

## Interannual variability of the carbon dioxide system in the southern Indian Ocean (20°S–60°S): The impact of a warm anomaly in austral summer 1998

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[1] The interannual variations of the carbon dioxide system and air-sea CO<sub>2</sub> fluxes are analyzed in the southwestern Indian Ocean from both in situ (Océan Indien Service d'Observations cruises in 1998 and 2000) and simulated oceanic CO<sub>2</sub> fugacity  $f_{\text{CO}_2}$ , dissolved inorganic carbon, total alkalinity, and nutrients. During austral summer of 1998, the ocean was warmer from 20°S to 60°S. In the Subtropical Zone (20°S–37°S), the warming, associated with the subtropical dipole pattern, creates an oceanic CO<sub>2</sub> source around 2 mmol m<sup>-2</sup> d<sup>-1</sup> in January 1998 where all previous observations, included in 2000, indicated that this region was a small sink in summer. In the Sub-Antarctic and Polar Front Zones (37°S–50°S), the  $f_{\text{CO}_2}$  interannual signal is not clearly detected because of the complex coupling between the horizontal and vertical dynamics and the biological activity that creates very high mesoscale  $f_{\text{CO}_2}$  variability in each summer. For the austral region, south of the Polar Front, we observe large variability in the CO<sub>2</sub> sources and sinks between summer 1998 and 2000. In the Seasonal Ice Zone (south of 58°S) the interannual variation of the CO<sub>2</sub> flux was mainly controlled by a warmer ocean, from 1.1°C, in summer 1998 (CO<sub>2</sub> source of 3.2 mmol m<sup>-2</sup> d<sup>-1</sup>) and an enhanced primary production occurring during summer 2000 (CO<sub>2</sub> sink of -3.8 mmol m<sup>-2</sup> d<sup>-1</sup>). In the Permanent Open Ocean Zone (POOZ, latitude 50°S–57°S), despite the half-degree warmer sea surface in 1998, we observe lower  $f_{\text{CO}_2}$  (-15 μatm (1 atm = 10<sup>5</sup> × 1.01325 N m<sup>-2</sup>) on average) during this period, and consequently, we estimate a double of the oceanic CO<sub>2</sub> sink in this region in 1998 than 2000, from -3.2 mmol m<sup>-2</sup> d<sup>-1</sup> in 1998 to -1.3 mmol m<sup>-2</sup> d<sup>-1</sup> in 2000. The strong oceanic CO<sub>2</sub> sink in 1998 is associated with a dramatic decrease of nutrients (silicates less than 2 μmol kg<sup>-1</sup>). The enhanced biological activity in summer 1998, probably diatoms bloom, is also apparent in Sea-viewing Wide Field-of-view Sensor (SeaWiFS) chlorophyll *a* estimates. Using a one-dimensional biogeochemical model applied in the POOZ and SeaWiFS time series, we predict a stronger oceanic sink during summer season than the winter, which is contrary to previous studies. The model suggests that biological activity controls most of the sink anomaly in summer 1998 (for about 70%) and dominates the warming effect. INDEX

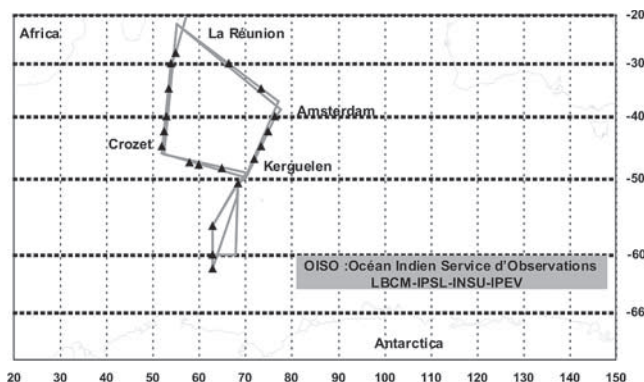
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### 1. Introduction

[2] To better predict the future atmospheric CO<sub>2</sub> concentrations and related climate change at global and regional scales, it is necessary to understand how the CO<sub>2</sub> is exchanged between the different climate components, the

atmosphere, the ocean, and the terrestrial biosphere. We need to determine where the sources and sinks of carbon are and how and why these sources and sinks are changing at seasonal, interannual, and decadal scales. The oceans contain about 50 times more CO<sub>2</sub> than does the atmosphere and absorb a large portion, approximately one third, of the carbon annually released by human activities. How the oceanic carbon uptake will change in the future is a fundamental



**Figure 1.** Track of OISO cruises in the southwestern Indian Ocean. Triangles indicate the position of hydrological stations made during cruises (from the surface to 1000 m depth or from the surface to the bottom).

issue to reduce uncertainties associated to climate change projections. However, the future oceanic uptake is difficult to predict because the complex coupling between the biogeochemical cycle and the ocean dynamics and thermodynamics are poorly observed and not well understood. Interannual observations of biogeochemistry in the ocean can help to elucidate the mechanisms of this complex coupling and to improve future prediction of the oceanic uptake of CO<sub>2</sub>.

[3] At a global scale, the estimation of the interannual variability of air-sea CO<sub>2</sub> fluxes has been derived from inverse atmospheric models [Keeling *et al.*, 1995; Ciais *et al.*, 1995; Francey *et al.*, 1995; Rayner *et al.*, 1999; Bousquet *et al.*, 2000]. These atmospheric estimates generally suggest that interannual variability in the global air-sea CO<sub>2</sub> fluxes can be as large as  $\pm 1$  to  $\pm 2$  Gt C yr<sup>-1</sup>. From an oceanic point of view, estimates of interannual variability of air-sea CO<sub>2</sub> fluxes at local and regional scales are dominated by repeated observations in the equatorial Pacific [Feely *et al.*, 1997, 1999, 2002; Inoue and Sugimura, 1992; Wong *et al.*, 1993], the subarctic Pacific [Murphy *et al.*, 1998], and at subtropical Joint Global Ocean Flux Study (JGOFS) time series in the Atlantic Ocean, Bermuda Atlantic Time-series Study (BATS) [Bates *et al.*, 1998] and in the Pacific Ocean, HOT [Winn *et al.*, 1994]. Additional estimates are made by combining observations with semiprognostic models [Metzl *et al.*, 1998; Louanchi *et al.*, 1999]. Diagnostic methods, based on properties/properties relationship (e.g.,  $p\text{CO}_2$ /sea surface temperature (SST)) are also used to estimate interannual air-sea CO<sub>2</sub> fluxes in the Indian Ocean [Metzl *et al.*, 1995], the equatorial Pacific [Boutin *et al.*, 1999], or at a global scale [Lee *et al.*, 1998]. These studies, either regional or global, all concluded that there is relatively small interannual variability in the air-sea CO<sub>2</sub> fluxes (from  $\pm 0.1$  to  $\pm 0.5$  Gt C yr<sup>-1</sup>) compared to the atmospheric inverse approaches. A recent global-scale ocean modeling study also reached a similar conclusion when it estimated global air-sea interannual variability of  $\pm 0.4$  Gt C yr<sup>-1</sup> [Le Quéré *et al.*, 2000]. In this study, variability is dominated by the changing distribution of  $p\text{CO}_2$  in the equatorial Pacific. For the tropical Pacific region, the spot of marked El Niño Southern Oscillation (ENSO) events, all these studies show a certain coherence on the CO<sub>2</sub> fluxes variability. For all the other oceanic regions, the interannual

