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TELLUS

Changes of the oceanic CO_2 sink in the Eastern Indian sector of the Southern Ocean

By EMILY BRÉVIÈRE[1,2,*], NICOLAS METZL[2], ALAIN POISSON[2] and BRONTE TILBROOK[3], [1]*SOLAS Project Office, School of Environmental Sciences, University of East Anglia, Norwich, Norfolk NR4 7TJ, United Kingdom;* [2]*Laboratoire d'Océanographie et du Climat: Expérimentations et Approches Numériques (LOCEAN/IPSL), Université Pierre et Marie Curie, Paris, France;* [3]*CSIRO Marine and Atmospheric Research and Antarctic Climate and Ecosystems CRC, Hobart 7001, Australia*

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ABSTRACT

Changes in the carbon dioxide fugacity (fCO_2) and air–sea CO_2 flux observed in the Southern Ocean, south of Tasmania were analysed and compared for two different years: 1996/1997 and 2002/2003. The CO_2 flux showed large and contrasting interannual changes in the permanent open ocean zone (POOZ, 53–61°S) between the 2 yr where the oceanic CO_2 sink increased from about -0.3 mmol m^{-2} d^{-1} in February 1997 to -20.6 mmol m^{-2} d^{-1} in February 2003. The strong sink in February 2003 was associated with increased phytoplankton biomass in this High-Nutrient, Low-Chlorophyll (HNLC) region. Three hypotheses that may have influenced the biomass and fCO_2 changes in the POOZ were investigated: sea surface temperature (SST) and El Niño/ Southern Oscillation (ENSO) event, total stratospheric ozone column and ultraviolet (UV) radiation, and atmospheric dust inputs. The strong CO_2 sink in 2003 in the POOZ cannot be explained by the observed changes in SST or UV, but would be qualitatively consistent with the presence of episodic atmospheric dust inputs.

1. Introduction

To reach a more realistic estimate of the global carbon budget, it is essential to have a better understanding of the temporal variations of the oceanic carbon cycle at the global scale. Recent models indicate that the Southern Ocean is very sensitive to climate change and its study is crucial to reliably predict changes in the ocean carbon cycle in the coming decades. In the Southern Ocean, especially south of 50°S, most data are collected in summer and show the carbon dioxide fugacity (fCO_2) is highly variable, creating a mosaic of CO_2 sources and sinks, the intensity and location of which evolve with time (e.g. Metzl et al., 1991). The magnitude and interannual variability of the air–sea CO_2 flux are controversial. Different magnitudes of the fluxes are noted between the global ocean circulation models (Orr et al., 2001), the climatology (Takahashi et al., 2002) and atmospheric (Gurney et al., 2002, Roy et al., 2003, Gurney et al., 2004) or oceanic (Gloor et al., 2003, Mikaloff Fletcher et al., 2006) inverse models. The lack of long-term monitoring data leads to large carbon budget uncertainties.

*Corresponding author.
e-mail: e.breviere@uea.ac.uk
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For observing interannual oceanic biogeochemical variations, including CO_2 fluxes, repeat cruises to the same regions over several years are necessary. Long-term time-series monitoring of fCO_2 was set up in the Indian and Southern Ocean, in the frame of the ongoing programs MINERVE and OISO. In 2002/2003, transects between Hobart (Tasmania, Australia) and Dumont D'Urville (Adelie Land, Antarctica) were sampled from the end of austral winter (October) to late austral summer (March) (Fig. 1). This paper investigates the variability of air–sea CO_2 fugacity along transects and compares the spring–summer changes with observations made along the same track in 1996/1997. The variability of the air–sea CO_2 fluxes, and potential causes of large interannual changes in the air–sea fluxes, between the 1996/1997 and 2002/2003 summer periods are assessed.

2. Methods

The region covered by this study extended from 43°S to 67°S, in the vicinity of 145°E, south of Tasmania. Observations of sea surface fCO_2, temperature and salinity were carried out onboard the *RSV Astrolabe* during three cruises between Hobart and Dumont D'Urville: two cruises (October 1996 and February 1997) and one cruise in October 2002. The same measurements were also carried out onboard the *RSV Marion-Dufresne* during a

