Ocean productivity south of Australia during spring and summer

Michael L. Bender¹, Bronte Tilbrook², Nicolas Cassar³, Bror Jonsson¹, Alain Poisson⁴, and Thomas W. Trull²

Abstract: We estimated mixed layer gross and net community production on a total of 20 crossings in the Australian sector of the Southern Ocean during the summer half-years (October to March) of 2007-2010. These estimates were calculated from measurements of O₂/Ar ratios and triple isotope compositions of O₂ in ~ 250 seawater samples collected underway. For comparison purposes, we also measured the seasonal drawdown of mixed layer NO₃⁻ and SiO₂ concentrations during 2006-2007 and 2007-2008. Across all samples, average values of gross and net O₂ production (measured by O₂/Ar and O₂ isotopes), were about 86 ± 90 and 18 ± 17 mmol O₂ m⁻² day⁻¹, respectively. Gross production was highest at the Subtropical Front (up to ~230 mmol O₂ m⁻² day⁻¹), and decreased southward (to ~10 near the southern boundary of the Antarctic Circumpolar Current). In contrast, net community production showed little meridional variation. Net and gross O₂ production increased throughout the spring-to-fall period, although most SiO₂ drawdown occurred in December. Consistent with satellite chlorophyll estimates, we saw no evidence for an intense spring bloom (e.g. as has been observed in the North Atlantic). Volumetric net and gross O₂ production in the mixed layer, normalized to chlorophyll, increased (with considerable scatter) with average irradiance in the mixed layer. These relationships provide a basis for estimating production from Argo float data and properties observed by satellite.

Affiliations:

1. Department of Geosciences, Princeton University, Princeton, NJ 08544, USA
2. CSIRO Oceans and Atmosphere and Antarctic Climate and Ecosystems Cooperative Research Centre, Hobart, 7001, Australia
3. Division of Earth and Ocean Sciences, Nicholas School of the Environment, Duke University, Durham, NC 27708, USA
4. Laboratoire IMAGES, Université de Perpignan Via Domitia, 52 avenue Paul Alduy, 66860 Perpignan Cedex, France Via Domitia, France.

1. Introduction

The Southern Ocean has long been identified as the largest High Nutrient Low Chlorophyll (HNLC) region in the global ocean. Variations in its status are important to nutrient delivery to other ocean basins [Sarmiento *et al.*, 2004]. They are also relevant to global carbon sequestration by the biological pump on decadal to at least interglacial timescales [Matear and Hirst, 1999; Sarmiento and Orr, 1991; Sigman and Boyle, 2000]. The High Nutrient appellation is an oversimplification. Phosphate and nitrate remain abundant across the entire Southern Ocean throughout the year. However silica, an essential nutrient for diatoms, which are the dominant large phytoplankton group, becomes depleted in summertime waters between the Subantarctic Front and the Polar Front [Nelson *et al.*, 2001; Treguer and Jacques, 1992; Trull *et al.*, 2001a]. Moreover, iron and light are actually the key limiting “nutrients”. Lack of iron, an essential element in enzymes, is now known to limit primary production essentially everywhere in the Southern Ocean except near localized sources such as islands, plateaus, and boundary currents [Boyd and Ellwood, 2010]. Low light availability driven by both deep surface mixed layers and cloudiness also limits photosynthesis, including physiological interactions between Fe scarcity and the efficiency of light utilization [Hiscock *et al.*, 2008a; B Greg Mitchell *et al.*, 1991; Strzepek *et al.*, 2012; Sunda and Huntsman, 1997]

The Low Chlorophyll status is often linked to these limitations on primary production, but that is also an oversimplification, because loss by grazing and export is at least as strong a control on phytoplankton biomass accumulation [Cullen, 1995; Frost, 1991; F M M Morel *et al.*, 1991; Smith and Lancelot, 2004]. Indeed, despite the LC status, Southern Ocean fluxes of organic carbon to the deep ocean in sinking particles are close to the global median [Honjo *et al.*, 2008; Honjo *et al.*, 2000; Trull *et al.*, 2001c]. The efficiency of this grazing and export may be modulated as mixed layer [SiO₂] varies seasonally and meridionally.

Improving the understanding of the controls on the HNLC status of the Southern Ocean requires measurements of upper ocean carbon fluxes. In this context, we present a large set of paired estimates of gross primary production (GPP) and net community production (NCP)

derived from gas tracers. The observations cover the north-south extent of Southern Ocean waters and the late spring to early autumn half-year that dominates annual production. As outlined in the Methods section, GPP estimates are based on the isotopic composition of dissolved oxygen, and the NCP estimates are based on the excess of dissolved oxygen concentrations relative to biologically inert argon concentrations. Conversion of these state variables to the GPP and NCP rate variables is achieved by calculation of gas exchange rates with the atmosphere. Both terms reflect productivities averaged over the gas exchange time of the mixed layer (=mixed layer depth/ gas transfer velocity) or ~ 2 weeks.

The results are based on discrete samples collected from the underway clean seawater supply intake onboard the French Antarctic station resupply ship, *Astrolabe*. *Astrolabe* made three round-trip crossings of the Southern Ocean each year between Hobart, Tasmania, Australia and Dumont d'Urville, Adelie Land, Antarctica. These crossings made it possible to characterize GPP and NCP six times each season. We can thus evaluate spatial, seasonal, and interannual variations in the rate terms. Using supplementary surface seawater nutrient analyses, Argo float data, and satellite observations, we then can also explore relationships with surface phytoplankton biomass, light levels, and mixed layer depths. Specifically, we focus on three questions:

1. Is there a spring phytoplankton bloom seen in primary production (as opposed to biomass accumulation) in the Australian sector of the Southern Ocean, and if so, what are its characteristics?
2. How does productivity vary with latitude in the Australian sector, and what factors mediate this relationship?
3. Is there a simple relationship between NCP and GPP and autonomously sensed properties, such as chlorophyll (Chl), sea-surface photosynthetically available radiation (PAR), and mixed layer depth (MLD)? If so, can it be used, in conjunction with ancillary data, to scale local estimates of NCP and GPP to broader regions?

As expanded on in the Discussion section, our study region south of Australia has the advantage of much background information. Material includes (a) understanding of circulation from six repeats of the WOCE/CLIVAR SR3 hydrographic section from Tasmania to Antarctica [Rintoul and Trull, 2001b; S. Sokolov and Rintoul, 2002; S. Sokolov and Rintoul, 2009; Trull *et al.*, 2001a], (b) special volumes from two month-long process studies of trace element (Fe, Mn, etc.) and macro-nutrient (N, P, Si) biogeochemistry and microbial tropho-dynamics in the SAZ [Bowie *et al.*, 2011a; Bowie *et al.*, 2011b; Bowie *et al.*, 2009; Cassar *et al.*, 2011a; Cassar *et al.*, 2015; Sedwick *et al.*, 2008; Sedwick *et al.*, 1999; Trull *et al.*, 2001b], and (c) an ongoing program, as part of Australia's Integrated Marine Observing System (www.imos.org.au), of air to sea and surface to deep ocean carbon fluxes at the Southern Ocean Time Series site [Rigual-Hernández *et al.*, 2015; Shadwick *et al.*, 2015b; Trull *et al.*, 2001c; Weeding and Trull, 2014b]. These efforts provide important context for our observations.

2. Experimental methods

2.1 Principles of the dissolved O₂ tracer methods for GPP and NCP

The O₂/Ar ratio of dissolved gases in the mixed layer, and the $\delta^{17}\text{O}$ and $\delta^{18}\text{O}$ of O₂, can be used to constrain mixed layer NCP and GPP, respectively [Castro-Morales *et al.*, 2013; Luz and Barkan, 2000; Prokopenko *et al.*, 2011; Reuer *et al.*, 2007]. The O₂/Ar ratio allows physical supersaturation (reflected in both O₂ and Ar concentrations) to be removed from the total O₂ supersaturation. O₂/Ar measurements thus quantify that part of O₂ supersaturation derived from photosynthesis in excess of respiration. Multiplying biological O₂ supersaturation by the gas transfer velocity and saturation O₂ concentration gives the sea-air flux of biological O₂ (termed “net O₂ bioflux” or just “net bioflux”). At steady state, and neglecting vertical mixing (see below), this flux must be sustained by photosynthesis in excess of respiration, and hence provides an estimate of NCP.

The estimation of GPP is based on recognition that there are two sources of dissolved O_2 in the mixed layer, atmospheric O_2 , and O_2 produced *in situ* by photosynthesis. Measuring both $\delta^{17}O$ and $\delta^{18}O$ of the dissolved O_2 constrains the fraction deriving from each of these sources. The tracer quality of the two δ terms comes from the fact that O isotopes are mass-dependently fractionated by photosynthesis and respiration (for which the change in $\delta^{17}O$ is ~ 0.5 times the change in $\delta^{18}O$). However, photochemical exchange in the stratosphere leads to non-mass-dependent fractionation, which in turn causes the global atmosphere to be depleted in ^{17}O relative to ^{18}O [Luz and Barkan, 2000; Luz *et al.*, 1999]. The relative proportions of atmospheric and photosynthetic O_2 in the dissolved O_2 pool can be computed from the magnitude of the ^{17}O deficiency. Then, from the proportion of photosynthetic O_2 , the O_2 concentration, and the gas transfer velocity, one can calculate the sea-air flux of photosynthetic O_2 (“gross bioflux”). At steady-state, and again neglecting vertical mixing, this term provides an estimate of O_2 gross primary production.

In the text that follows, we generally use the terms NCP and GPP to represent true value of production. We generally use the terms “net O_2 bioflux” and “gross O_2 bioflux” to represent the sea-air fluxes, calculated from observations, that approximate true net and gross production.