signal has to be more clearly defined. For the Southern Hemisphere, particularly the austral regions, in situ observations of the oceanic carbon system at interannual scale are very poorly documented. Although in the Southern Ocean a large-scale climatic signal occurs [White and Peterson, 1996] and could create, as ENSO does in the Pacific, a nonnegligible interannual variability of the air-sea CO<sub>2</sub> fluxes through variability of winds, ocean dynamics, solubility, biological productivity, and biogeochemical components.

[4] In this context, this paper presents a new analysis of variability in the air-sea CO<sub>2</sub> fluxes in the southern Indian Ocean and discusses how the oceanic carbon system responds when the ocean is warmed at large scale, as was observed in this region in 1998. After a description of the cruises and measurements (section 2), we compare the differences in sea surface  $f_{\text{CO}_2}$  observed between 1998 and 2000 in the subtropical, the frontal, and the austral regions (section 3). In the austral regions, we observed a large interannual variability in the CO<sub>2</sub> sources and sinks. The analysis of this dramatic air-sea CO<sub>2</sub> flux variations in the Southern Ocean is presented in section 4, based on the biogeochemical properties observed during the cruises. To quantify the processes that are controlling the observed interannual variability of the oceanic CO<sub>2</sub> sink in the Permanent Open Ocean Zone (POOZ, 50°S–57°S), we apply a one-dimensional (1-D) biogeochemical model (section 5). Concluding remarks are presented in section 6.

## 2. Data Collections During Océan Indien Service d'Observations Cruises

[5] One purpose of the observational program Océan Indien Service d'Observations (OISO) is to document the interannual variability of the CO<sub>2</sub> cycle and associated air-sea CO<sub>2</sub> fluxes in the southwestern region of the Indian Ocean, from the Subtropical Zone (SBT) to the Southern Ocean (Figure 1). The cruises are conducted on board RV *Marion-Dufresne* during resupply voyages to the French sub-Antarctic Islands. This long-term oceanic CO<sub>2</sub> monitoring program complements other international cruises in other ocean basins.

[6] During the OISO cruises, continuous sea surface measurements are recorded for temperature (SST), salinity (SSS), fluorescence, CO<sub>2</sub> fugacity ( $f_{\text{CO}_2}$ ), total dissolved inorganic carbon (DIC), and total alkalinity (TA). Regular discrete samples of sea surface water, six per day, are also performed to measure salinity, nutrients (nitrates + nitrites, silicates), and chlorophyll *a* (chl *a*) concentrations. In addition to surface data, the water column data are collected at several stations in the different regions and at the same position during each OISO cruise (Figure 1). To study the temporal stability (or variability) of the oceanic properties below the mixed layer, the first 1000 m of the water column is sampled at each station. Three additional full-depth stations are obtained; one in the subtropical region (4500 m for station 17 at 30°S–66°E), one situated south of the Polar Front (PF, 1650 m for station 10 at 50°40'S, also known as Kerfix/JGOFS time series station [Jeandel *et al.*, 1998]), and one in the Southern Ocean (4750 m for station 11 at 56°30'S–63°E). The deep samples are also used to check the accuracy of the measurements from cruise to cruise.

Hydrocast stations are performed with a Conductivity-Temperature-Depth/Rosette (CTD Seabird 911 plus); temperature and conductivity probes are calibrated before each cruise. At each station, bottles are fired at standard depths to measure biogeochemical properties (nutrients, chl *a*, DIC, and TA). Continuous surface and water column salinity records are controlled by regular conductivity measurements (Guildline Autosol 8400B). Sea surface temperature is also checked against CTD's surface temperature.

[7] Nutrients (silicates and nitrates + nitrites) are measured with an Autoanalyzer II Technicon, following the methods described by Tréguer and Le Corre [1975]. Series of diluted standards (0–40  $\mu\text{M L}^{-1}$  for nitrates and 0 to 150  $\mu\text{M L}^{-1}$  for silicates) are prepared from reference solutions (OSI, Ocean Scientific International, United Kingdom). The results of replicate measurements for about 50 deep samples give a measure of the reproducibility (mean difference) of 0.25  $\mu\text{mol kg}^{-1}$  for silicates and 0.1  $\mu\text{mol kg}^{-1}$  for nitrates, corresponding to about 0.3% for both nutrients.

[8] TA and total DIC are measured on board using a technique based on the potentiometric method of Edmond [1970] with a closed cell described by Goyet *et al.* [1991]. For continuous measurements of sea surface TA and DIC, the system has been automatized: a program developed on LabVIEW controls the thermostation of the seawater samples, the transfer of the water into the cell, the acidification, and the titration. The calculation of the equivalent point is estimated using a nonlinear regression method [Department of Energy, 1994]. The same instrument is used to measure the pairs of DIC and TA sampled on hydrocast stations (these samples are immediately poisoned with saturated mercuric chloride solution). In that case, the automatized sampling of sea surface water is stopped and discrete samples are measured during resupply to Crozet, Kerguelen, and Amsterdam Islands (this is generally done within 1–2 days after sampling). During each cruise, we use the Certified Referenced Material (CRM) provided by A. Dickson (Scripps Institution of Oceanography (SIO), University of California). Based on CRM analyses (between 30 and 50 depending the cruise), the precision of TA and DIC data is estimated at 1.9 and 2.0  $\mu\text{mol kg}^{-1}$ , respectively, which is the precision achieved during most of WOCE/JGOFS cruises [e.g., Millero *et al.*, 1998; Sabine *et al.*, 1999]. The reproducibility estimated from about 80 replicate analyses of deep samples is 1.4 and 1.3  $\mu\text{mol kg}^{-1}$  for TA and DIC, respectively. Comparison of TA and DIC measured in deep waters (Theta < 0.4°C) for the four cruises does not reveal significant offsets (i.e., larger than the precision of 2.0  $\mu\text{mol kg}^{-1}$ ). Therefore one has to consider that observed temporal difference of 3  $\mu\text{mol kg}^{-1}$  period<sup>-1</sup> (in DIC and TA) is the lower acceptable limit to interpret real changes between cruises.