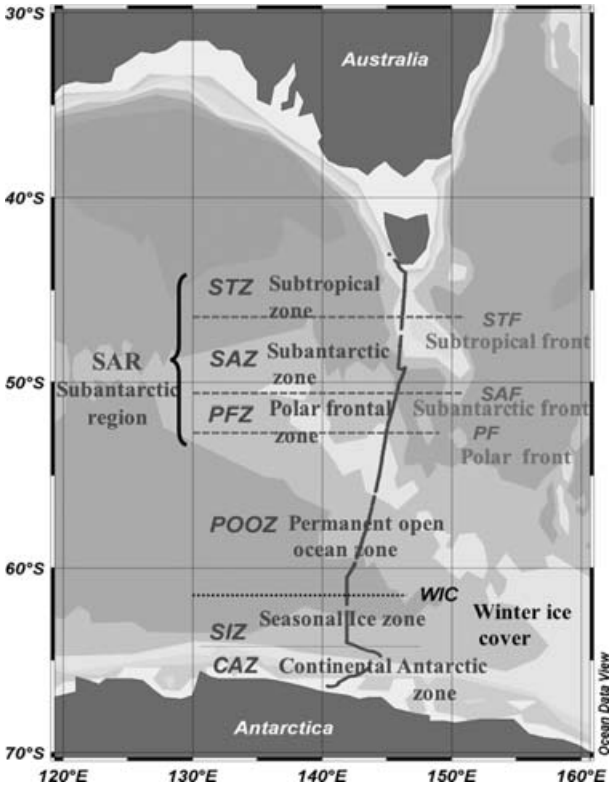

Fig. 1. Map of the transect south of Tasmania. The mean location of the major oceanic fronts and remarkable features delimits the studied provinces.

complementary cruise (OISO-10) on the same transect in February 2003.

During all cruises, sea surface temperature (SST) and sea surface salinity (SSS) were measured continuously with a Seabird thermosalinometer SBE 21 calibrated with regular salinity samples. The sea surface continuous fCO_2 measurement technique has been previously described (Poisson et al., 1993; Metzl et al., 1995, 1999; Jabaud-Jan et al., 2004). fCO_2 in seawater was measured continuously using infrared gas analysers calibrated against CO_2 in air standards analysed every 6 hr. The CSIRO system on the *Astrolabe* in 2002 used a 'shower type' equilibrator with all gases dried prior to measurement using a LICOR 6252. The LBCM/IPSL system was used on all other cruises (MINERVE in 1996/1997 and OISO in 2003); the seawater was equilibrated in a 'thin film' type equilibrator, the gas dried through Peltier cold-traps and analysed using an ULTRAMAT-5F (Siemens) NDIR. The measurements made on dry gases were corrected for water vapour at the equilibrator temperature using Weiss and Price (1980). The fCO_2 values measured at the equilibrator temperature were corrected to *in situ* temperature using the formula proposed by Copin-Montegut (1988, 1989). For all cruises the warming between the equilibrator water temperature and *in situ* temperature were less than 1°C. An international at-sea fCO_2 intercomparisons study conducted in May 1996 onboard the *RV Meteor* showed the CSIRO and LBCM/IPSL systems agree to ± 1 μatm (Körtzinger et al., 2000). The oceanic fCO_2 data are considered accurate to ± 1 μatm.

The net CO_2 flux (F) across the air–sea interface was determined from:

$$F = ks\Delta fCO_2,$$

where k, the gas transfer velocity, was evaluated using the cubic wind speed relationship of Wanninkhof and McGillis (1999); s, the solubility of CO_2 in seawater at the *in situ* temperature and salinity was calculated using the algorithm of Weiss (1974); ΔfCO_2 was the difference in fCO_2 in surface seawater (fCO_2^{sw}) and in the atmosphere (fCO_2^{air}):

$$\Delta fCO_2 = fCO_2^{sw} - fCO_2^{air}.$$

The ΔfCO_2 sets the direction of CO_2 gas exchange and is controlled by complex physical, chemical and biological interactions in the ocean that determine the fCO_2 of seawater.

ΔfCO_2 used in the flux calculation was obtained by averaging ΔfCO_2 values for each cruise (northbound and southbound) for the track segments within each of four biogeochemical regions described below. We used wind speed from ERS-2 and QuikSCAT MWF products, which provide a monthly satellite wind field over a global grid with 0.5° × 0.5° resolution (http://www.ifremer.fr/cersat/en/index.htm). These data were obtained from CERSAT, at IFREMER, Plouzané (France).

3. Regional and seasonal variability

The tracks of the cruises crossed several hydrological fronts and we used criteria based on sea surface salinity in the same region (Chaigneau and Morrow, 2002) to locate the fronts (Fig. 1). The sub-tropical front (STF), the sub-Antarctic front (SAF) and the polar front (PF) are found in the northern part of the transect. In the south, hydrological gradients were observed near the ice and in the vicinity of the Antarctic Divergence. The northernmost limit of the winter ice cover (WIC) was located near 61°S (http://www.natice.noaa.gov/). The fronts, the WIC and the Antarctic Divergence near 64°S delimit different biogeochemical regions (Fig. 1):

(1) The sub-Antarctic region, SAR, located at 43–54°S in 1996/1997 and 43–52.5°S in 2002/2003.