The estimates of NCP and GPP from the O_2 measurements rely on the assumptions of biological steady-state (constant NCP and GPP) and negligible mixing across the base of the surface mixed layer, as well as the parameterizations of air-sea gas exchange (as detailed below). Probable biases in net O_2 bioflux estimates of NCP have been calculated in a model context [Jonsson *et al.*, 2013]. On average, when O_2 is biologically supersaturated in the mixed layer, net O_2 bioflux underestimates NCP by about 10%. Additional random errors are much larger, probably about $\pm 40\%$, because of non-steady-state conditions, mixing across the base of the mixed layer, and analytical uncertainties. Experimental studies also show that net and gross bioflux can be influenced by upwelling, vertical mixing, and entrainment (e. g., [Castro-Morales *et al.*, 2013]). We minimize these issues by excluding from our analysis samples that are

undersaturated in O₂ (which signifies upwelling). In any case, it is means and trends in our data that are robust, rather than individual excursions.

[*Nicholson et al.*, 2014] compared gross bioflux with GPP as simulated in a model context. In their simulations, gross bioflux underestimated GPP in spring. This discrepancy was associated with a modeled spring bloom that is not observed in our data. In summer and autumn, modeled gross bioflux was 2-3 times higher than simulated GPP in the mixed layer. This error derived largely from the entrainment and mixing of deeper euphotic zone waters into the mixed layer, a phenomenon unlikely to be important in our study area. In our data set, MLD was > 40 m at 88% of XBT stations, and > 30 m at 80% of stations. According to [*Westwood et al.*, 2011], most primary production in the Australian sector occurs above 30 m depth, and about 90% above 40 m depth. Therefore it is improbable that our data is seriously aliased by entrainment and mixing of deeper photosynthetic O₂ into the mixed layer. There are additional errors associated with lateral advection and non steady-state conditions [*Nicholson et al.*, 2014]. Again, we need to focus on average values and trends rather than individual samples.

2.2 Dissolved O₂ sample collection and analysis:

The water samples (Fig. 1) were collected from the underway clean seawater supply, which has its intake at about 3 m depth on the resupply ship *Astrolabe*. Water samples were collected into pre-evacuated and HgCl₂ pre-poisoned 500 ml bottles with Leuwers-Happert O-ring valves, as described by [*Emerson et al.*, 1995] and [*Reuer et al.*, 2007]. The gases equilibrated passively between the water and headspace before analysis in Princeton, and most of the water was removed by aspiration. Non-condensable gases (essentially O₂, N₂, and Ar) were frozen onto molecular sieve and then passed through a gas chromatographic column to separate O₂ and Ar from N₂. The O₂ and Ar were frozen into a cryo-cooler at ~15 °K, warmed, and admitted to the Finnigan MAT 252 mass spectrometer as described by [*Reuer et al.*, 2007]. ¹⁸O¹⁶O, ¹⁷O¹⁶O, and ¹⁶O₂ were measured by simultaneous collection for 3 cycles with 24 blocks per cycle. ¹⁷Δ of O₂ is defined by this equation:

$$^{17}\Delta = \ln(0.001 \times \delta^{17}\text{O} + 1) - 0.516 \times \ln(0.001 \times \delta^{18}\text{O} + 1) \quad (1)$$

The uncorrected $^{17}\Delta$ of O_2 was computed, using eq. 1, from the raw data, based only on the observed ion current ratios. A number of factors can affect the raw values of $\delta^{17}\text{O}$ and $\delta^{18}\text{O}$. We evaluated these factors by analyzing the variability of between 76 and 311 air standards measured during each of 5 analysis periods. In total, these extended from February 14, 2006, to January 1, 2011. We rejected all samples where the mass 28 peak was greater than 0.3% of the mass 32 peak, because high N_2 leads to large errors in $\delta^{17}\text{O}$. We observed no dependence of $\delta^{17}\text{O}$ and $\delta^{18}\text{O}$ on the ratio of 28 to 32 ion current for the remaining samples, or on the initial pressure imbalance between sample and reference sides (data not shown). We did not correct for the zero enrichment, which effectively cancels out when seawater is compared with water equilibrated in the laboratory.

The $^{17}\Delta_{\text{raw}}$ values were corrected for differential gas depletion between sample and reference sides during analysis (Stanley et al., 2010), according to the formula:

$$\text{Imbalance term} = (1 - V_{32\text{sample-end}}/V_{32\text{sample-start}}) - (1 - V_{32\text{reference-end}}/V_{32\text{reference-start}}) \quad (2)$$

The imbalance term registers the change in the ion current imbalance between sample and reference during 1 block. “V” is the voltage, and the subscripts “end” and “start” refer to the voltages read at the end of the first block by the interfering masses program and that read at the start of the first block by the standard Finnigan ion current acquisition software from which δ values are calculated.

$^{17}\Delta$ data for a large number of equilibrated lab water samples analyzed in this way were higher than the equilibrium value (discussed below) by about 0.005 ‰. We therefore subtracted the average difference, calculated for each of our 5 analysis periods, from the corrected $^{17}\Delta$ value of O_2 . Knowing $^{17}\Delta$ and $\delta^{18}\text{O}$, we then calculated $\delta^{17}\text{O}$ from equation (1). Because of numerical issues, this approach is marginally more accurate than calculating $^{17}\Delta$ after correcting the measured $\delta^{17}\text{O}$ and $\delta^{18}\text{O}$ values independently. The choice of the coefficient (in equation 1) is

immaterial to the result: we calculate the same $\delta^{17}\text{O}$ value if we use a coefficient of 0.518. We do not use $^{17}\Delta$ for any other purpose in the analysis of our results.

$\delta^{18}\text{O}$ of O_2 is fractionated by 0.1 – 0.3 ‰ during the process of sample preparation. A correction to the measured $\delta^{18}\text{O}$ value is made which does not affect $^{17}\Delta$. O_2/Ar ratios were normalized to air and a correction was applied for the residual gas in the water sample after equilibration with the headspace. Uncertainties are ± 3 ‰ for the O_2/Ar ratio, ± 0.1 ‰ for $\delta^{18}\text{O}$ of O_2 , and ± 0.008 ‰ for $^{17}\Delta$ of O_2 .

2.3 Calculation of net and gross production from the O_2 tracers

From $\delta^{17}\text{O}$ and $\delta^{18}\text{O}$ calculated as described above, the ratio of $G/k[\text{O}_2]_{\text{eq}}$ is calculated from equation (7) of [Prokopenko *et al.*, 2011], transformed to δ notation as described in their paper:

$$\frac{G}{k\text{O}_2} = \frac{\left(1 - \frac{10^{-3}\delta^{17}\text{O}_{eq} + 1}{10^{-3}\delta^{17}\text{O}_{dis} + 1}\right) - \lambda \left(1 - \frac{10^{-3}\delta^{18}\text{O}_{eq} + 1}{10^{-3}\delta^{18}\text{O}_{dis} + 1}\right)}{\left(\frac{10^{-3}\delta^{17}\text{O}_p + 1}{10^{-3}\delta^{17}\text{O}_{dis} + 1} - 1\right) - \lambda \left(\frac{10^{-3}\delta^{18}\text{O}_p + 1}{10^{-3}\delta^{18}\text{O}_{dis} + 1} - 1\right)} \quad (3)$$

G is gross O_2 production, k is the gas transfer velocity (m day^{-1}), and O_2 is the saturation O_2 concentration. The subscripts “ p ”, “ eq ”, and “ dis ” refer to photosynthetic O_2 , equilibrium O_2 , and dissolved O_2 , respectively. $\lambda = 0.518$. As shown in this equation, $G/k[\text{O}_2]_{\text{eq}}$ is a function of the equilibrium difference between the $\delta^{17}\text{O}$ and $\delta^{18}\text{O}$ of dissolved O_2 and that of air O_2 , the $\delta^{17}\text{O}$ and $\delta^{18}\text{O}$ of dissolved O_2 (which we measure), and the $\delta^{17}\text{O}$ and $\delta^{18}\text{O}$ of photosynthetic O_2 . Air O_2 is the reference.

$\delta^{18}\text{O}_p$ is set at the value for seawater relative to air, -23.320‰, essentially indistinguishable from the value of -23.324‰ reported by [Barkan and Luz, 2011]. $\delta^{17}\text{O}$ is then set to -11.856‰. With this assignment, the $^{17}\Delta$ of seawater and photosynthetic O_2 with respect to atmospheric O_2 is then +0.249‰. These settings invoke the assumption that the $\delta^{17}\text{O}$ and $\delta^{18}\text{O}$ of

photosynthetic O_2 is identical to that of the substrate water. We note the discovery that photosynthetic O_2 is fractionated inside cells by respiration before it becomes part of the ambient O_2 pool [Luz and Barkan, 2011]. However, as long as the adopted parameters give a $^{17}\Delta$ value of +0.249‰ for seawater with respect to air, calculated values of $G/k[O_2]_{eq}$ will be insensitive to the exact values selected for the isotopic composition of photosynthetic O_2 .

Finally we need to adopt values for $\delta^{17}O_{eq}$ and $\delta^{18}O_{eq}$. $\delta^{18}O_{eq}$ was accurately measured by as a function of temperature [Benson and Krause, 1980], and we use their temperature-dependent values. We adopt the $^{17}\Delta$ value for dissolved O_2 in equilibrium with air, +0.008‰, measured by [Reuer *et al.*, 2007] and subsequently by [Stanley *et al.*, 2010].

Gas transfer velocities between the ocean and atmosphere were calculated as weighted-average values for the 60 days before collection of the sample, using wind speeds from the NOAA/Seawinds blended product [Chin *et al.*, 1998], and the gas transfer parameterization of [Sweeney *et al.*, 2007]. Gas transfer velocities closer to days of collection were weighted more heavily [Bender *et al.*, 2011; Reuer *et al.*, 2007]. The sea-air biological O_2 flux (“net O_2 bioflux”) was then simply calculated from the product of the biological O_2 supersaturation, O_2 solubility (i.e. O_2 concentration at saturation), and the gas transfer velocity (e. g., [Reuer *et al.*, 2007]). Gross O_2 bioflux was calculated from the equation of Prokopenko *et al.* (above).