[9] The  $f_{\text{CO}_2}$  measurement technique is the same as the one described by Poisson *et al.* [1993] and Metzl *et al.* [1995, 1999]. This is also the same instrument used by our group during the recent international at-sea intercomparison of  $f_{\text{CO}_2}$  systems [Koertzing *et al.*, 2000]. Sea surface water is continuously equilibrated using a “thin film” type equilibrator thermostated with surface seawater. The CO<sub>2</sub> in the dried gas is measured with a nondispersive infrared analyzer (NDIR, Siemens Ultramat 5F). Standard gases for calibra-

tion (270, 350, 480 ppm) and atmospheric CO<sub>2</sub> are measured every 7 hours. The temperature in the equilibrium cells is 0.2°–0.7°C warmer than the SST unless in frontal region where temperature difference can reach 2°C episodically. To correct  $f_{\text{CO}_2}$  measurements to in situ data, we used polynomials given by Weiss and Price [1980] for vapor pressure and by Copin-Montégut [1988, 1989] for temperature. The same instrument and data processing are used during all OISO cruises. The oceanic  $f_{\text{CO}_2}$  data are accurate to about  $\pm 0.7$   $\mu\text{atm}$  (1 atm =  $10^5 \times 1.01325$  N m<sup>-2</sup>). Average atmospheric CO<sub>2</sub> measured on board was 362.7 ppm ( $\pm 0.5$ ) in January 1998 and 365.4 ppm ( $\pm 0.4$ ) in August–September 1998. The summer to winter increase of +2.7 ppm represents a strong anomaly in this area compared to the amplitude of the seasonal cycle and the annual growth rate generally observed at CO<sub>2</sub> monitoring stations in the southern subtropical, subpolar, and high latitudes [Conway *et al.*, 1994]. This signal observed during OISO cruises in 1998 is confirmed with observations at Amsterdam Island, located at 38°S in the South Indian Ocean: here, the monthly CO<sub>2</sub> concentrations were 362.6 ppm ( $\pm 0.4$ ) for January–February 1998 and 364.7 ppm ( $\pm 0.2$ ) in August–September 1998 (M. Ramonet, LSCE/IPSL, personal communication, 1999). At South Pole, CO<sub>2</sub> concentrations were 362.2 and 365.0 ppm for the same periods [Keeling and Whorf, 1999]. In January 2000, the average atmospheric CO<sub>2</sub> recorded during OISO cruise, from 20°S to 60°S, was 366.2 ppm ( $\pm 0.5$ ), which is also close to CO<sub>2</sub> measurements at Amsterdam station, 366.4 ppm ( $\pm 0.1$ ) in January–February 2000 (M. Ramonet, LSCE/IPSL, personal communication). The good quality of the atmospheric data measured on board demonstrates that the  $f_{\text{CO}_2}$  instrument was operating well during the cruises and that variation of about 3  $\mu\text{atm}$  in oceanic  $f_{\text{CO}_2}$  is a realistic lower limit to interpret the interannual signals. We will see in the next section that oceanic  $f_{\text{CO}_2}$  interannual variability is on average much higher, around 10–30  $\mu\text{atm}$  depending on the region.

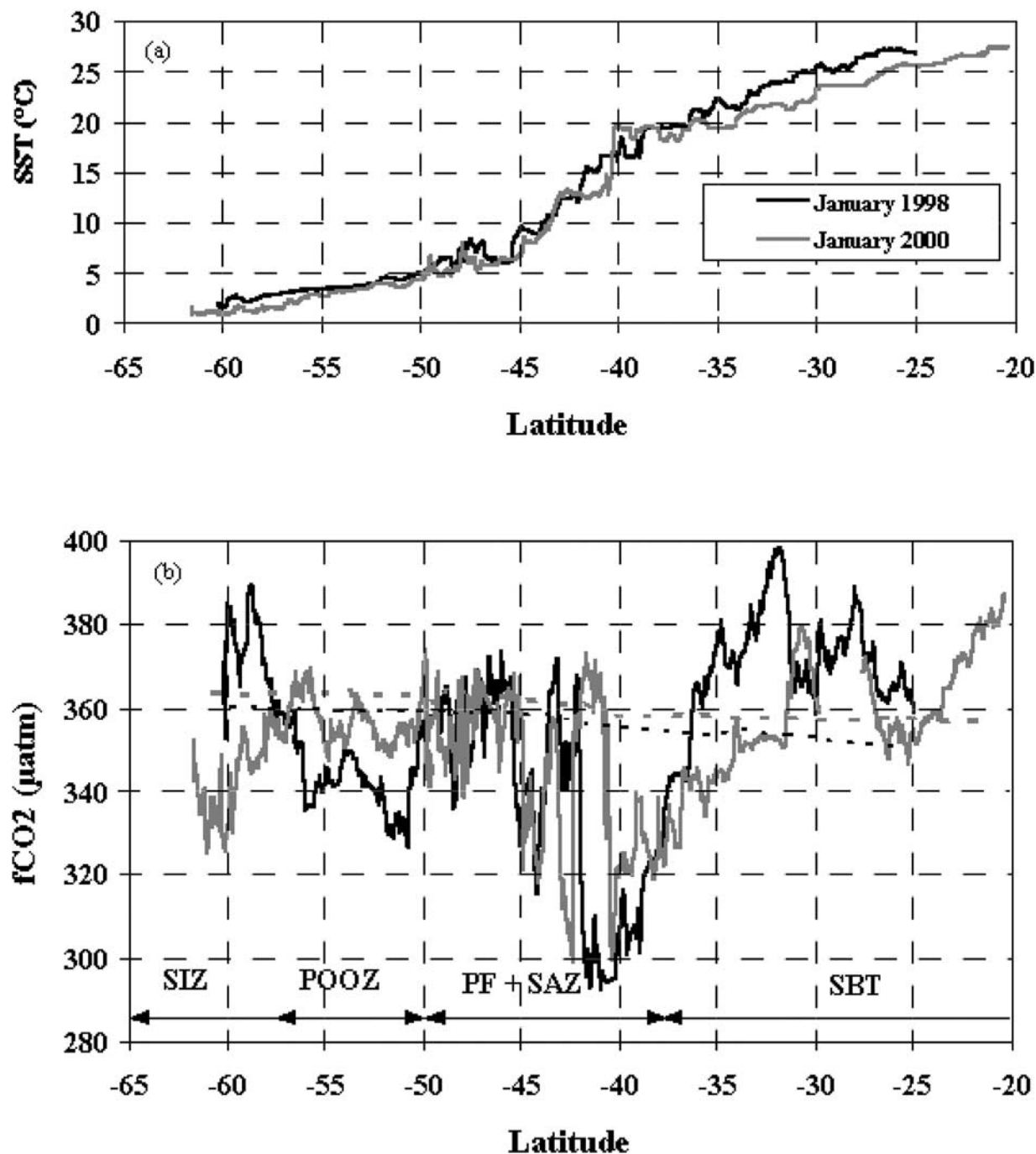
### 3. Large-Scale Distribution of CO<sub>2</sub> Sources and Sinks in Austral Summer 1998 and 2000