(2) The permanent open ocean zone, POOZ, at 54–61°S in 1996/1997 and 52.5–61.5°S in 2002/2003.

(3) The seasonal ice zone, SIZ, at 61–64.5°S in 1996/1997 and 61.5–64.5°S in 2002/2003.

(4) The continental Antarctic zone, CAZ, at 64.5–67°S.

The distributions of ΔfCO_2, during the cruises of spring–summer 1996/1997 and 2002/2003 are presented Fig. 2. The seasonal ΔfCO_2 patterns are clearly different in each biogeochemical region and described below for the SAR, the POOZ and the region south of 62°S.

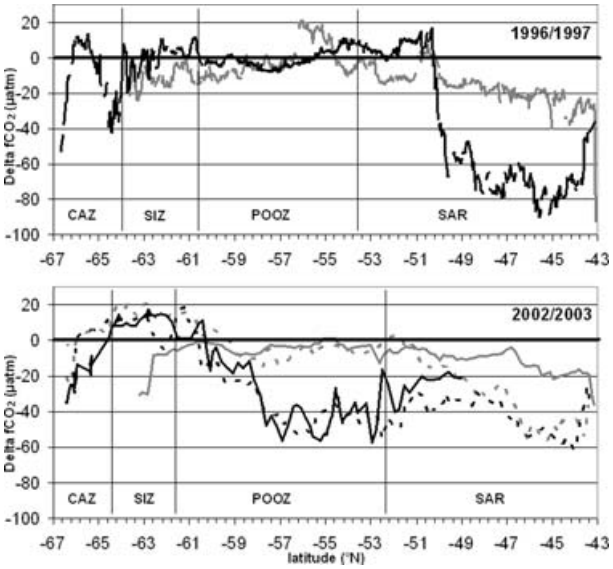

Fig. 2. Seasonal distributions of ΔfCO_2 south of Tasmania in 1996/1997 and 2002/2003. October (grey solid line), late December–January (grey dashed line), early February (black solid line) and late February (black dashed line). Regions are also presented.

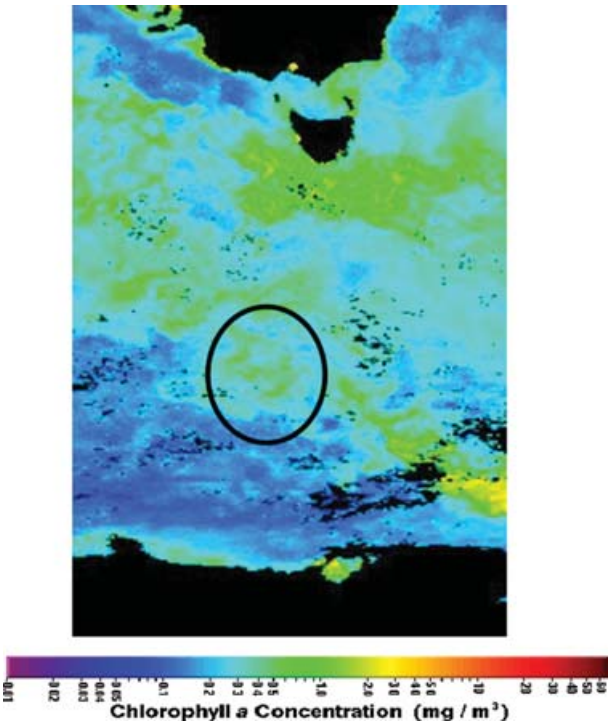

Fig. 3. SeaWiFS distribution of monthly mean chlorophyll-*a* (mg m^{-3}) in February 2003. The ellipse encircles the POOZ, where the phytoplankton biomass increased.

3.1. The sub-Antarctic region

The surface ocean in the SAR was undersaturated in CO_2, with respect to atmospheric CO_2, during both years sampled. A strong CO_2 sink occurred north of the SAF, which was located at 50°S in 1996/1997 and 49.5°S in 2002/2003. North of the SAF, the sink was always more intense in summer than in spring. A large seasonal fCO_2 decrease, up to 50 μatm below atmospheric values, was observed around 45°S in 1996/1997 (Fig. 2, upper panel), with a maximum decrease of about 40 μatm was observed at the same latitude in 2002/2003 (Fig. 2, lower panel). South of the SAF, in the polar front zone (PFZ), the ΔfCO_2 seasonality was variable, sometimes positive (spring to summer increase of about +10 μatm in 1996/1997), sometimes negative (spring to summer decrease in 2002/2003).