The calculation of net and gross bioflux relies on the assumption that there is negligible addition of air during the collection of the water samples using the ship’s underway supply, e.g. from either air forced under the hull by ship motions or exposure to air in the pumping system. The relationship between $\delta^{18}O$ of O_2 and the O_2/Ar ratio provides an assessment of air injection during collection. In the presence of only biological processes, these two variables are anti-correlated along a line that passes slightly above the atmospheric equilibrium point (at $O_2/Ar \sim -86$ per mil relative to air, and $\delta^{18}O$ of O_2 about 0.8 ‰, in the Southern Ocean). This occurs because the photosynthetic addition of O_2 raises O_2/Ar while lowering $\delta^{18}O$ of O_2 (since photosynthetic O_2 has the same $\delta^{18}O$ as seawater, around -24‰ with respect to air O_2 , which is the standard). Conversely, respiration consumes O_2 , lowering O_2/Ar while raising $\delta^{18}O$ of O_2

(because respiration discriminates against the heavy isotope). The co-occurrence of gas exchange and variable ratios of net/gross production complicates this correlation, but nonetheless, a compact relationship between $\delta^{18}\text{O}$ of O_2 and the O_2/Ar ratio was observed (as shown in Supplementary Material Figure S1). This suggests minimal addition of air to the samples, because air addition changes the O_2/Ar ratio while leaving $\delta^{18}\text{O}$ nearly unchanged. Based on this diagnostic, visual inspection indicates that less than 5% of the samples may have experienced significant air injection and only 3 points were sufficiently compromised to require removal from the dataset (Supplementary Material). We also assume that there is negligible removal of O_2 as seawater flows through the ship's line from the intake to the sampling space (for a cautionary note see [Juranek *et al.*, 2010]).

Finally for this section, we note that either upwelling or net heterotrophy [Hamme *et al.*, 2012] can cause net O_2 bioflux to be less than zero. We believe that upwelling is by far the dominant cause, for two reasons. First, deep waters, undersaturated in O_2 , mix to the surface of the Southern Ocean, particularly south of the Polar Front. Second, in the Southern Ocean as elsewhere, there is not enough POC in the mixed layer to sustain net heterotrophy for a large fraction of the summer half-year. We include samples with negative bioflux in our data set (Supplementary Material), but, to keep the focus on productivity controls, we exclude biologically under-saturated samples from the net bioflux values plotted in all other figures. The exception is samples that were biologically undersaturated by $<0.3\%$ (86 % of all samples), which is the analytical uncertainty. In contrast, we include all gross O_2 bioflux values in all figures, (regardless of their associated O_2/Ar values). This is so because gross bioflux is much less affected by upwelling than net bioflux (since the oxygen isotopic gradient across the bottom of the mixed layer is relatively much smaller than the O_2/Ar gradient). Exchange across the base of the mixed layer also impacts NCP estimates based on nutrient depletions. From model calculations, the bias has been suggested to be up to 2-fold in spring when mixed layer variability is large, and $\sim 10\%$ in summer when stratification is well established [Wang and Matear, 2001; Wang *et al.*, 2001].

2.4 Additional data sources

Mixed layer macro-nutrient (nitrate, phosphate, and silicic acid) samples were collected in parallel to the oxygen samples (Fig. 2). Nutrient samples were frozen onboard and analyzed in Hobart following WOCE protocols.

MLDs used in the interpretation of Astrolabe bioflux data were estimated from XBT's deployed by the ship. The base of the mixed layer was identified using a temperature criterion (T_z was 0.2 °C lower than T_{surf}) [Montegut *et al.*, 2004].

PAR and Chl were characterized by remote sensing [Arrigo *et al.*, 2008], and the extinction coefficient was parameterized in terms of Chl [A Morel and Maritorena, 2001]. Average mixed layer PAR was calculated from the following equation:

$$\text{Average ML PAR} = \text{Sea surface PAR} * \int \exp(-kz) dz / \text{MLD} \quad (4)$$

k is the light attenuation coefficient and z is depth below the sea surface. The integral is evaluated from the sea surface to the MLD.

We also computed chl-normalized net and gross O₂ bioflux, dividing bioflux by Chl as inferred from remote sensing. Specifically, we used NASA MODIS/Aqua 2009b daily 4 km level 3 fields for remotely sensed SST, PAR, and Chl (<http://oceancolor.gsfc.nasa.gov>). For the purpose of calculating chl-normalized bioflux, the in-situ bioflux observations are matched to nearest available satellite data points, and Chl is registered for the closest observation within a 1 degree radius from the day of collection to 3 days before and after. The inclusion of such potentially distant matches is appropriate as the waters we sample were likely mixed over a large area during the period prior to sampling. We performed sensitivity tests that suggest that the typical maximum error in Chl due to matching of distant locations is less than $\pm 50\%$, which is within the acceptable range for our application. However, the satellite chlorophyll retrievals add scatter to the calculated values of chl-normalized production. We also note that a number of

papers discuss evidence that Southern Ocean chlorophyll retrievals are aliased toward values lower than in situ measurements [Johnson *et al.*, 2013; Marrari *et al.*, 2006; B. G. Mitchell and Kahru, 2009; Szeto *et al.*, 2011]. This effect may alias our calculations and can be addressed when the Southern Ocean chl algorithm is better constrained.

The complete oxygen-based dataset is tabulated in the Supplementary Material.

3. Results

We first present contextual observations on the spatial and seasonal variations of water column conditions. These come from the in-situ discrete nutrient analyses (section 3.1) and the ARGO and satellite remote sensing estimates of mixed layer depth, insolation, and chlorophyll (section 3.2). We then discuss net and gross bioflux estimates (section 3.3), and examine the relationship between chl-normalized production and mixed layer irradiance.

3.1. Macro-nutrient drawdown

[NO₃⁻] and [SiO₂] are plotted vs. latitude and salinity in Fig. 2 for 2006-2007 and 2007-2008,. The data show the well-known first-order features of Southern Ocean nutrient distributions: increasing concentrations towards higher latitudes (as a result of upwelling), and decreasing concentrations as the growing season progresses (driven by NCP). They also reveal significant interannual variations as well as meridional differences in the stoichiometry of nutrient drawdown. Nitrate concentrations were near zero in Subtropical waters at the very northern end of the transects, and increased strongly across the SAZ to values greater than 20-25 μM at the Subantarctic Front. [NO₃⁻] continued to rise slowly across the PFZ and AZ to reach values of nearly 30 μM south of the ACC. In contrast, silicate concentrations were lower over a

much larger portion of the Southern Ocean. $[\text{SiO}_2]$ was $<5 \mu\text{M}$ in both Subtropical and Subantarctic waters. Going southward, we observed both progressive and stepwise SiO_2 increases.

Fig. 2 shows that the extent and timing of nutrient depletion, as well as Si/N drawdown ratios, varied strongly from zone to zone. In the AZ, where winter Si concentrations were very high, the Si/N drawdown ratio over the season was nearly 10. In the PFZ, it was closer to 4, and decreased further to ~ 1 in the SAZ and to $\sim 0.1\text{-}0.2$ in Subtropical waters. High ratios for the Si/N decrease in the upper water column, such as we find, were previously observed along the nearby WOCE/CLIVAR SR3 north-south hydrographic section [Trull *et al.*, 2001a], and in the AESOPS study south of New Zealand (e. g., [Morrison *et al.*, 2001; Smith Jr *et al.*, 2000]). There are multiple contributing factors to these variations. In the AZ and PFZ, production is dominated by diatoms. Their Si uptake is known to increase with Si availability, and the Si/N uptake ratio is expected to be higher under the Fe limited conditions that prevail there [Bowie *et al.*, 2009; Cassar *et al.*, 2011b; Sosik and Olson, 2002]. The uptake ratio under Fe limitation is near 4 vs. near 1 for Fe replete growth, at least for some species [Brzezinski *et al.*, 2002; Franck *et al.*, 2000; Takeda, 1998]. In the SAZ and especially Subtropical waters, non-siliceous organisms contribute significantly to NCP, lowering the Si/N drawdown ratio [Kopczynska *et al.*, 2001; Trull *et al.*, 2001c]. In addition to these uptake-based controls on Si/N drawdown, remineralization is important. N is recycled more rapidly than Si, attenuating the ratio of $\text{NO}_3^-/\text{SiO}_2$ drawdown [Green and Sambrotto, 2006; Rubin, 2003]. In partial support, there is some evidence for significant nitrification within the surface mixed layer [Fripiat *et al.*, 2015]. N/P drawdown stoichiometry also revealed zonal variations, with the median uptake ratio of N/P ranging from 13-19 in the Subtropical, Subantarctic, and Antarctic zones, but only ~ 9 in the PFZ. This result agrees well with [Lourey and Trull, 2001b], who observed a ratio of 15.1 in the SAZ and 8.3 in the PFZ, but the cause remains unclear.