[10] Five main biogeochemical provinces are sampled during OISO cruises: the SBT (approximately 20°S–37°S), the Sub-Antarctic Zone (SAZ,  $\approx 37^\circ\text{S}$ – $45^\circ\text{S}$ ), the Polar Front Zone (PFZ,  $\approx 45^\circ\text{S}$ – $50^\circ\text{S}$ ), the POOZ  $\approx 50^\circ\text{S}$ – $57^\circ\text{S}$  and the Seasonal Ice Zone (SIZ,  $\approx$  south of  $58^\circ\text{S}$ ). These regions are characterized by distinct spatial and temporal biogeochemical variations. In this section, we describe for each region the large-scale  $f_{\text{CO}_2}$  differences observed between austral summer 1998 and 2000. We focus the description of the  $f_{\text{CO}_2}$  and temperature distributions along the western track (Figures 2a and 2b) and the time/latitudinal distribution of Sea-viewing Wide Field-of-view Sensor (SeaWiFS)-derived monthly level 3 chl *a* data (NASA/GSFC V0 Daac Version 6.0) extracted along the ship track (Figure 3).

#### 3.1. Interannual Variations of Air-Sea CO<sub>2</sub> Fluxes in the Subtropical Region

[11] The southern boundary of the SBT is located around 37°S where we observed a sharp north-south  $f_{\text{CO}_2}$  gradient

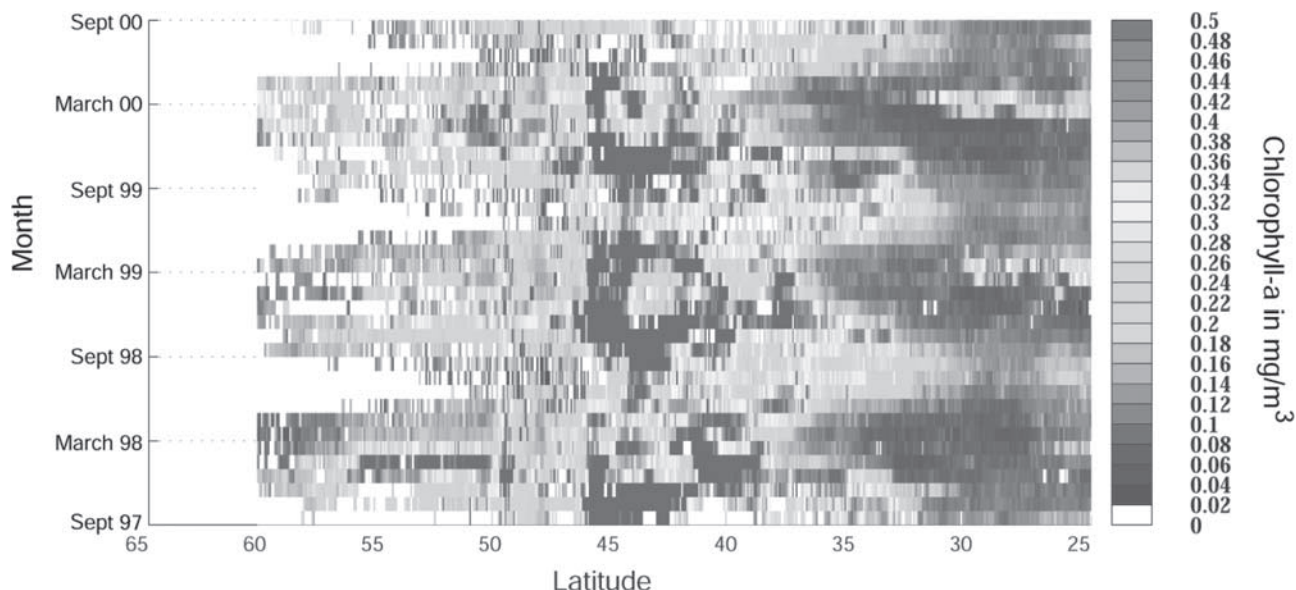




**Figure 2.** Evolution along the western track of OISO cruises (from La Réunion to 60°S) of (a) the sea surface temperature in January 1998 (solid black line) and 2000 (solid shaded line) measured on board during cruises and (b) the oceanic  $f\text{CO}_2$  during January 1998 (solid black line) and January 2000 (solid shaded line) with atmospheric  $\text{CO}_2$  in 1998 (dashed black line) and 2000 (dashed shaded line) measured on board during cruises. Different regions SBT, Sub-Antarctic, and Polar Front Zones (PF + SAZ), POOZ, and SIZ are indicated.

each summer (Figure 2b) associated with a marked increase in SeaWiFS chl *a* concentrations (Figure 3). In the western SBT region, the surface ocean was warmer in 1998 (Figure 2a); this corresponds to higher  $f\text{CO}_2$  measured during

this year. The SST anomaly observed in the western Indian Ocean during austral summer 1998 is related to the sub-tropical dipole pattern (SDP) described by *Behera and Yamagata* [2001]: in the southwestern region of the Indian