On average the whole SAR's surface waters were an intense CO_2 sink in February 1997 and 2003, slightly larger in 2003 than in 1997 (−13.6 mmol m^{-2} d^{-1} and −12.3 mmol m^{-2} d^{-1}, respectively, Table 1). These sinks are associated with an increase of phytoplankton biomass in summer as shown on the SeaWiFS data of February 2003, where high chlorophyll-*a* concentrations up to 1.2 μg/L occurred just south of Tasmania (Fig. 3). The spring–summer cycle of the CO_2 fluxes in the SAR in February 1997 and 2003 exhibits a pattern already noticed south of Tasmania (Inoue et al., 1999; Metzl et al., 1999; Popp et al., 1999; Ishii et al., 2002; Inoue and Ishii, 2005). The SAR displayed no significant interannual variability of the CO_2 fluxes compared to south of the Polar Front.

3.2. The permanent open ocean zone

The surface water and atmosphere were near equilibrium in the spring and summer of both years for the southern POOZ waters

Table 1. Averaged ΔfCO_2 and air–sea CO_2 fluxes in the four different regions of the Southern Ocean in October 1996 and 2002 and February 1997 and 2003. Numbers in brackets indicate standard deviations.

	SAR		POOZ		SIZ		**CAZ**	
	Delta fCO_2 (μatm)	Flux (mM m^{-2} d^{-1})	Delta fCO_2 (μatm)	Flux (mM m^{-2} d^{-1})	Delta fCO_2 (μatm)	Flux (mM m^{-2} d^{-1})	Delta fCO_2 (μatm)	Flux (mM m^{-2} d^{-1})
October 1996	−13.6 (10)	−9.0	−4.6 (9)	−3.3	−8.3 (6)	−4.8		
February 1997	−34.3 (37)	−12.3	−0.5 (5)	−0.3	−6.4 (15)	−1.6	−6.4 (20)	−3.1
October 2002	−10.6 (11)	−5	3.9 (8)	1.4	−12.4 (10)	−5.0		
February 2003	−31.2 (13)	−13.6	−35.4 (13)	−20.6	13.3 (3)	2.6	−7.5 (16)	−4.7

between 58°S and 61.5°S. In the northern part of the POOZ, ΔfCO_2 was close to 0 μatm in spring of both years and the summer of 1997. Large ΔfCO_2 changes were found in summer 2003 (Table 1). In February 2003, at latitude 52.5°S–58°S, ΔfCO_2 was about -45 μatm below the atmospheric equilibrium values. Here, the oceanic fCO_2 decreased rapidly between late January 2003 and early February 2003. The intense CO_2 sink in February 2003 (-20.5 mmol m^{-2} d^{-1}) is associated with increased phytoplankton biomass estimated from satellite derived sea surface chlorophyll-*a* concentrations in February (http://seawifs.gsfc.nasa.gov/cgi/level3.pl; ellipse on Fig. 3).

Previous analysis based on December–January data suggest the POOZ is typically a small sink (ΔfCO_2 of about -10 μatm) or close to equilibrium, with a small spring–summer seasonal variability (Inoue and Ishii, 2005). This is in good agreement with our observations of 1996/1997 and similar to climatological analysis (Takahashi et al., 2002), highlighting the dramatic ocean CO_2 sink anomaly we observed in austral summer 2003.

3.3. The region south of 62°S: seasonal ice zone and continental Antarctic zone

When approaching the ice-edge and Antarctic coastal regions, the ΔfCO_2 is highly variable and a clear seasonal signal is difficult to establish at the regional scale. The surface water of the SIZ displayed a negative ΔfCO_2 in spring 1996 and 2002 of -8.3 and -12.4 μatm, respectively. In summer, ΔfCO_2 was negative in 1997 (-6.4 μatm) and positive in 2003 (13.3 μatm). In the CAZ, ΔfCO_2 was always negative during the summer months, with slightly larger values in 2003 (Table 1). The presence of CO_2 sinks located close to the sea-ice edge and the continent are likely to be controlled by increased phytoplankton biomass. The biomass typically shows a high mesoscale variability and this is reflected in large ΔfCO_2 variability around the mean (Table 1).

4. Large change of the CO_2 sink in the POOZ

Figure 4 shows the calculated mean air–sea CO_2 fluxes over the different regions in February 1997 and 2003. The largest change observed in the calculated air–sea CO_2 fluxes was between February 1997 and 2003 in the POOZ (Fig. 4). The different features in the CO_2 flux patterns in the POOZ of the Southern Ocean highlight the need to progress understanding of the mechanisms driving such variability.