Calculation of NCP from the seasonal surface nitrate drawdown is difficult, because vertical distributions were not measured yet appear to be important especially in the PFZ [Lourey and

Trull, 2001a; Parslow *et al.*, 2001]. As well, both vertical mixing and northward Ekman transport provide significant nutrient resupply [DiFiore *et al.*, 2006a; Green and Sambrotto, 2006; Rubin, 2003]. Model estimates suggest that these two transport modes double fluxes calculated from springtime nutrient drawdown [Wang and Matear, 2001; Wang *et al.*, 2001]. Keeping these caveats in mind, we make approximate first estimates as follows. The seasonal $[\text{NO}_3^-]$ decrease in the Subantarctic Zone was 3-4 $\mu\text{mol kg}^{-1}$, and about 2-3 $\mu\text{mol kg}^{-1}$ between the SAF and the SACCF (Fig. 2). For a summer MLD of ~ 40 m in the Subantarctic Zone and ~ 80 m to the south, and assuming $\text{C/N} = 7$, then seasonal NCP would be about 1 mol C m^{-2} across the entire transect from the Subtropical Front to the SACCF. In comparison, depth resolved nitrate estimates suggest $\sim 3\text{-}4$ mol C m^{-2} for the SAZ from October to March, a period over which mixed depth shoaled from 200 to ~ 50 m [DiFiore *et al.*, 2006b; Lourey and Trull, 2001a].

There is considerable interannual variability in nutrient drawdown. As an example of interannual variability, October NO_3^- and SiO_2 concentrations were meridionally constant, across the AZ from the Polar Front ($\sim 56^\circ$ S) south to about 61° S, in each year of our observations. However, $[\text{NO}_3^-]$ seasonal drawdown was about 1 μM in 2006-7 but closer to 3 μM in 2007-8. In this same region, Si was almost fully depleted by January in 2007 but only about half depleted in 2008. The timing of SiO_2 depletion in our data set is similar to other observations, which also suggest that the seasonal consumption mainly occurs before the end of January [Green and Sambrotto, 2006; Morrison *et al.*, 2001; Rubin, 2003].

3.2 Mixed layer depth, irradiance, and chlorophyll variations

In Hovmöller diagrams, Fig. 3 shows weekly estimates for mixed layer depth (MLD), euphotic depth (Z), Chlorophyll concentrations (Chl), and surface photosynthetically available radiation (PAR). Also shown is depth averaged PAR for the mixed layer, and the differences

between the mixed layer and euphotic depths. Also shown are our sparser measurements of net O_2 bioflux, i.e. estimates of NCP.

There are clear seasonal cycles and considerable interannual variability in all these properties. MLD follows the well-known seasonal cycle: deep in winter and shallow in summer. The seasonality is at its largest in the Subantarctic Zone [Rintoul and Trull, 2001b; Rintoul *et al.*, 1997; S. Sokolov, 2008; Trull *et al.*, 2001a]. To generalize from 3 years' data, mixed layers at the beginning of October are >150 m deep north of the Polar Front, and shallower to the south. Mixed layers shoal in October in the north and November in the south. In the fall there is a slow transition back to deep wintertime values. Sea surface PAR values track the sun, with maximum values close to summer solstice. Average values of mixed layer PAR have greater variation in seasonal timing. In 2007-2008, mixed layer average PAR in the northern part of the study region was highest before summer solstice. In the next two years it was highest after. This pattern tracks MLD, and is a result of interannual variability in water column stratification. Shallower MLD's were more prevalent in the early part of the growing season in 2007-2008, and more prevalent in the latter part the following 2 years.

In the northern part of the study area, Chl levels rose later in the growing season than the timing of seasonal shoaling of MLD or the increase in mixed layer PAR. In other words, biomass (or at least chlorophyll) continued to increase after stratification. Highest Chl levels were not observed until the end of January, or February. During summer, Chlorophyll concentrations reached a maximum around the Subtropical Front, then decreased towards the south. There was a suggestion that Chl levels also increased at the southern end of the transect, as Antarctica was approached - a feature that is seen more strongly to the east of the Astrolabe transit in satellite data [S. Sokolov, 2008]). Even the higher Chl accumulations that occurred in the SAZ were not large, rarely reaching more than $1 \mu g L^{-1}$ (3). These levels are similar to those reached in the iron-limited sub-arctic North Pacific [Harrison *et al.*, 2004] and can be contrasted with levels 3-5 times higher that develop in the more iron-rich [Ryan-Keogh *et al.*, 2013] sub-arctic North Atlantic spring bloom (e.g.[Mahadevan *et al.*, 2012]).

During the mid-summer periods of highest Chl accumulations, the mixed layer was generally shallower than the euphotic zone (see citations in section 2.1). Vertical mixing would cause our fluxes to overestimate ML production. The effect is strongest north of the subtropical front in summer (Fig. 3), and would cause us to overestimate summertime NCP between 43°-46° south. This effect is unlikely to be important south of the Subtropical Front, where only a small fraction of primary production takes place below the mixed layer (see section 2. 1.).

3.3 Spatial and temporal distributions of Gross and Net Community Production

Net O₂ bioflux (as an estimate of NCP) and gross O₂ bioflux (as an estimate of GPP) are plotted vs. latitude and averaged over 1° latitude bands in Fig. 4 (data for all samples are provided in Supplementary Figure S2). In this and subsequent discussions of meridional variations, we exclude from the data set results from the Astrolabe cruise that is to the east of the other cruise tracks (Fig. 1). Net bioflux (averaged over all samples with biological O₂ supersaturation > -0.3%) was $18 \pm 17 \text{ mmol m}^{-2} \text{ day}^{-1}$ (1σ , n=213). Gross O₂ bioflux was 91 ± 110 . As discussed above, NCP is likely to be slightly higher than net bioflux, which is aliased low by vertical mixing. The ratio of both average and median net O₂ bioflux/gross O₂ bioflux was 0.20. These values are all similar to those of other studies of large domains in the Southern Ocean [Green and Sambrotto, 2006; Morrison et al., 2001; Reuer et al., 2007; Rubin, 2003].

Gross bioflux was highest from 43-48° S; in contrast, net bioflux showed no strong meridional trend (Fig. 4). This situation may be compared with that during the autumn SAZ-SENSE voyage (Feb. 2007). During this study, ¹⁴C production, gross O₂ bioflux, and net O₂ bioflux were all highest just to the south of the Subtropical Front near 45-46° S [Cassar et al., 2011b]. However, it is important to note that the SAZ-SENSE data did not show a monotonic southward decrease of net bioflux. Net bioflux was very low in the southern part of the SAZ, then rose south of the Polar Front. In our Astrolabe data, gross bioflux was highest at the northern end of the transect, but not net O₂ bioflux. On the other hand, data from the New

Zealand sector [Reuer *et al.*, 2007] showed a persistent southward decrease in both net and gross bioflux from November or December through February.

Monthly average values (and standard errors) for gross and net bioflux are plotted vs. Southern Ocean Zone, from STZ to SACC, in Fig. 5. These data again illustrate the main temporal and meridional trends. There is a southward decrease in gross bioflux but little or no decrease in net bioflux. Maximum values of net bioflux come after the New Year. South of the Subtropical Front, maximum values of both net and gross bioflux occur in January. This feature may be linked to the rapid drawdown of SiO_2 in the December-January period.

The maintenance of high gross production into the beginning of March may be associated with elevated chlorophyll and shallow mixed layer depths during this time (Fig. 3). Iron is likely decreasing throughout the growing season [Bowie *et al.*, 2009; Sedwick *et al.*, 2008], but light and chlorophyll apparently win in our domain.

For the period from October to the beginning of March (~ 130 days), net O_2 bioflux averaged 17, 18, and 17 $\text{mmol m}^{-2} \text{ day}^{-1}$ for the Subtropical Zone, Subantarctic Zone, and Polar Front zone respectively. For a growing season of 130 days and $\Delta\text{CO}_2/\Delta\text{O}_2 = 117/170$ (Sarmiento and Anderson, 1994), NCP inferred from our O_2/Ar data would be $1.5 \text{ mol C m}^{-2} \text{ year}^{-1}$. Thus our estimates are in reasonably good agreement with values estimates from a detailed seasonal cycle study at $46^\circ 56' \text{ S}$, $142^\circ 15' \text{ E}$ using hourly autonomous moored observations of O_2 and pCO_2 ($\sim 2 \text{ mol C m}^{-2}$) [Shadwick *et al.*, 2015a; Weeding and Trull, 2014b]). Estimates from NO_3^- drawdown are similar (section 3.2). As with the Chl abundances, our NCP estimates are relatively low overall ($< 20 \text{ mmol m}^{-2} \text{ day}^{-1}$). These values are similar to sub-Arctic North Pacific waters (e.g. [Harrison *et al.*, 2004]). They are about an order of magnitude below NCP values during the spring bloom in the sub-Arctic North Atlantic [Alkire *et al.*, 2012; Bender *et al.*, 1992; Quay *et al.*, 2012]. There, NCP can be as high as 1 mole C m^{-2} over a period of order 1 week, equivalent to net O_2 bioflux values of about $200 \text{ mmol m}^{-2} \text{ day}^{-1}$.

The seasonal cycle of NCP inferred here for the Subantarctic Zone is very different from that of [Weeding and Trull, 2014a], who measured $[\text{O}_2]$ and gas tension using a mooring.

They diagnosed very high values of O_2 NCP in early spring. These high values were required to explain the observed time series of $[O_2]$, given evidence for rapid entrainment from deep mixed layers and highly variable mixed layer depths. This entrainment kept O_2 undersaturated at the surface until the middle of December, after which time vertical mixing has little influence on the O_2 balance of the mixed layer (their Fig. 5C). In contrast, the mixed layer was generally supersaturated in biological O_2 during our cruises (Table S-1). Also, we calculated NCP only for waters that were biologically supersaturated in O_2 . Thus biases from entrainment in our observations are likely to be small.