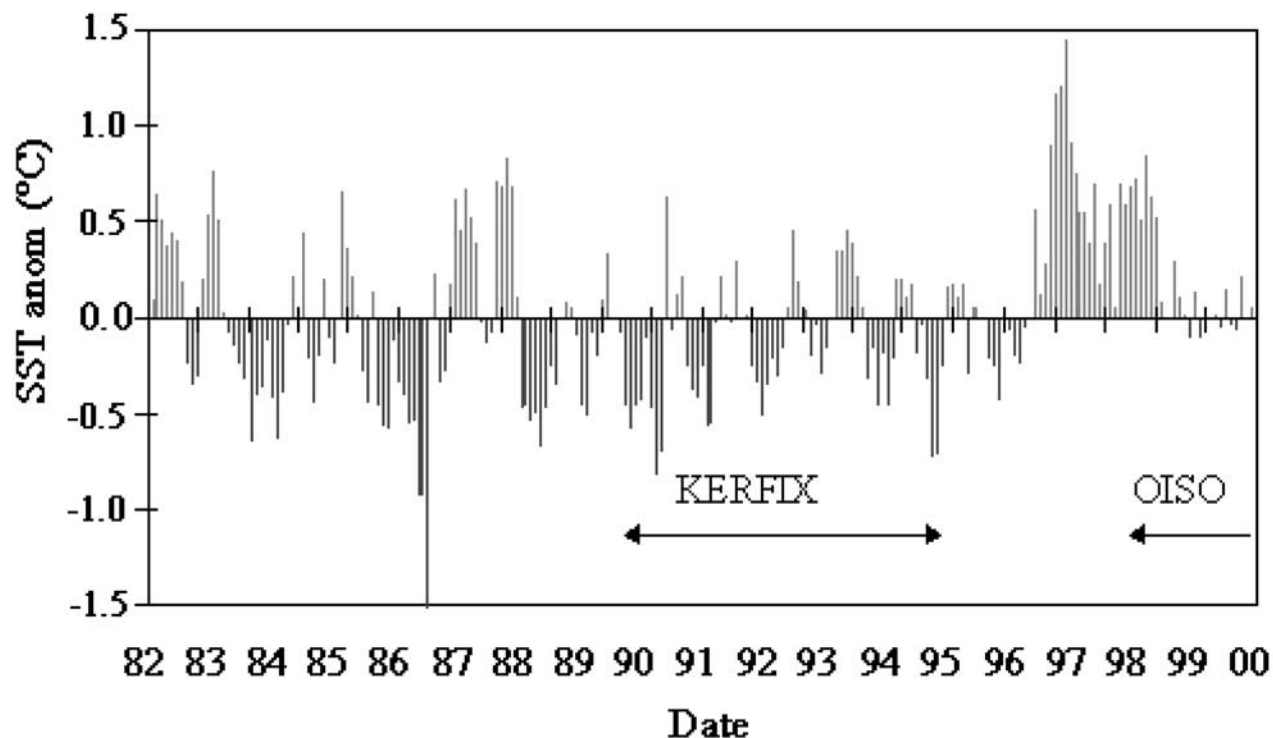
**Figure 3.** Evolution along the western track of OISO cruises (from La Réunion to 60°S) of the monthly chl *a* concentration in  $\text{mg m}^{-3}$  (SeaWiFS level 3 binned data) from September 1997 to September 2000. Blank indicates no data. See color version of this figure at back of this issue.

Ocean, where mixed layer is relatively shallow in austral summer, reduced evaporation and upwelling cause a positive SST anomaly. This mechanism is controlled by variability in the winds, and it was operating in summer 1998. During both the summer 1998 and 2000 cruises, the deep chl *a* maximum concentrations between 0.3 and 0.5  $\text{mg m}^{-3}$  are observed around 100–120 m well below the mixed layer in the SBT. The primary production is relatively low in surface waters of the SBT especially during austral summer, where nutrients are very low. Therefore the sea surface interannual  $f_{\text{CO}_2}$  variations are mostly controlled by temperature changes. This thermodynamical control on  $f_{\text{CO}_2}$  is a well-known process that explains most of the seasonal sea surface  $f_{\text{CO}_2}$  variations in the subtropical oceans. This was detailed in the South Indian SBT [Poisson *et al.*, 1993; Metzl *et al.*, 1995, 1998], in the SBT of the North Atlantic [Bates *et al.*, 1996, 1998] or in the SBT in the North and South Pacific [Inoue *et al.*, 1995; Weiss *et al.*, 1982; Winn *et al.*, 1994]. Considering that interannual variability of primary production in the mixed layer is negligible in the subtropical gyres, we conclude that the thermal control process is also valid to explain most of the interannual  $f_{\text{CO}_2}$  changes in the SBT governed by the SDP in the South Indian gyre. It is worth noting that during austral summer the SBT zone of the Indian Ocean is generally found as a moderate CO<sub>2</sub> sink or near equilibrium [Poisson *et al.*, 1993; Metzl *et al.*, 1995, 1998; Louanchi *et al.*, 1996; Sabine *et al.*, 2000; Takahashi *et al.*, 2002] such as the most recent January 2000 observations in this region. Using our observations and the formulation of the gas transfer coefficient proposed by Wanninkhof [1992], we estimate an oceanic CO<sub>2</sub> source around 2  $\text{mmol m}^{-2} \text{d}^{-1}$  in January 1998 and a sink of  $-1 \text{ mmol m}^{-2} \text{d}^{-1}$  in January 2000 for the SBT region. The ocean CO<sub>2</sub> source estimated during January 1998 in the western subtropical Indian gyre was

much larger than any reported observations in the same region.