For the Southern Ocean, both temperature and biological activity cause significant seasonal changes in surface CO_2 (Takahashi et al., 2002) and could influence interannual variability. The Southern Ocean is the largest High-Nutrient, Low-Chlorophyll (HNLC) area where phytoplankton does not usually exhaust all the available macronutrients in surface waters (Bathmann et al., 1997; Smetacek et al., 1997). Despite the active debates of the last decade, the paradox of the HNLC regions remains and the

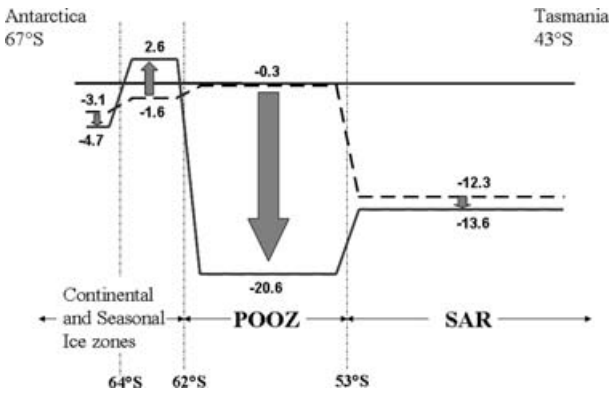

Fig. 4. Diagram depicting the air–sea CO_2 fluxes (mM m^{-2} d^{-1}) over the four distinct zones in February 1997 (black dashed line) and 2003 (black solid line).

balance of processes that drive variability in the carbon cycle are obscure. Macronutrient concentrations are rarely limiting and trace metal limitation, especially iron (De Baar et al., 1995), incident solar irradiance (Lubin et al., 2004), total stratospheric ozone column (Arrigo et al., 2003) and surface temperature (Fiala et al., 1998), may contribute towards direct or indirect limiting phytoplankton growth (Löscher et al., 1997; Hutchins et al., 2001; Marchant et al., 2001; Leblanc et al., 2002) and ocean carbon sink variability (Jabaud-Jan et al., 2004). In the following sections, we investigate how these factors, temperature and mixed layer depth may have influenced the fCO_2 changes observed in the POOZ region.

4.1. Mixed layer depth and temperature effects

Interannual changes in mixed layer depth could alter the surface CO_2 through a number of processes including variations in the entrainment of CO_2-rich deeper waters into the surface mixed layer, enhanced surface stratification and warming, and changes to the light and nutrient regimes that influence biological production. Chaigneau et al. (2004) developed a climatology of mixed layer depth south of Tasmania and found surface waters of the POOZ are well-mixed down to about 150 m depth in winter, and about 40–60 m in summer. They found the mixed layer depths (MLD) were fairly constant over the zone. Table 2 presents the mean MLD in the POOZ in October 1996, February 1997 and October 2002 estimated from the expendable bathythermograph (XBT) profiles of the SURVeillance de l'Océan auSTRAL program (SURVOSTRAL) obtained from the Coriolis website (http://www.coriolis.eu.org/) and, from the hydrographic stations done during the cruise of February 2003 (OISO-10). The MLD in summers 1997 and 2003 were similar and close to the climatological MLD of Chaigneau et al. (2004). The mean monthly wind speed over the POOZ in February 1997 was about 10.6 m s^{-1} and, in February 2003, was about 10.1 m s^{-1} (ERS-2 and QuikSCAT MWF products). The

Table 2. MLD in the POOZ estimated from expendable bathythermograph (XBT) profiles of the SURVOSTRAL program for October 1996, 2002 and February 1997 and from hydrographic stations profiles done during the cruise of February 2003 (OISO-10). Climatological MLD of the Northern Antarctic zone (54°S–59.5°S) (Chaigneau et al., 2004)

	23–25 October 1996	28 January 1997[a]	31 October 2002	1–2 February 2003
MLD (m)	140	50–60	100	40–50
	October	January	October	February
Climatological MLD (m)	115	50	115	55

[a]Two weeks before the cruise.

Table 3. Mean sea surface temperature (°C) over the POOZ in October 1996 and 2002 and February 1997 and 2003. Numbers in brackets indicate standard deviations

	1996/1997	2002/2003
October	0.9 (1.1)	2.2 (1.0)
February	4.7 (0.6)	5.2 (0.4)

similarity in wind speeds and MLD between years suggests that changes in upper ocean mixing and gas exchange rates cannot explain the observed variability in surface fCO_2. Takahashi et al. (1993) measured the temperature effect on pCO_2 in isochemical conditions ($\partial \ln pCO_2/\partial T$) as $+4.23°C^{-1}$. Surface seawater of the POOZ was warmed from October 1996 to February 1997 by about 4°C (Table 3), which would have increased fCO_2 by about 60 μatm. Measurements show fCO_2 increased by less than 5 μatm for the period. The seasonal variation of a surface warming of about 3°C between October 2002 and February 2003 would have increased fCO_2 by about 50 μatm, but the measured fCO_2 decreased by about 40 μatm. While temperature effects will influence surface fCO_2 they cannot explain the large CO_2 sink observed in February 2003 in the POOZ when compared to the 1996/1997 data. Other processes, such as enhanced biological carbon uptake in the summer of 2002/2003 appear to dominate thermodynamic and mixed-layer influences on the development of the large CO_2 sink in the POOZ in 2002/2003.

