).

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

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

285 Figure 9a presents  $F_o$  along with dFe from the tow fish. The ship passed through  
286 a region of elevated phytoplankton biomass during the morning of 2 October,  
287 followed by low biomass in the afternoon and then elevated biomass during the night

288 (these are the regions near the centre and on the flanks of the eddy seen in Figure 6).  
 289 A least-square regression (with  $F_0$  values interpolated to the dFe sample times)  
 290 suggests that dFe is positively correlated with  $F_0$  ( $r^2 = 0.27$ ;  $p = 0.006$ , Figure 9d).  
 291 That dFe is correlated with phytoplankton biomass is also illustrated when dFe is  
 292 superimposed on satellite SSC where high values of dFe coincide with high SSC  
 293 (Figure 8b).

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

301 The upper mixed layer depth and  $F_0$  were uncorrelated ( $r^2 = 0.01$ ;  $p = 0.175$ , not  
 302 shown), so they can be used as independent variables in a linear predictor to estimate  
 303 dFe,

$$304 \quad dFe_{est} = a \times uMLD + b \times F_0 + c .$$

305 A multiple least-squares linear regression using the data shown in Figure 9, returns  $a$   
 306  $= 0.003 \pm 0.0018 \text{ nmol L}^{-1} \text{ m}^{-1}$  and  $b = 0.37 \pm 0.14 \text{ nmol L}^{-1}$ , and  $c = -0.18 \pm 0.10 \text{ nmol}$   
 307  $\text{L}^{-1}$ , with an overall p-value of 0.005. A fit with uMLD and  $F_0$  de-meaned and  
 308 normalised by their respective standard deviations, returns  $a = 0.019 \pm 0.01$ ,  $b = 0.$   
 309  $025 \pm 0.1$ , and  $c = 0.17 \pm 0.1$ , indicating that the contribution to iron uptake/loss by  
 310 normalised fluorescence variations is about 1.3 times that of normalised mixed layer  
 311 depth variations. Whether computed dimensionally or non-dimensionally, the fit,  
 312  $dFe_{est}$ , accounts for 36% of the variance in dFe (Figure 9f). The implications of this  
 313 fit are discussed in detail later, but for now we simply comment that positive  $a$  is  
 314 consistent with entrainment of high-dFe water from depth, whereas positive  $b$  is  
 315 unexpected since one might expect uptake of iron to lead to negative  $b$ .

#### 4. Discussion

316 The remnant mixed layer in eleven of the twelve casts had near-identical density  
 317 ( $\sigma_t = 26.499 \pm 0.001 \text{ kg m}^{-3}$ ), suggesting that the deeper water within Eddy A<sub>1</sub> had a

318 common origin. This eddy was likely formed near East Cape where one to two eddies  
319 are shed per year that propagate south-west, but are topographically blocked by the  
320 Chatham Rise so that in the mean, the circulation is as shown in Figure 1 (Chiswell,  
321 2005). The CTD casts were made only to 500 m at most, so that the full water-column  
322 baroclinic structure of eddy A<sub>1</sub> cannot be determined from this survey. However, the  
323 mean Wairarapa Eddy extends to at least 1000 m depth (Roemmich and Sutton,  
324 1998), and individual mesoscale eddies are as deep (Chiswell, 2003), hence the  
325 observation that the remnant mixed layer structure is consistent with anticyclonic  
326 rotation (i.e. deeper mixed layer in the centre of the eddy, Figure 5). The shallow  
327 upper mixed layer (mean depth = 30 m) had negligible impact on the circulation.

328         The upper mixed layer during the survey was on average 0.37°C warmer and  
329 0.019 fresher than the remnant mixed layer. This temperature difference is similar to  
330 that seen earlier in the cruise as the ocean became warmer and began to stratify  
331 (Chiswell et al., 2019). The decrease in salinity in the upper mixed layer can be  
332 accounted for by local precipitation exceeding evaporation (0.01m of rain would lead  
333 to about 0.01 salinity decrease if mixed over a 30-m upper mixed layer). Thus, the T-  
334 S properties of the upper mixed layer were likely to have been heavily modified by  
335 recent heating and rainfall, and so cannot be used to determine if regions of high- and  
336 low- POM were of coastal vs offshore origin (or vice-versa).