The difference between the two studies may also reflect spatial variations in the depth and intensity of late winter and early spring deep convection south of Australia. , The Southern Ocean Time Series site southwest of Tasmania studied by Weeding and Trull [2014b] exhibits deeper winter mixed layers than along the our Astrolabe transects and the nearby WOCE/Clivar SR3 line [Dong *et al.*, 2008; Rintoul and Bullister, 1999; Rintoul and Trull, 2001a]. Interannual variability may also contribute to the difference. We observed biological supersaturation in the spring of 2007 and 2008, but mixed saturation status in spring, 2009. Significant interannual variability in biological O_2 supersaturation and net O_2 bioflux is hardly surprising given the large difference in Subantarctic Zone nitrate concentrations in October and November, 2006, relative to 2007. Spatio-temporal variations may also explain high NCP inferred from nutrient and carbon dioxide observations in Drake Passage in springtime [Munro *et al.*, 2015]. Overall, these other observations of high springtime NCP do not invalidate our conclusion that, in O_2 -supersaturated waters south of Australia, springtime productivity is not particularly high.

4. Discussion

4.1 Relationships between Chl, mixed layer PAR, and gross and net bioflux

The relationship between sea surface PAR, MLD, and production has been at the heart of attempts to understand seasonal and spatial variability in productivity back to at least the classic paper of [Sverdrup, 1953]. It is also at the center of algorithms to calculate ^{14}C production from

remotely observed properties of the sea surface (e. g., [Behrenfeld and Falkowski, 1997]). Here we analyze the relationship between these properties using a simple approach that has basic mechanistic content. Our objectives are to identify a simple and transparent relationship between these properties, and to advance the possibility of simulating NCP from remotely sensed properties and Argo/BioARGO float data.

Specifically, we focus on the relationship between average volumetric production in the mixed layer normalized to Chl, and average irradiance in the mixed layer. One expects these terms to be related, because productivity increases as irradiance rises and as Chl rises. Our work follows, for example, the Vertically-Generalized Productivity Model of [Behrenfeld and Falkowski, 1997]. Upon rearranging Eq. (10) of their paper, one gets an equation stating that the rate of Chl-normalized volumetric primary production in the euphotic zone scales with irradiance. Their irradiance term takes into account light saturation, and their model is more process- rich than ours (therefore also more complex), but the basic premises are similar.

In our model, the relevant terms are calculated as follows:

$$\text{Chl-normalized volumetric gross bioflux in the ML} = \text{gross O}_2 \text{ bioflux} / (\text{MLD} * \text{Chl}) \quad (5)$$

$$\text{Chl-normalized volumetric net bioflux in the ML} = \text{net O}_2 \text{ bioflux} / (\text{MLD} * \text{Chl}) \quad (6)$$

The equation for calculating average ML irradiance was presented earlier. Chl is in mg m^{-3} . The units of volumetric bioflux are $\text{mmol O}_2 \text{ mg Chl}^{-1} \text{ day}^{-1}$. Average ML irradiance has the units of $\text{photons m}^{-2} \text{ day}^{-1}$.

In evaluating the relationship between normalized bioflux and average ML irradiance, we exclude all samples with $\text{Chl} < 0.10 \text{ mg m}^{-3}$, because at these low levels Chl retrieved from space may have large uncertainties (e. g., [Johnson et al., 2013]). As outlined above, we exclude samples from the net bioflux analysis that were biologically undersaturated in O_2 by $>0.3\%$. We also exclude two measurements of gross bioflux that were extreme outliers. We then plot normalized gross O_2 bioflux and normalized net O_2 bioflux vs. average ML irradiance (Fig. 6).

Chl-normalized volumetric gross and net O_2 bioflux both rise with increasing values of average ML irradiance, albeit with continuous scatter. Out to a light flux of $\sim 25 \text{ mol m}^{-2} \text{ day}^{-1}$,

there is no strong indication that irradiance is saturating. The R^2 is 0.44 for both normalized net bioflux and normalized gross bioflux plotted vs. average ML irradiance. This level of shared variance is remarkably high given errors in derived properties (satellite Chl observations, gas transfer velocity, mixed layer PAR, light attenuation coefficient) and the simplistic nature of the assumptions for calculating O_2 bioflux (steady state, no water flux across the base of the mixed layer).

We analyzed the apparent absence of light saturation in the plots of normalized production vs. average ML irradiance using a simple model based on curves of 1-hour ^{14}C production vs. irradiance [Westwood *et al.*, 2011] on the SAZ-SENSE cruise. According to Westwood *et al.*, saturating irradiance during the SAZ-SENSE cruise averaged $56 \mu\text{mol photons m}^{-2} \text{sec}^{-1}$ when evaluated for those stations where the MLD was within the euphotic zone. For a 16-hour day, this number corresponds to about $3 \text{ mol photons m}^{-2} \text{day}^{-1}$. It is approximately equal to the median value of average daily ML irradiance for our samples (Fig. 6). However, Fig. 6 shows that gross and net O_2 bioflux continue to increase at higher irradiances, despite the fact that average daily ML irradiance is saturating. There are 2 reasons. First, light decreases exponentially with depth. Therefore, most of the mixed layer lies below the depth of average ML irradiance. Second, irradiance will be low, and production will be light-limited, towards the beginning and end of the photoperiod. For these 2 reasons, production during the photoperiod may be light-limited even if average irradiance in the mixed layer exceeds the light-saturated value.

In Fig. 7, we quantify these effects by plotting relative O_2 production vs. irradiance for the case where saturating irradiance = $56 \mu\text{mol photons m}^{-2} \text{sec}^{-1}$. We assume that instantaneous production increases linearly with irradiance until it reaches the saturation value (at $56 \mu\text{mol photons m}^{-2} \text{sec}^{-1}$), and is constant thereafter. In our calculation, insolation decreases symmetrically about the noontime value along a hyperbolic curve that goes to zero at 4 AM and 8 PM (i.e., 16h/8 h light/dark cycle). We calculate relative ML production for MLD from 1-100 m, in 1 m intervals. Corresponding values of average ML irradiance decrease from 45 to $6 \mu\text{mol}$

m⁻² sec⁻¹. Mixed layer O₂ production saturates at an average irradiance of ~ 15 mol photons m⁻² day⁻¹, approaching the highest value observed for our samples. Hence most of the mixed layer is light-limited for most of the day, and we expect that gross O₂ production will increase with increasing irradiance, given the ambient range of ML irradiances. In practice, of course, production must eventually saturate at high irradiance. This dependence may be present in the curve of normalized gross O₂ evolution (Fig. 7, left), but there is a lot of scatter.

Various factors complicate the relationship between light, Chl, and O₂ production is the iron sufficiency of the ecosystem, which also influences μ . There are ample data suggesting that iron sufficiency changes in the study region (e. g., [Cassar *et al.*, 2011b]). [Hiscock *et al.*, 2008b] argued that the effect of iron on the physiology of individual phytoplankton is to increase the initial slope of the P-I curve while leaving light-saturated production unchanged. In their experiments, the slope varied by about a factor of 2 between Fe-limited and Fe-sufficient phytoplankton. This variability would lead to scatter in the curve of gross O₂ production (or bioflux) vs. average ML irradiance, but the basic relationship would be unchanged. In fact the greatest relevant effect of Fe is simply to cause a rise in the Chl. We implicitly acknowledge this effect by normalizing to Chl. Other obvious confounding factors include water temperature, light acclimation, and the changing nature of Southern Ocean ecosystems.

Despite these complications, our results show that there is a strong relationship between volumetric net and gross bioflux normalized to Chl, and average ML irradiance. These relationships can be exploited, using remotely sensed Chl values and MLD's from ARGO floats and other sources, to scale local estimates of GPP and NCP to broader areas of the Southern Ocean. Scaled values can be used to achieve a better understanding of process-level controls on production, and to improve the simulation of biogeochemical cycles in models.

4.2 **Net/gross bioflux variations** There are many possible controls on the overall trend of lower NCP/GPP ratios in the northern regions of the Southern Ocean observed here (Figure 4). For example, zooplankton grazing and remineralization are both more effective in warmer waters [Laws *et al.*, 2000; Steinberg *et al.*, 2008]. In addition, lower levels of SiO₂ in the north may limit the growth of diatom micro-plankton, whose size allows them to more readily escape grazing pressure [Assmy *et al.*, 2013; F M M Morel *et al.*, 1991]. Importantly, the complexity of tropho-dynamics means that high NCP can still occur in small phytoplankton communities, as has been observed previously in this region [Cassar *et al.*, 2015], including contributions to carbon export to the ocean interior [Richardson and Jackson, 2007].

V. Conclusions

Our data allow us to answer our three key research questions, outlined in the Introduction, as follows:

1. In the Australian sector of the Southern Ocean there is no evidence of a spring bloom in the sense of very high rates of photosynthesis and carbon export.
2. The oxygen based GPP and NCP estimates (and the SiO₂ and NO₃⁻ drawdowns) show considerable latitudinal and year to year variability, but with an overall trend of higher GPP in the north and little latitudinal variation in NCP.
3. Chl-normalized values of net and gross O₂ bioflux scale with average mixed layer irradiance, albeit with a high level of scatter. This relationship provides a useful check on models of productivity based on remote sensing, that can be evaluated further in future studies. The relationship also provides a basis for scaling GPP and NCP in the Southern Ocean based on Chl observed from satellites and mixed layer depths from ARGO floats and other sources.

Acknowledgements:

Sampling on Astrolabe was supported by a French-Australian research collaboration and involved many dedicated field personnel and help from ships officers and crew. The Institute Polaire Francais are thanked for allowing access to the ship and for help with field operations, as is the assistance provided by Drs. C. Goyet and N. Metzl. The Australian Climate Change Research Program, Antarctic Climate and Ecosystems Co-operative Research Centre, and the Integrated Marine Observing System funded the collection of water samples and nutrient analyses, and the operation of underway sensors reported in this study. Bender, Cassar, and Jonsson gratefully acknowledge financial support from NASA and the Office of Polar Programs, NSF. Alan Poole and Matt Sherlock from CSIRO were frequently called on to ensure equipment functioned for the voyages.