4.2. *Warm SST anomaly and El Niño/Southern Oscillation (ENSO) event*

Increased temperature in the POOZ region has been linked with enhanced microalgal growth rates (Fiala et al., 1998), and could affect ocean fCO_2. Such a scenario for sea surface fCO_2 was

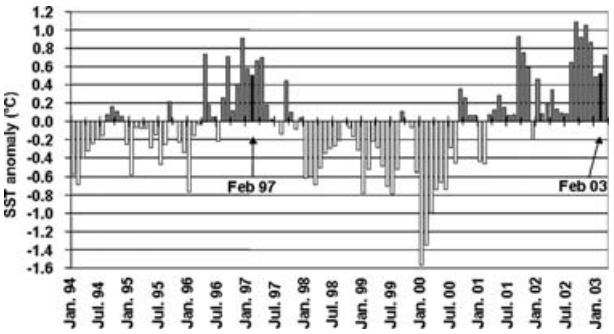

Fig. 5. Mean SST anomalies in the POOZ (145°E; 52.5°–60.5°S) from January 1994 to March 2003, Reynolds and Smith (1994).

analysed in the western Indian POOZ (50–57°S, 65°E) where a large CO_2 sink (oceanic $fCO_2 < 330$ μatm) was observed during a warm period (Jabaud-Jan et al., 2004). The authors attributed the strong sink in summer 1998 to an increase in the biological activity (lower fCO_2) that dominated the thermodynamic effect of warming (higher fCO_2). The high productivity in the Indian POOZ, was associated with low silicic acid concentrations (<2 μmol kg^{-1}), but they were not able to identify a mechanism to explain the high productivity. The analysis of the SST anomalies in the eastern Indian POOZ also reveals the presence of cold and warm anomalies of about 1°C over the period 1994–2004 (Fig. 5). In both February 1997 and 2003, the warm anomaly was about +0.5°C, yet the strong CO_2 sink was observed in February 2003, but no sink was observed in February 1997. Thus, a warming event in the POOZ is not always associated with an enhanced ocean CO_2 sink.

Jabaud-Jan et al. (2004) did suggest a relationship between warming in high latitudes, air–sea CO_2 flux anomalies, and the 1997/1998 ENSO event through an atmospheric connection (e.g. change in clouds and light that could also affect primary production). A moderate ENSO event was noticed in 2002/2003 and not in 1996/1997. The mechanisms by which remote ENSO events can influence Southern Ocean net primary production are unknown, but examples might include ENSO-related drought conditions over continents, such as over Australia, that could supply iron in dust and enhance primary production (see Section 4.4).

4.3. *Total stratospheric ozone column and ultraviolet (UV) radiation*

Despite numerous studies, the affects of the high energy part of the UV spectrum (UVB and UVC <315 nm) on phytoplankton remains controversial and complex. Positive and negative effects of UV radiation on biogeochemical cycles in the upper ocean have been suggested. The harmful effects on phytoplankton growth (Smith et al., 1992; Neale et al., 1998; Helbling et al., 2001; Larsen, 2005) are well documented. UV radiation could also modify the bioavailability of dissolved organic matter (Sommaruga et al., 1997) virus development (Lindell et al.,

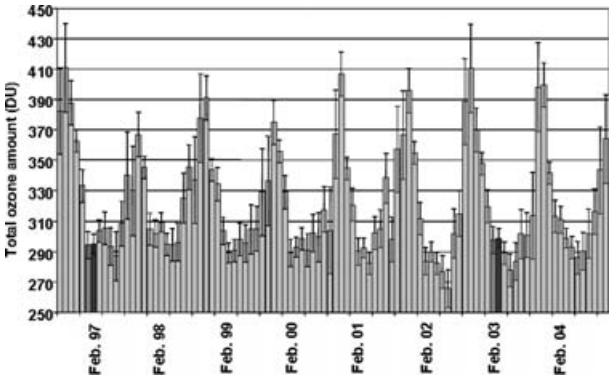

Fig. 6. Mean total ozone in DU over the POOZ (143–147.5°E; 52.5°–58°S) from August 1996 to October 2004. Total Ozone Mapping Spectrometer satellite data (http://jwocky.gsfc.nasa.gov/).

1996), photochemical production of toxic compounds (copper and aluminium, superoxide radicals and hydrogen peroxide; Karentz et al., 1994), and iron speciation (Rijkenberg et al., 2004). These processes all have potential to influence biological activity and associated carbon uptake in the surface waters.