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

348         Apart from a suggestion of nitrate and phosphate uptake, there was no  
349 statistically significant relationship between phytoplankton biomass in the upper

350 mixed layer and macronutrients, although this lack of relationship may well have been  
351 due more to the lack of vertical nutrient sampling.

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

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

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

373         About 27% of the variance in near-surface dFe is explained by increased F<sub>o</sub>,  
374 (Figure 9d). This positive relationship between dFe and phytoplankton biomass is a  
375 little surprising because one would expect uptake during a bloom to lead to a negative  
376 relationship. When combined with the satellite images showing apparent lateral  
377 entrainment of neritic water, we suggest that the most likely explanation for this result  
378 is that the eddies transport high-dFe coastal water offshore. This conclusion is  
379 supported by results from the same region in 2008, where Ellwood et al. (2014) found  
380 that mixed layer Mn:Fe and Mn:Al ratios were elevated compared to crustal values,

381 and based on these ratios and particle-backtracking simulations, suggested the most  
382 likely source of iron was continental water.

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

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

393 Atmospheric deposition of dust in the New Zealand region has long been  
394 ascribed to Australian dust storms (Boyd et al., 2004; Kidson, 1930; Mahowald et al.,  
395 2009), but there are not sufficient data to determine the temporal and spatial scales of  
396 the deposition in the region. Ellwood et al. (2018; their Figure 5) present a map of  
397 annual mean Fe deposition based on modelling of global dust distributions (Albani et  
398 al., 2014) that shows broad area of deposition extending east of Australia and  
399 covering New Zealand. This modelled deposition has little of the spatial structure of  
400 surface chlorophyll seen in Figure 1, in particular there is no evidence of high  
401 deposition along the STF, and since the STF is considered to be region of high dFe  
402 (Banse and English, 1997; Boyd et al., 2004) this argues that deposition is not the  
403 major source of iron to the STF. Such featureless deposition might be expected in the  
404 annual mean, but various attempts to investigate the role of atmospheric deposition  
405 from satellite imagery have not provided more precise information on the temporal  
406 and spatial scales of deposition at shorter time scales. Boyd et al. (2004) inferred  
407 oceanic supply of dFe from episodic increases in chlorophyll concentrations in SAW  
408 seen in ocean colour images between 1997 and 2001, and found no evidence that  
409 these events were mediated by atmospheric iron supply, although they also could not  
410 explain these events from lateral advection or vertical mixing, and did comment that  
411 dust storms during this time sent plumes over both STW and SAW. They concluded  
412 that more data are needed, including on rainfall patterns in relation to dust plume  
413 trajectories to distinguish wet from dry deposition.

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

416 This then raises the question of whether our implied finding that dFe is high and  
417 variable in coastal water is consistent with previous work. There is surprisingly little  
418 information on dFe in coastal North Island water, however, it is reasonable to assume  
419 that coastal water, in general is higher in dFe than offshore water. For example,  
420 Hutchins et al. (1998) found dFe in California coastal water to vary between  $<0.1$  and  
421  $>8.0$   $\text{nmol L}^{-1}$ , and explained this high variability was due to uneven distributions of  
422 sources of iron such as rivers and resuspension of shelf sediments. More locally,  
423 Sander et al. (2015) found dFe on average to be  $1.0 \pm 0.4$   $\text{nmol L}^{-1}$  in neritic water on  
424 the Otago shelf, compared to  $0.2 \pm 0.1$   $\text{nmol L}^{-1}$  offshore in SAW. Croot and Hunter  
425 (1998) reported dFe as high as  $6$   $\text{nmol L}^{-1}$  on the Otago shelf, which they suggested  
426 was due to wind-induced upwelling rather than fluvial input. Upwelling occurs around  
427 the North Island east coast (Sharples and Greig, 1998) and may also be a significant  
428 source of dFe. On balance, it seems that high and variable dFe in coastal water is a  
429 reasonable finding.