References

- Alkire, M. B., et al. (2012), Estimates of net community production and export using high-resolution, Lagrangian measurements of O₂, NO₃⁻, and POC through the evolution of a spring diatom bloom in the North Atlantic, *Deep-Sea Research I*, 64, 157-174.
- Assmy, P., V. Smetacek, M. Montresor, C. Klaas, J. Henjes, V. H. Strass, J. M. Arrieta, U. Bathmann, G. M. Berg, and E. Breitbarth (2013), Thick-shelled, grazer-protected diatoms decouple ocean carbon and silicon cycles in the iron-limited Antarctic Circumpolar Current, *Proceedings of the National Academy of Sciences*, 110(51), 20633-20638.
- Behrenfeld, M. J., and P. G. Falkowski (1997), Photosynthetic rates derived from satellite-based chlorophyll concentration, *Limnology and Oceanography*, 42, 1-20.
- Bender, M. L., S. Kinter, N. Cassar, and R. Wanninkhof (2011), Evaluating gas transfer parameterizations using upper ocean radon distributions, *Journal of Geophysical Research*, 116, doi: 10.1029/2009JC0058-1025.
- Bender, M. L., H. Ducklow, J. Kiddon, J. Marra, and J. Martin (1992), The carbon balance during the 1989 spring bloom in the North Atlantic Ocean, 47° N, 20°W, *Deep-Sea Research Part A*, 39, 1707-1725.
- Bowie, A. R., T. W. Trull, and F. Dehairs (2011a), Estimating the Sensitivity of the Subantarctic Zone to Environmental Change: the SAZ-Sense project., *Deep Sea Research II*, 58(21-22), 2051-2058.
- Bowie, A. R., F. B. Griffiths, F. Dehairs, and T. W. Trull (2011b), Oceanography of the subantarctic and Polar Frontal Zones south of Australia during summer: Setting for the SAZ-Sense study, *Deep-Sea Research II*, 58(21-22), 2059-2070.
- Bowie, A. R., D. Lannuzel, T. A. Remenyi, T. Wagener, P. J. Lam, P. W. Boyd, C. Guieu, A. T. Townsend, and T. W. Trull (2009), Biogeochemical iron budgets of the Southern Ocean

- 636 south of Australia: Decoupling of iron and nutrient cycles in the subantarctic zone by
 637 the summertime supply, *Global Biogeochemical Cycles*, 23.
- 638 Boyd, P. W., and M. J. Ellwood (2010), The biogeochemical cycle of iron in the ocean, *Nature*
 639 *Geoscience*, 3, 675 - 682.
- 640 Brzezinski, M. A., C. J. Pride, V. M. Franck, D. M. Sigman, J. L. Sarmiento, K. Matsumoto, N.
 641 Gruber, G. H. Rau, and K. H. Coale (2002), A switch from Si(OH)_4 to NO_3^- depletion in
 642 the glacial Southern Ocean, *Geophysical Research Letters*, 29(12), 5-1-5-4.
- 643 Cassar, N., P. DiFiore, B. A. Barnett, M. L. Bender, A. R. Bowie, B. Tilbrook, K. Petrou, K.
 644 Westwood, S. Wright, and T. Wagener (2011a), The influence of iron and light on net
 645 community production in the Subantarctic and Polar Frontal Zones, *Biogeosciences*, 7,
 646 5649-5674.
- 647 Cassar, N., P. J. DiFiore, B. A. Barnett, M. L. Bender, A. R. Bowie, B. Tilbrook, K. Petrou, K. J.
 648 Westwood, S. W. Wright, and D. Lefevre (2011b), The influence of iron and light on net
 649 community production in the Subantarctic and Polar Frontal Zones, *Biogeosciences*, 8,
 650 227-237.
- 651 Cassar, N., et al. (2015), The relation of mixed-layer carbon export production to plankton
 652 community in the Southern Ocean, *Global Biogeochemical Cycles*, .
- 653 Castro-Morales, K., N. Cassar, D. R. Shoosmith, and J. Kaiser (2013), Biological production in
 654 the Bellingshausen Sea from oxygen-to-argon ratios and oxygen triple isotopes,
 655 *Biogeosciences*, 10, 2273-2291.
- 656 Chin, T. M., R. F. Milliff, and W. G. Large (1998), Basin-scale, high-wavenumber sea surface
 657 wind fields from a multiresolution analysis of scatterometer data, *Journal of Atmospheric*
 658 *and Oceanic Technology*, 15, 741-763.
- 659 Cullen, J. J. (1995), Status of the iron hypothesis after the Open-Ocean Enrichment
 660 Experiment, *Limnology and Oceanography*, 40(7), 1336-1343.

- 661 DiFiore, P. J., D. M. Sigman, T. W. Trull, M. J. Lourey, K. Karsh, G. Cane, and R. Ho (2006a),
 662 Nitrogen isotope constraints on subantarctic biogeochemistry, *Journal of Geophysical*
 663 *Research*, *111*, doi 10.1029/2005JC003216.
- 664 DiFiore, P. J., D. M. Sigman, T. W. Trull, M. J. Lourey, K. Karsh, G. Cane, and R. Ho (2006b),
 665 Nitrogen isotope constraints on Subantarctic biogeochemistry, *Journal of Geophysical*
 666 *Research*, *111*, C08016, doi:08010.01029/02005JC003216.
- 667 Dong, S., J. Sprintall, S. T. Gille, and L. Talley (2008), Southern Ocean mixed layer depth from
 668 Argo float profiles, *Journal of Geophysical Research-Oceans*, *113*, 2008.
- 669 Franck, V. M., M. A. Brzezinski, K. H. Coale, and D. M. Nelson (2000), Iron and silicic acid
 670 concentrations regulate Si uptake north and south of the Polar Frontal Zone in the
 671 Pacific Sector of the Southern Ocean, *Deep-Sea Research Part II*, *47*, 3315-3338.
- 672 Fripiat, F., et al. (2015), Enhanced nitrification in a natural iron-fertilized bloom of the
 673 Southern Ocean, *Global Biogeochemical Cycles*, in review.
- 674 Frost, B. (1991), The role of grazing in nutrient-rich areas of the open sea, *Limnology and*
 675 *Oceanography*, *36*, 1834-1850.
- 676 Green, S. E., and R. N. Sambrotto (2006), Net community production in terms of C, N, P and
 677 Si in the Antarctic Circumpolar Current and its influence on regional water mass
 678 characteristics, *Deep-Sea Research Part I*, *53*, 111-135.
- 679 Hamme, R. C., et al. (2012), Dissolved O₂/Ar and other methods reveal rapid changes in
 680 productivity during a Lagrangian experiment in the Southern Ocean, *Journal of*
 681 *Geophysical Research Oceans*, *117*, doi: 10.1029/2011JC007046.
- 682 Harrison, P. J., F. A. Whitney, A. Tsuda, H. Saito, and K. Todokoro (2004), Nutrient and
 683 plankton dynamics in the NE and NW gyres of the Subarctic Pacific Ocean, *Journal of*
 684 *Oceanography*, *60*, 93-117.
- 685 Hiscock, M. R., V. P. Lance, A. M. Apprill, R. R. Bidigare, Z. I. Johnson, B. G. Mitchell, W. O.
 686 Smith, and R. T. Barber (2008a), Photosynthetic maximum quantum yield increases are

- 687 an essential component of the Southern Ocean phytoplankton response to iron,
 688 *Proceedings of the National Academy of Sciences*, 105(12), 4775-4780.
- 689 Hiscock, M. R., V. P. Lance, A. M. Apprill, R. R. Bidigare, Z. I. Johnson, B. G. Mitchell, W. O.
 690 Smith, and R. T. Barber (2008b), Photosynthetic maximum quantum yield increases are
 691 an essential component of the South Ocean phytoplankton response to iron, *Proceedings*
 692 *of the National Academy of Sciences*, 105, 4775-4780.
- 693 Honjo, S., S. J. Manganini, R. A. Krishfield, and R. Francois (2008), Particulate organic carbon
 694 fluxes to the ocean interior and factors controlling the biological pump: A synthesis of
 695 global sediment trap programs since 1983, *Progress in Oceanography*, 217-285, 217-
 696 285.
- 697 Honjo, S., R. Francois, S. Manganini, J. Dymond, and R. Collier (2000), Particle fluxes to the
 698 interior of the Southern Ocean in the Western Pacific sector along 170°W, *Deep-Sea*
 699 *Research II*, 47(15-16), 3521-3548.
- 700 Johnson, R., P. G. Strutton, S. W. Wright, A. McMinn, and K. M. Meiners (2013), Three
 701 improved satellite chlorophyll algorithms for the Southern Ocean, *Journal of Geophysical*
 702 *Research: Oceans*, 118(7), 3694-3703.
- 703 Jonsson, B. F., S. C. Doney, J. Dunne, and M. Bender (2013), Evaluation of the Southern
 704 Ocean O₂/Ar-based NCP estimates in a model framework, *Journal of Geophysical*
 705 *Research: Biogeosciences*, 118(2), 385-399.
- 706 Juranek, L. W., R. C. Hamme, J. Kaiser, R. Wanninkhof, and P. D. Quay (2010), Evidence of O₂
 707 consumption in underway seawater lines: IMplications for air-sea O₂ fluxes,
 708 *Geophysical Research Letters*, 37, DOI: 10.1029/2009GL04023.
- 709 Kopczynska, E. E., F. Dehairs, M. Elskens, and S. Wright (2001), Phytoplankton and
 710 microzooplankton variability between the Subtropical and Polar Fronts south of
 711 Australia (SAZ' 98): thriving under regenerative and new production in late summer,
 712 *Journal of Geophysical Research*, 106(C12), 31,597-531,609.