### 3.2. Interannual $f_{\text{CO}_2}$ Variations are not Captured in the Frontal Zone

[12] At 40°S we enter the Sub-Antarctic and Polar Front Zones designed by a complex frontal structure: the Southern Subtropical Front and the Agulhas Front to the north and the Sub-Antarctic Front and the Polar Front to the south [Belkin and Gordon, 1996]. In the southwestern Indian Ocean, north of the Crozet Archipelago, the Agulhas Front, the Subtropical Front, and the Sub-Antarctic Front sometimes collapse to form a triple front, called the “Crozet Front.” This is a region where primary production is enhanced (Figure 3) [Moore and Abbott, 2000] and where minima of  $f_{\text{CO}_2}$  are observed each summer (Figure 2b). All around the circumpolar zone, the SAZ is an oceanic CO<sub>2</sub> sink with the largest sinks observed in austral summer when the sea surface  $f_{\text{CO}_2}$  is under the control of primary production [Metzl *et al.*, 1999]. In the summer of 1998 the  $f_{\text{CO}_2}$  minimum was centered at 40°S (Figure 2b). Whereas in the summer of 2000, we observed two minima, at 40°S and 42°S, where SST gradients are sharp. Around 45°S, an  $f_{\text{CO}_2}$  minimum was observed for both periods. In the PFZ (around 47°S–50°S), the spatial distribution of  $f_{\text{CO}_2}$  is less organized, with high, low, and near-equilibrium values. On average the SST was warmer by about 1°C in 1998 in the frontal zone, but we did not detect clear difference of  $f_{\text{CO}_2}$  as was observed in the SBT region. The position of the circumpolar frontal system is mostly controlled by topography [Moore *et al.*, 2000]; however, north and south displacement of the fronts, meanders, eddies, and filament occur. It is also likely that the variability of the circumpolar circulation has a strong impact on the primary productivity and thus on summer  $f_{\text{CO}_2}$  distribution; this adds complexity



**Figure 4.** Monthly SST anomalies from 1982 to 2000 in the POOZ of the Indian Ocean (location 50.5°S–68.5°E). The period of observational programs KERFIX and OISO is noticed.

when trying to analyze interannual variations. In addition, in the frontal zone, where mixed layers are much deeper than in the subtropics, an anomaly in the climatic forcing (such as the SDP in the subtropics) may not have the same dramatic impact on surface properties as observed in the southern gyre. Because of a complex (and not yet understood) interplay between circulation, mixing, and primary productivity in the frontal zone, we must conclude that the interannual variability in this sector is very difficult to capture without a more comprehensive and long-term data set.

### 3.3. Interannual Variation of CO<sub>2</sub> Sources and Sinks in the Southern Ocean

[13] South of the PF, variations of the CO<sub>2</sub> sources and sinks are much clearly visible compared to the frontal zone. At 50°S, we enter the so-called Permanent Open Ocean Zone. The SIZ is located south of 58°S in the South-Western Indian sector. Continuous  $f_{\text{CO}_2}$  measurements (Figure 2) show that the CO<sub>2</sub> sink was potentially much stronger in the POOZ during summer 1998 compared to 2000, and interestingly, opposite to the variations in the SIZ.

[14] In the Southern Ocean, south of 50°S the sea surface was warmer in summer 1998 (Figure 2a). The warm anomaly observed along the ship track during OISO 1998 cruise was also apparent on a regional scale in the Indian sector of the Southern Ocean as deduced from SST monthly fields [Reynolds and Smith, 1994]. The origin of this warm anomaly may be related to the Antarctic Circumpolar Wave (ACW) [White and Peterson, 1996]. During the period 1980–1991, White and Peterson report a maximum range

of SST anomalies of 1.6°C in the POOZ (for latitude 56°S). To estimate the magnitude of anomalies for the period of our cruises, we computed the monthly SST anomalies from 1982 to 2000 at location 50.5°S–68.5°E (Figure 4), which is the position of one of our hydrocast stations in the POOZ (OISO st-10). The station is known as KERFIX/JGOFS time series station, which was operating in 1990–1995 [Jeandel et al., 1998]. In this region, SST varies from about 2°C in winter to 4.5°C in summer. Interannual SST anomalies were as large as 1.5°C (Figure 4). The OISO–1 cruise in January–February 1998 was situated in the middle of a long warm event occurring in 1997–1998. The maximum anomaly was reached in March 1997. During January–February 2000, the period of OISO-4 cruise, SST was close to climatological values. Figure 4 also shows that the KERFIX/JGOFS time series station was sampled during a relatively cold period (1990–1995). This may explain some of the differences we detected between KERFIX and OISO biogeochemical properties and air-sea CO<sub>2</sub> fluxes (this will be discussed in section 5).

[15] Average anomalies calculated for the whole region between 60°E–68°E and south of 50°S, show the same interannual variability as presented at the Kerfix site in Figure 4. Although the warm anomaly observed in the South Indian Ocean in 1998 is quite large, it is not unique. Warm anomalies also occurred during 1982–1983 and 1987–1988 (Figure 4) and may be associated with planetary event, such as ENSO. At the global scale, during ENSO conditions, positive SST anomalies were observed and simulated in the South Indian and southeast Pacific sectors



of the Southern Ocean [Li, 2000]. Moreover, it was suggested that these anomalies were likely controlled by the increase of surface solar radiation and changes in clouds cover [Li, 2000]. This is an interesting coupling of external forcings, as for  $f_{\text{CO}_2}$ , the impact of warming and clouds will be different; a warming will increase  $f_{\text{CO}_2}$ , whereas more light (less clouds) may enhance the primary production and decrease  $f_{\text{CO}_2}$ . We now attempt to identify what processes can explain the interannual  $f_{\text{CO}_2}$  variations observed in the Southern Indian Ocean.

#### 4. Observed Interannual Variations of Biogeochemistry in the Southern Ocean

[16] To quantify the differences of the air-sea CO<sub>2</sub> fluxes observed between summer 1998 and 2000, we use the observed variations of hydrological and biogeochemical properties in the following the two main regions of the Southern Ocean, the POOZ and the SIZ (Figures 5 and 6). These two regions are separated by a well-defined biogeochemical front at 56°S–57°S. Therefore when averaging the data, the POOZ will be represented by the data obtained in the band 50°S–56°S. The tracks of the two cruises were exactly the same from 50°S to 60°S. Measurements were made south of 60°S only during January 2000. These data are presented in the figures but were not included in the averaging procedure. Only measurements recorded from 58°S to 60°S were representative of average changing properties for the SIZ. The regional averages of sea surface properties are presented in Tables 1 and 2.