Ultraviolet radiation is strongly influenced by stratospheric ozone. Total ozone (the integral of the ozone concentration with respect to height) changes in the atmosphere are likely to influence UV radiation reaching the sea surface and could influence the net primary production. Was the amount of UV radiation reaching the sea surface ocean different between the summers 1997 and 2003 leading to a different net primary production between the two summers?

Total stratospheric ozone data were obtained from the Total Ozone Mapping Spectrometer satellite (TOMS/MW data/NOAA—http://jwocky.gsfc.nasa.gov/). Over the studied area (52.5°–58°S–143–147.5°E) the monthly mean of the total ozone did not present significant interannual variability especially in summer (Fig. 6). Although some variations were identified during austral winter (e.g. a minimum in September–October 1997 and September 2004), the summer concentrations (around 300 Dobson units, DU) were stable over the investigated period, 1996–2004. During the period of the cruises, the total ozone were 294 ± 9 and 295 ± 7 DU in January and February 1997, respectively. In January and February 2003, the total ozone was 298 ± 9 and 299 ± 7 DU, respectively. The similarity in total ozone between years suggests a similar impact of UV radiation for the both years, making it an unlikely contributor to variability of the phytoplankton biomass and sea surface fCO_2.

4.4. Dust storms and iron input

The availability of iron is a factor that may contribute to phytoplankton production and associated CO_2 uptake. Deckboard and iron addition experiments in the surface water of HNLC systems show iron, and possibly other trace nutrients, as limiting phytoplankton growth in HNLC waters of the POOZ (De Baar et al., 1995; Sedwick et al., 1997, 1999; Boyd et al., 2000, 2004, 2005; Jickells et al., 2005). Sohrin et al. (2000), suggested low iron concentrations in the study area appears to be a limiting factor of phytoplankton growth. Surface inputs of trace nutrients by upwelling of deeper waters (Tyrrell et al., 2005), melting of sea-ice and icebergs, or by atmospheric dust events tend to be small in the POOZ (Löscher et al., 1997; Sohrin et al., 2000).

Meteorological records show that since 1960 dust storms in Australia are highly episodic events (McTainsh et al., 2005). Severe drought conditions in eastern Australia produced a very active period of dust storms in the summer of 2002/2003 compared to 1996/1997 (McTainsh et al., 2005). Gabric et al. (2002) show a correlation between high phytoplankton biomass in summer for the SAR region and aerosol optical depth, indicating the high biomass could be related to dust inputs from Australia.

There are no dust monitoring sites south of Tasmania in the POOZ, but the HYSPLIT model (Draxler and Rolph, 2003) (http://www.arl.noaa.gov/ss/models/hysplit.html) provides backward air mass trajectories. The ship was sailing towards Antarctica and in the POOZ from 31 January to 2 February 2003 when the CO_2 sink was measured. The increase of the phytoplankton biomass is visible in SeaWiFS images for the week 2–9 February 2003, not before. Air-mass back trajectories were analysed from 24 January to 2 February 2003 to account for a 5–10 d lag observed in phytoplankton response to experimental releases of iron to surface waters of the Southern Ocean (Boyd et al., 2001; Coale et al., 2004). The provenance of hypothetical air parcels arriving in the POOZ for the period was primarily marine from polar/subpolar regions west (Figs. 7a and b), or from near Antarctica (Fig. 7c). The analysis shows air masses from the southern Australian region were transported to the POOZ on 24 and 25 January 2003 (Fig. 7d). A TOMS aerosol index map (Fig. 8) also shows elevated aerosol concentrations in the southeastern Australian region on the 24 and 25 January 2003. There are a wide range of air mass trajectories, but dust deposition into the POOZ from air near the Australian continent was possible at this time and may have been a source of iron that led to increased phytoplankton biomass and CO_2 drawdown in February 2003.

5. Conclusion

Surface CO_2 data collected south of Tasmania show the surface waters of the SAR were an intense CO_2 sink in summer 1997 and 2003, with similar uptake in both years. The observed changes are associated with increased biomass and are consistent with other studies in the region that attribute strong summertime drawdown in surface CO_2 to increased biological production. South of 62°S, in the sea-ice zone the sign of the CO_2 flux was variable and showed no definite change for both years. Large changes were observed in the POOZ between the summer months of 1997 and 2003. In this HNLC region, an unusual intense CO_2 sink was observed in summer 2003 compared to