430 Evidence that dFe is limiting in south-west Pacific Ocean STW is more scarce,  
431 perhaps because STW is generally considered oligotrophic in macronutrients and  
432 there has been relatively little interest in iron in these waters. However, Ellwood et al.  
433 (2008) found winter dFe levels  $\sim 0.1$   $\text{nmol L}^{-1}$  within and across the Subtropical Front  
434 in the Tasman Sea. Near  $30^\circ\text{S}$ , Ellwood et al. (2018) found surface dFe less than  $0.2$   
435  $\text{nmol L}^{-1}$  in the central Tasman Sea. Furthermore, phytoplankton located north of the  
436 Tasman Front (i.e. in STW, but to the west of our study region) were considered to be  
437 near the threshold for iron limitation (Ellwood et al., 2013).

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



447 bioavailable iron from the continental shelves is entrained into western boundary  
448 currents and then advected along the STF.

449

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**References**

464

## Figures

465

466 Figure 1. Mean sea surface chlorophyll (SSC) from 2002 - 2016 derived from the  
 467 Moderate Resolution Imaging Spectroradiometer (MODIS) on NASA's  
 468 Aqua satellite. **a)** Mean SSC for the southern hemisphere; and **b)** Mean  
 469 SSC for the New Zealand region. Black vectors are mean sea surface  
 470 velocity derived from AVISO ocean altimeter data. Red lines are 0.5 and  
 471 0.75 mg Chl m<sup>-3</sup> contour levels, dash-dot black line is 1000 m isobath,  
 472 showing the Chatham Rise extending east of New Zealand. EC, WE, and  
 473 CI indicate East Cape, Wairarapa Eddy, and Chatham Islands.

474

475 Figure 2. **a)** Sea surface chlorophyll (SSC) from 2 October 2012 derived from the  
 476 Moderate Resolution Imaging Spectroradiometer (MODIS) on NASA's  
 477 Aqua satellite. Vectors are surface currents derived from AVISO ocean  
 478 altimeter data. Labels A, and C indicate anti-cyclonic and cyclonic eddies,  
 479 numbered eddies are referred to in the text. The blue line near eddy A<sub>1</sub>  
 480 indicates the cruise track during the one-day survey; **b)** Rossby number  
 481 (vorticity divided by planetary vorticity,  $R_0 = \zeta / f$ ) for 2 October 2012.  
 482 Dashed contours indicate  $R_0 = \pm 0.1$ ; and **c-f)** Histograms of Rossby  
 483 number for regions where SSC exceeds various levels. Dash-dot line in **a**  
 484 and **b** indicates the 250 m isobath, and data are only shown where the water  
 485 depth is greater than 250 m to exclude the coastal zone.

486

487 Figure 3. Moderate Resolution Imaging Spectroradiometer (MODIS) images of sea  
 488 surface chlorophyll (SSC) from 30 September to 4 October, 2012. **a)** 30  
 489 September; **b)** 2 October; **c)** 4 October; and **d)** enlarged view of 2 October,  
 490 along with upper mixed-layer velocity from the ship's ADCP (red vectors).  
 491 Numbered circles show the CTD cast locations from the one-day survey  
 492 made in the anticyclonic eddy A<sub>1</sub> shown in Figure 2. Colours of the circles  
 493 in **d)** indicate whether the casts were considered high-POM (green) or low-  
 494 POM (blue, see text). Black vectors are surface currents derived from  
 495 AVISO satellite altimetry.

496

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

507

508 Figure 5. Gridded mixed-layer depths derived from CTD casts shown in Figure 4.  
 509 **a)** Upper mixed layer depth (uMLD); and **b)** Remnant mixed layer depth  
 510 (rMLD). Red and blue vectors are ADCP-derived velocities at 10 m and  
 511 100 depths, respectively, showing anticyclonic circulation.

512

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

521

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

527

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

539

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

548

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