##### 4.1. The Seasonal Ice Zone (South of 58°S)

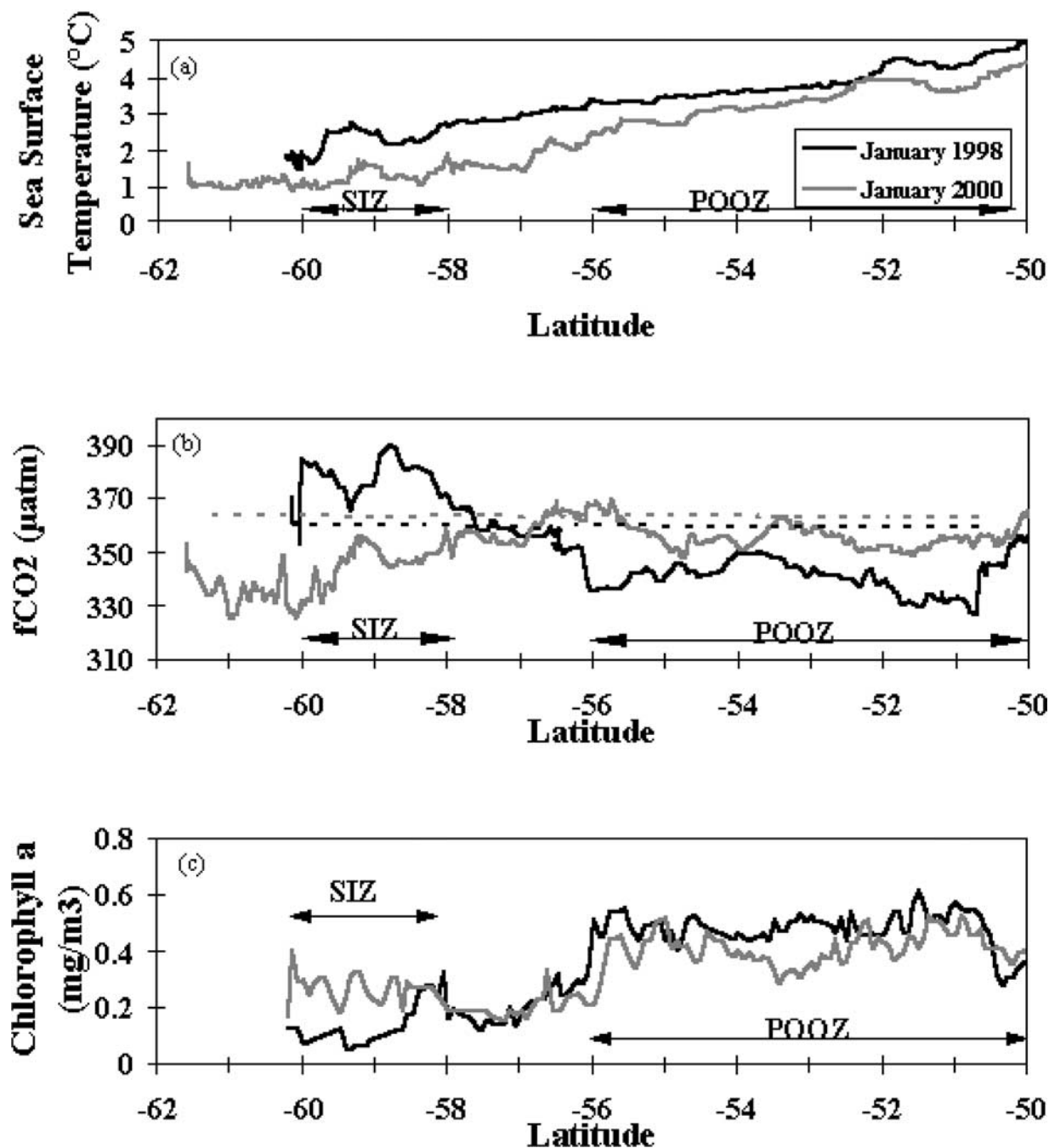
[17] The SIZ was a CO<sub>2</sub> source to the atmosphere in 1998 and a sink in summer 2000 (section 3.3). The average difference of  $f_{\text{CO}_2}$  between the two periods is 38  $\mu\text{atm}$  deduced from direct measurements (Table 1) and 33  $\mu\text{atm}$  when using calculated  $f_{\text{CO}_2}$  values based on average DIC and TA concentrations (using “CO<sub>2</sub>sys program” developed by Lewis and Wallace [1998] and the dissociation constants from the work of Goyet and Poisson [1989]). To separate the thermodynamics on  $f_{\text{CO}_2}$  from other processes, we calculated  $f_{\text{CO}_2}$  with DIC concentrations of 1998 but with SST observed in 2000. The warm anomaly in 1998 explained about 22  $\mu\text{atm}$  of the  $f_{\text{CO}_2}$  differences in the SIZ. The residual around 10–15  $\mu\text{atm}$  was explained by DIC variations, controlled by changes in primary production, air-sea fluxes and/or horizontal advection, and vertical mixing. The effect of air-sea flux was evaluated using the average mixed-layer depth (MLD, 50 m) and summer wind speed [Wanninkhof, 1992]. The calculation showed that the ocean lost about 2  $\mu\text{mol kg}^{-1} \text{ month}^{-1}$  of DIC in summer 1998 (a source) and gained 2  $\mu\text{mol kg}^{-1} \text{ month}^{-1}$  in summer 2000 (a sink). This effect increased the relative difference of DIC between the 2 years. Although the distributions of DIC presented quite a large spatial variability with a marked maximum at 59°S (Figure 6a), the detectable variation of DIC was likely mostly related to biological activity in the SIZ during summer as previously observed in the South-East Indian sector of the SIZ [Ishii et al., 1998] and at coastal site in Prydz Bay [Gibson and Trull, 1999].

[18] The chl *a* values observed in the SIZ during the warm period of January 1998 were lower than those measured in 2000 (Figures 3 and 5c). They were also low compared to previous observations in the same region [Fiala et al., 1998a]. Surprisingly, silicates’ data (but few points) showed that concentrations were almost the same for the 2 years in the SIZ, the difference being only 2  $\mu\text{mol kg}^{-1}$ , and opposite to nitrates and DIC variations (Figure 6). Whatever the year, the chl *a* values and silicate concentrations remain high in the Indian SIZ. To understand the origin of the interannual  $f_{\text{CO}_2}$ , DIC, and chl *a* variability in the SIZ, one must find an explanation other than a purely silicate-limited scenario.