- 713 Laws, E. A., P. G. Falkowski, W. O. J. Smith, H. Ducklow, and J. J. McCarthy (2000),
 714 Temperature effects on export production in the open ocean, *Global Biogeochemical*
 715 *Cycles*, 14(4), 1231-1246.
- 716 Lourey, M. J., and T. W. Trull (2001a), Seasonal nutrient depletion and carbon export in the
 717 Subantarctic and Polar Frontal Zones of the Southern Ocean south of Australia, *Journal*
 718 *of Geophysical Research-Oceans*, 106(C12), 31463-31487.
- 719 Lourey, M. J., and T. W. Trull (2001b), Seasonal nutrient depletion and carbon export in the
 720 Subantarctic and Polar Frontal Zones of the Southern Ocean south of Australia, *Journal*
 721 *of Geophysical Research*, 106, 31463-31487.
- 722 Luz, B., and E. Barkan (2000), Assessment of oceanic productivity with the triple-isotope
 723 composition of dissolved oxygen, *Science*, 288(5473), 2028-2031.
- 724 Luz, B., E. Barkan, M. L. Bender, M. H. Thieme, and K. A. Boering (1999), Triple-isotope
 725 composition of atmospheric oxygen as a tracer of biosphere productivity, *Nature*,
 726 400(6744), 547-550.
- 727 Mahadevan, A., E. D'Asaro, C. Lee, and M. J. Perry (2012), Eddy-driven stratification initiates
 728 North Atlantic spring phytoplankton blooms, *Science*, 54, 54-58.
- 729 Marrari, M., c. Hu, and K. Daly (2006), Validation of SeaWiFS chlorophyll a concentrations in
 730 the Southern Ocean: A revisit, *Remote Sensing of the Environment*, 105, 367-376.
- 731 Matear, R., and A. C. Hirst (1999), Climate change feedback on the future oceanic CO₂
 732 uptake, *Tellus*, 51(3), 722-733.
- 733 Mitchell, B. G., and M. Kahru (2009), Bio-Optical algorithmes for ADEOS-2 GLI, *Journal of the*
 734 *Remote Sensing Society of Japan*, 29, 80-85.
- 735 Mitchell, B. G., E. A. Brody, O. Holm-Hansen, C. McClain, and J. Bishop (1991), Light
 736 limitation of phytoplankton biomass and macronutrient utilization in the Southern
 737 Ocean, *Limnology and Oceanography*, 36(8), 1662-1677.

- 738 Montegut, C., G. Madec, A. S. Fischer, A. Lazar, and D. Iudicone (2004), Mixed layer depth
 739 over the global ocean: An examination of profile data and a profile-based climatology,
 740 *Journal of Geophysical Research*, *109*, doi: 10.1029/2004JC002378.
- 741 Morel, A., and S. Maritorena (2001), Bio-optical properties of oceanic waters: A reappraisal,
 742 *Journal of Geophysical Research: Oceans*, *106*, 7163-7180.
- 743 Morel, F. M. M., J. G. Reuter, and N. M. Price (1991), Iron nutrition of phytoplankton and its
 744 possible importance in the ecology of ocean regions with high nutrient and low
 745 biomass, *Oceanography*, *4*, 56-61.
- 746 Morrison, J. M., S. Gaurin, L. A. Codispoti, T. Takahashi, F. J. Millero, W. D. Gardner, and M. J.
 747 Richardson (2001), Seasonal evolution of hydrographic properties in the Antarctic
 748 circumpolar current at 170 W during 1997–1998, *Deep-Sea Research Part II*, *48*, 3943-
 749 3972.
- 750 Munro, D. R., N. S. Lovenduski, B. B. Stephens, T. Newberger, K. R. Arrigo, T. Takahashi, P. D.
 751 Quay, J. Sprintall, N. M. Feeman, and C. Sweeney (2015), Estimates of net community
 752 production in the Southern Ocean determined from time series observations (2002-
 753 2011) of nutrients, dissolved inorganic carbon, and surface ocean pCO₂ in Drake
 754 Passage, *Deep-Sea Research II*, *114*, 49-63.
- 755 Nelson, D. M., M. A. Brzezinski, D. E. Sigmon, and V. M. Frank (2001), A seasonal progression
 756 of Si limitation in the Pacific sector of the Southern Ocean, *Deep-Sea Research II*, *48*,
 757 3973-3995.
- 758 Nicholson, D., H. R. H. Stanley, and S. C. Doney (2014), The triple oxygen isotope tracer of
 759 primary productivity in a dynamic ocean model, *Global Biogeochemical Cycles*, *28*, 538-
 760 552.
- 761 Parslow, J., P. Boyd, S. R. Rintoul, and F. B. Griffiths (2001), A persistent sub-surface
 762 chlorophyll maximum in the Polar Frontal Zone south of Australia: seasonal
 763 progression and implications for phytoplankton-light-nutrient interactions, *Journal of*
 764 *Geophysical Research*, *106*, 31543-31557.

- 765 Prokopenko, M. G., O. M. Pauluis, J. Granger, and L. Y. Yeung (2011), Exact evaluation of
 766 gross photosynthetic production from the oxygen triple-isotope composition of O₂:
 767 Implications for the net-to-gross primary production ratios, *Geophysical Research*
 768 *Letters*, *38*, DOI: 10.1029/2011GL047652.
- 769 Quay, P. D., J. Stutsman, and T. Steinhoff (2012), Primary production and carbon export
 770 rates across the subpolar N. Atlantic Ocean basin based on triple oxygen isotope and
 771 dissolved O₂ and Ar gas measurements, *Global Biogeochemical Cycles*, *26*, doi:
 772 10.1029/2010GB004003.
- 773 Reuer, M. K., B. B. Barnett, M. L. Bender, P. G. Falkowski, and M. B. Hendricks (2007), New
 774 Estimates of Southern Ocean Biological Production Rates from O₂/Ar Ratios and the
 775 Triple Isotope Composition of O₂ *Deep-Sea Research Part I*, *54*, 951-974.
- 776 Richardson, T. L., and G. A. Jackson (2007), Small Phytoplankton and Carbon Export from
 777 the Surface Ocean, *Science*, *315*, 838 - 840, DOI: 10.1126/science.1133471.
- 778 Rigual-Hernández, A. S., T. W. Trull, S. G. Bray, A. Cortina, and L. K. A. Armand (2015),
 779 Latitudinal and temporal distributions of diatom populations in the pelagic waters of
 780 the Subantarctic and Polar Frontal zones of the Southern Ocean and their role in the
 781 biological pump, *Biogeosciences*, *12*, 5309–5337.
- 782 Rintoul, S. R., and J. L. Bullister (1999), A late winter hydrographic section from Tasmania
 783 to Antarctica, *Deep-Sea Research I*, *46*, 1417-1454.
- 784 Rintoul, S. R., and T. W. Trull (2001a), Seasonal evolution of the mixed layer in the
 785 Subantarctic Zone south of Australia, *Journal of Geophysical Research-Oceans*, *106*,
 786 31447-31462.
- 787 Rintoul, S. R., and T. W. Trull (2001b), Seasonal evolution of the mixed layer in the
 788 Subantarctic Zone south of Australia, *Journal of Geophysical Research-Oceans*, *106*(C12),
 789 31447-31462.

- 790 Rintoul, S. R., J. R. Donguy, and D. H. Roemmich (1997), Seasonal evolution of upper ocean
 791 thermal structure between Tasmania and Antarctica, *Deep-Sea Research*, 44(7), 1185-
 792 1202.
- 793 Rubin, S. I. (2003), Carbon and nutrient cycling in the upper water column across the Polar
 794 Frontal Zone and Antarctic Circumpolar Current along 170°W, *Global Biogeochemical*
 795 *Cycles*, 17(3), 1087.
- 796 Ryan-Keogh, T. J., A. I. Macey, M. V. Nirldfoyyit, M. I. Lucas, S. S. Steigenberger, M. C.
 797 Stinchcombe, E. P. Achterberg, T. S. Bibby, and C. M. Moore (2013), Spatial and temporal
 798 development of phytoplankton iron stress in relation to bloom dynamics in the high-
 799 latitude North Atlantic Ocean, *Limnology and Oceanography*, 58, 2013.
- 800 Sarmiento, J. L., and J. C. Orr (1991), Three-dimensional simulations of the impact of
 801 Southern Ocean nutrient depletion on atmospheric CO₂ and ocean chemistry, *Limnology*
 802 *and Oceanography*, 36, 1928-1950.
- 803 Sarmiento, J. L., N. Gruber, M. A. Brzezinski, and J. P. Dunne (2004), High-latitude controls of
 804 thermocline nutrients and low latitude biological productivity, *Nature*, 427, 56-60.
- 805 Sedwick, P. N., A. R. Bowie, and T. W. Trull (2008), Dissolved iron in the Australian sector of
 806 the Southern Ocean (CLIVAR SR3 section): Meridional and seasonal trends, *Deep-Sea*
 807 *Research I*, 55(8), 911-925.
- 808 Sedwick, P. N., G. R. DiTullio, D. A. Hutchins, P. W. Boyd, F. B. Griffiths, A. C. Crossley, T. W.
 809 Trull, and B. Queguiner (1999), Limitation of algal growth by iron deficiency in the
 810 Australian Subantarctic region, *Geophysical Research Letters*, 26(18), 2865-2868.
- 811 Shadwick, E. H., B. Tilbrook, N. Cassar, T. W. Trull, and S. R. Rintoul (2015a), Summertime
 812 physical and biological controls on O₂ and CO₂ in the Australian Sector of the Southern
 813 Ocean, *Journal of Marine Systems*, 147, 21-28.
- 814 Shadwick, E. H., T. W. Trull, B. Tilbrook, A. J. Sutton, E. Schulz, and C. L. Sabine (2015b),
 815 Seasonality of biological and physical controls on surface ocean CO₂ from hourly