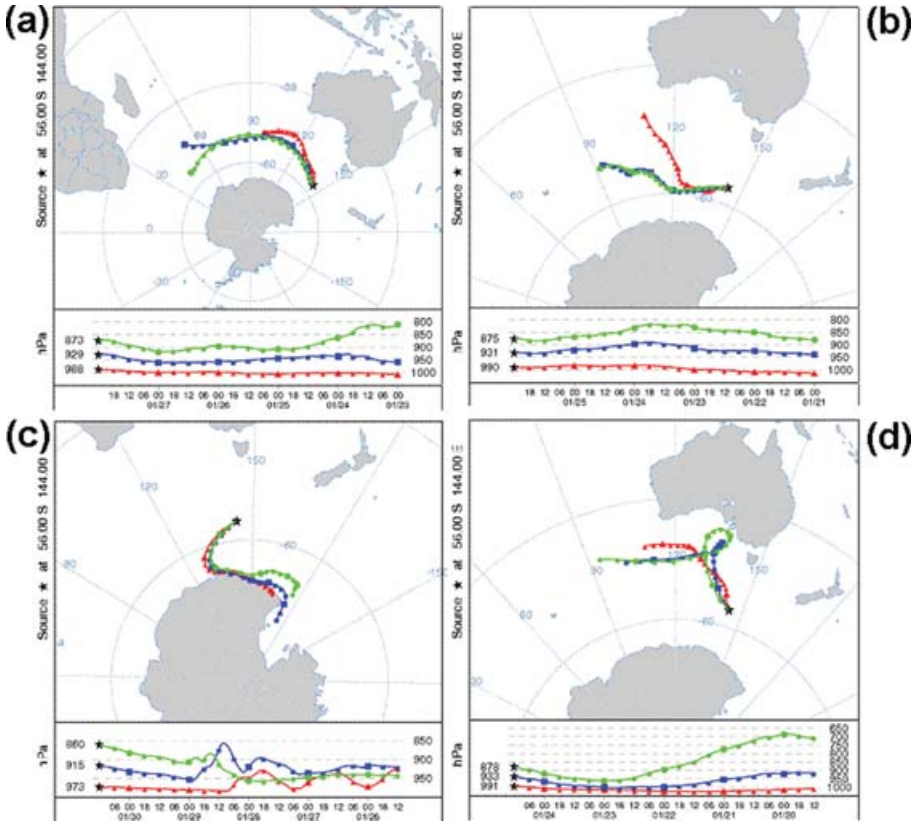

Fig. 7. Air mass back trajectories for a site in the POOZ (56°S, 144°E) in late January 2003. Trajectory: (a) and (b) polar/subpolar marine; (c) polar marine, Antarctica and (d) Australian coastal. The upper part of each panel denotes the trajectory routes based on heights of 10 500 and 1000 m, and the lower part shows the predicted trajectory heights overtime.

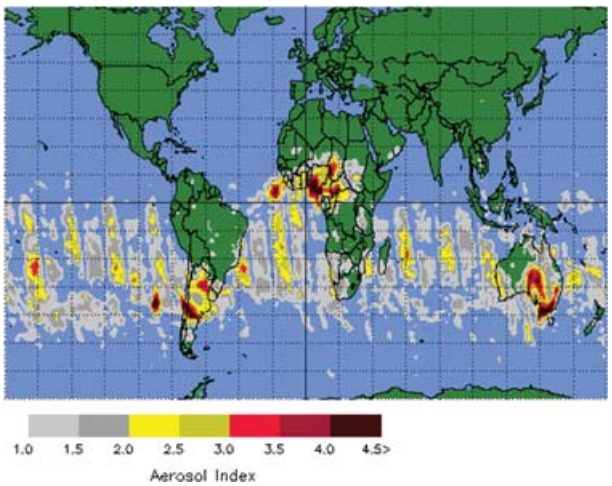

Fig. 8. TOMS aerosol index (http://toms.gsfc.nasa.gov/aerosols/aerosols˙v8.html) for the 24th of January 2003.

1997 and previous reports (Inoue and Ishii, 2005), and is associated with increased phytoplankton biomass.

A comparison of various forcings between 1997 and 2003 suggests that the intense CO_2 sink in February 2003 was not directly related to changes in total stratospheric ozone warming, or anomalous surface mixing. The occurrence of an ENSO event in 2002/2003 and episodic atmospheric iron inputs that may be linked to drought conditions in Australia could be significant in causing the interannual change.

Almost all climate models suggest that Australia and the SAR will experience negative precipitation anomalies in the future (IPCC, 2001). Dust inputs and fertilization of surface waters may occur more often in the future with the ocean carbon sink in the SAR and POOZ regions enhanced. The affect would produce a negative climate feedback that could partially cancel the positive feedback associated with sea surface warming and increased stratification. The transient responses described here provide fundamental data to test ocean biogeochemical models. Future observational programs to characterize and investigate the processes driving these responses will be a valuable tool in understanding how the ocean biogeochemical system will respond to future climate forcing.

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