- 816 observations at the Southern Ocean Time Series site south of Australia, *Global*
817 *Biogeochemical Cycles*, 2014GB004906, 004901-004916.
- 818 Sigman, D. M., and E. A. Boyle (2000), Glacial/Interglacial variations in atmospheric carbon
819 dioxide, *Nature*, 407, 859-869.
- 820 Smith Jr, W. O., R. F. Anderson, J. Keith Moore, L. A. Codispoti, and J. M. Morrison (2000),
821 The US Southern Ocean Joint Global Ocean Flux Study: an introduction to AESOPS, *Deep*
822 *Sea Research Part II: Topical Studies in Oceanography*, 47(15–16), 3073-3093.
- 823 Smith, W. O., and C. Lancelot (2004), Bottom-up versus top-down control in phytoplankton
824 of the Southern Ocean, *Antarctic science*, 16(04), 531-539.
- 825 Sokolov, S. (2008), Chlorophyll blooms in the Antarctic Zone south of Australia and New
826 Zealand in reference to the Antarctic Circumpolar Current fronts and sea ice forcing,
827 *Journal of Geophysical Research: Oceans*, 113(C3), C03022.
- 828 Sokolov, S., and S. R. Rintoul (2002), Structure of Southern Ocean fronts at 140E, *Journal of*
829 *Marine Systems*, 37(1-3), 151-184.
- 830 Sokolov, S., and S. R. Rintoul (2009), Circumpolar structure and distribution of the Antarctic
831 Circumpolar Current fronts: 1. Mean circumpolar paths, *Journal of geophysical research*,
832 114(C11), C11018.
- 833 Sosik, H. M., and R. J. Olson (2002), Phytoplankton and iron limitation of photosynthetic
834 efficiency in the Southern Ocean during late summer, *Deep-Sea Research Part II*, 49,
835 1195-1216.
- 836 Steinberg, D. K., B. A. Van Mooy, K. O. Buesseler, P. W. Boyd, T. Kobari, and D. M. Karl
837 (2008), Bacterial vs. zooplankton control of sinking particle flux in the ocean's twilight
838 zone, *Limnology and Oceanography*, 53(4), 1327-1338.
- 839 Strzepek, R. F., K. A. Hunter, R. D. Frew, P. J. Harrison, and P. W. Boyd (2012), Iron-light
840 interactions differ in Southern Ocean phytoplankton, *Limnology and oceanography*,
841 57(4), 1182-1200.

- 842 Sunda, W. G., and S. A. Huntsman (1997), Interrelated influence of iron, light and cell size on
 843 marine phytoplankton growth, *Nature*, 390, 389-392.
- 844 Szeto, M., P. J. Werdell, T. S. Moore, and J. W. Campbell (2011), Are the world's oceans
 845 optically different?, *Journal of Geophysical Research*, 116, doi: 10.1029/2011JC007230.
- 846 Takeda, S. (1998), Influence of iron availability on nutrient consumption ratio of diatoms in
 847 oceanic waters, *Nature*, 393(6687), 774-777.
- 848 Treguer, P., and G. Jacques (1992), Dynamics of nutrients and phytoplankton, and fluxes of
 849 carbon, nitrogen, and silicon in the Antarctic ocean, *Polar Biology*, 12, 149-162.
- 850 Trull, T. W., S. R. Rintoul, M. Hadfield, and E. R. Abraham (2001a), Circulation and seasonal
 851 evolution of polar waters south of Australia: Implications for iron fertilization of the
 852 Southern Ocean, *Deep-Sea Research Part II-Topical Studies in Oceanography*, 48(11-12),
 853 2439-2466.
- 854 Trull, T. W., P. N. Sedwick, F. B. Griffiths, and S. R. Rintoul (2001b), Introduction to special
 855 section: SAZ Project, *Journal of Geophysical Research-Oceans*, 106(C12), 31425-31429.
- 856 Trull, T. W., S. G. Bray, S. J. Manganini, S. Honjo, and R. François (2001c), Moored sediment
 857 trap measurements of carbon export in the Subantarctic and Polar Frontal Zones of the
 858 Southern Ocean, south of Australia, *Journal of Geophysical Research*, 106(C12), 31489-
 859 31510.
- 860 Wang, X., and R. J. Matear (2001), Modeling the upper ocean dynamics in the Subantarctic
 861 and Polar Frontal Zones in the Australian sector of the Southern Ocean, *Journal of*
 862 *Geophysical Research*, 106(C12), 31511-31524.
- 863 Wang, X., R. J. Matear, and T. W. Trull (2001), Modeling seasonal phosphate export and
 864 resupply in the Subantarctic and Polar Frontal Zones in the Australian sector of the
 865 Southern Ocean, *Journal of Geophysical Research*, 106(C12), 31525-31542.
- 866 Weeding, B., and T. W. Trull (2014a), Hourly oxygen and total gas tension measurements at
 867 the Southern Ocean Time Series site reveal winter ventilation and spring net
 868 community production, *Journal of Geophysical Research: Oceans*, 119(1), 348-358.

869 Weeding, B., and T. W. Trull (2014b), Hourly oxygen and total gas tension measurements at
870 the Southern Ocean Time Series site reveal winter ventilation and spring net
871 community production, *Journal of Geophysical Research - Oceans*, 119, 348-358,
872 doi:310.1002/2013JC009302, 002014.

873 Westwood, K. J., F. B. Griffiths, J. P. Webb, and S. W. Wright (2011), Primary production in
874 the Sub-Antarctic and Polar Frontal Zones south of Tasmania, Australia; SAZ-Sense
875 survey, 2007, *Deep-Sea Research II*, 58, 2162-2178.

876

877

878

Figures

Fig. 1. Map of the study region showing the location of each water sample.

Fig. 2. $[\text{SiO}_2]$ and $[\text{NO}_3^-]$ vs. latitude south of Tasmania during the summer half-years of 2006-2007 (top) and 2007-2008 (bottom). Red is October/November, green is January, and blue is February/March. Also plotted is $[\text{NO}_3^-]$ vs. salinity.

Fig. 3. Hovmoller diagrams showing net O_2 bioflux (sea-air biological O_2 flux, $\text{mmol m}^{-2} \text{ day}^{-1}$), MLD, surface PAR, chlorophyll concentration on a log scale ($\mu\text{g/L}$), euphotic depth, $\log(\text{chl})$, and depth-averaged PAR vs. time for a longitude band of ~ 30 degrees south of Australia.

Fig. 4. Left: Net and gross bioflux vs. latitude for all samples. Right: Chlorophyll-normalized net and gross bioflux vs. latitude for all samples. Error bars correspond to standard errors. Upper part of the plot shows the number of samples for each degree of latitude.

Fig. 5. Average net (blue) or gross (green) O_2 bioflux vs. month of observation for the entire domain.

Fig. 6. Net (left) and gross (right) O_2 volumetric bioflux normalized to chlorophyll, plotted vs. average mixed layer irradiance.

Fig. 7. Relative production vs. average daily irradiance in the mixed layer. Production is assumed to increase linearly up to saturating irradiance (taken as $56 \mu\text{mol photons m}^{-2} \text{ sec}^{-1}$), and constant at higher irradiances. This plot shows that daily production increases linearly with average mixed layer irradiance up to irradiance values much higher than the instantaneous saturating irradiance, $3.2 \text{ mol m}^{-2} \text{ day}^{-1}$. The reason is that irradiance early in the day, late in the day, and lower in the mixed layer can be undersaturated even when average daily ML irradiance is much greater than the instantaneous saturating value.

Fig. 1. Map of the study region showing the location of each water sample.

Fig. 2. $[\text{SiO}_2]$ and $[\text{NO}_3^-]$ vs. latitude south of Tasmania during the summer half-years of 2006-2007 (top) and 2007-2008 (bottom). Red is October/November, green is January, and blue is February/March. Also plotted is $[\text{NO}_3^-]$ vs. salinity.

Fig. 3. Hovmoller diagrams showing net O₂ bioflux (sea-air biological O₂ flux, mmol m⁻² day⁻¹), MLD, surface PAR, chlorophyll concentration on a log scale (μg/L), euphotic depth, log(chl), and depth-averaged PAR vs. time for a longitude band of ~ 30 degrees south of Australia. Chl concentrations are from remote sensing, and MLD is from Argo floats.

Fig. 4. Left: Net and gross bioflux vs. latitude for all samples. Right: Chlorophyll-normalized net and gross production vs. latitude for all samples. Error bars correspond to standard errors. Upper part of the plot shows the number of samples for each degree of latitude.

Fig. 5. Average net and gross bioflux for the major zones of the Southern Ocean south of Australia as a function of the month of sampling. STZ, Subtropical Zone (north of 46° S), SAZ, Subantarctic Zone (52°-46° S); PFZ, Polar Frontal Zone (56°-52° S); AZ, Antarctic Zone (63°-56° S), and SACC, South of the Antarctic Circumpolar Current (S of 63° S).

Fig. 6. Net (left) and gross (right) O₂ volumetric bioflux normalized to chlorophyll, plotted vs. average mixed layer irradiance. Net and gross O₂ bioflux are calculated from mixed layer data and approximate O₂ NCP and O₂ GPP in the mixed layer. Normalized net bioflux is plotted only for samples where the biological O₂ supersaturation is > -0.3% (within uncertainty of zero).

Fig. 7. Relative production vs. average daily irradiance in the mixed layer. Production is assumed to increase linearly up to saturating irradiance (taken as $56 \mu\text{moles photons m}^{-2} \text{sec}^{-1}$), and constant at higher irradiances. This plot shows that daily production increases linearly with average mixed layer irradiance up to irradiance values much higher than the instantaneous saturating irradiance, $3.2 \text{ moles m}^{-2} \text{day}^{-1}$. The reason is that irradiance early in the day, late in the day, and lower in the mixed layer can be undersaturated even when average daily ML irradiance is much greater than the instantaneous saturating value.