[19] One possibility was the control of temperature on the polar microalgae. Indeed, in the Indian Ocean SIZ, a parabolic chl *a*/temperature relationship with a well-defined maximum at 0°C was proposed by Fiala et al. [1998a]. This may correspond to a direct control of temperature on phytoplanktonic species living in cold waters. Following this curve, a warming of 1°C would create a chl *a* reduction of 0.15  $\text{mg m}^{-3}$ . This was almost the difference we observed for SST and chl *a* between summer 1998 and 2000 (Table 1). Another assumption was that the high solar radiation that might be higher during warm summer 1998 (section 3.3) may be too high for an optimal phytoplanktonic growth in surface. This could enhance the primary production in deeper layers, where both light and temperature would offer better ecophysiological adaptations for antarctic species. In the Southern Ocean, regional and temporal variations of phytoplankton species also depend on specific macronutrient and micronutrient demands (e.g., iron) and grazing pressure [e.g., Lancelot et al., 2000]. For example, it was possible that during summer 2000 the high chl *a*, low DIC, and nitrate but high silicate concentrations were related to significant production of nonsiliceous *Phaeocystis* colonies. Unfortunately, to quantify and demonstrate these assumptions, we had insufficient subsurface data and no species distributions in the SIZ. Additional observations and a coupled physical-biogeochemical modeling are needed to test these hypothesis and to quantify the control of the thermodynamics and primary production on the  $f_{\text{CO}_2}$  and air-sea CO<sub>2</sub> fluxes’ interannual variations in the SIZ.

##### 4.2. The Permanent Open Ocean Zone (50°S–57°S)

[20] As in the SIZ, air-sea CO<sub>2</sub> fluxes in the POOZ are dramatically different in 1998 and 2000. They are controlled by the variations of oceanic  $f_{\text{CO}_2}$  distribution. The fluxes deduced from average  $f_{\text{CO}_2}$  and SST observations in the POOZ were  $-3.2 \text{ mmol m}^{-2} \text{ d}^{-1}$  in midsummer 1998 and  $-1.3 \text{ mmol m}^{-2} \text{ d}^{-1}$  in midsummer 2000. In the POOZ, the summer SST anomaly was smaller than in the SIZ. From in situ data we calculate an average difference of 0.5°C in the POOZ between the summers 1998 and 2000 (Table 1). The warm water in 1998 causes a relative increase of  $f_{\text{CO}_2}$  of about 10  $\mu\text{atm}$  that has to be subtracted from the observed  $f_{\text{CO}_2}$  difference of  $-14.6 \mu\text{atm}$  in the POOZ. We must now explain a temperature-normalized  $f_{\text{CO}_2}$  difference of  $-24.6 \mu\text{atm}$  between the two periods. The higher chl *a* concentrations observed in the POOZ in 1998 ( $+0.1 \text{ mg m}^{-3}$  on average) are consistent with a greater biological uptake



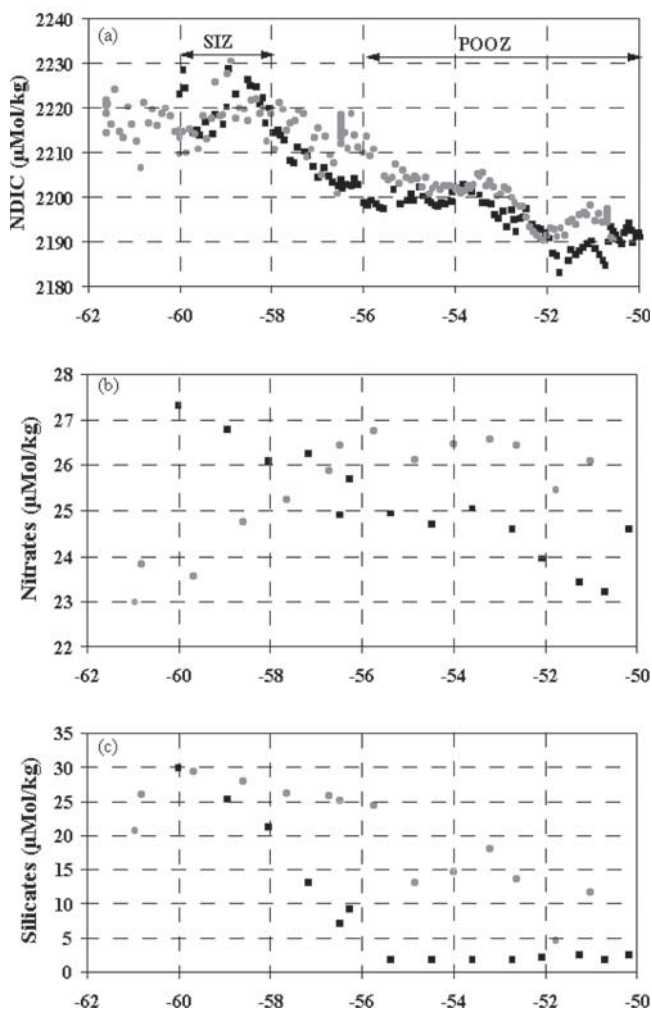
**Figure 5.** A focus on the Southern Ocean (south of 50°S) in January 1998 (black line) and 2000 (shaded line) along the western track of OISO cruises. Arrows indicate the area taken into account to average data. (a) Evolution of the continuously measured SST, (b) evolution of the continuous oceanic  $f_{\text{CO}_2}$  in solid line and atmospheric  $f_{\text{CO}_2}$  in dashed line, and (c) evolution of monthly chl  $a$  concentration (SeaWiFS data) are shown.

of CO<sub>2</sub> during the summer 1998. The enhanced biological activity in 1998 may be related to a stronger stratification when ocean is warmer, to nutrient availability, or to the effect of light limitation (associated to cloud cover). All these processes are closely coupled.

[21] As opposed to the SIZ, a dramatic decrease of silicates was observed in the POOZ (Figure 6c and Table 2). The

interannual depletion in silicates ( $-10 \mu\text{mol kg}^{-1}$ ) was as large as the seasonal variations observed in the POOZ between summer and winter 1998. In summer 1998, surface silicate concentrations were very low, below  $2 \mu\text{mol kg}^{-1}$  (Figure 6c), suggesting that a strong diatom bloom occurred throughout the whole POOZ (50°S–57°S). This large decrease was also observed at depth (Figure 7) at the northern





**Figure 6.** (a) Sea surface NDIC, (b) nitrates, and (c) silicates, measured during OISO cruises in January 1998 (black squares) and January 2000 (shaded circles) in the Indian sector of the Southern Ocean. Arrows indicate the area taken into account to average data.

and southern boundaries of the POOZ (station 10 at 50°40'S and station 11 at 56°30'S). At 56°30'S, silicate concentration was about 7  $\mu\text{mol kg}^{-1}$  in the mixed layer (0–50 m). This contrasts with the silicate concentrations observed at the same stations in January 2000 (Figure 7c) and with the profiles reported before in the same region [Jeandel *et al.*, 1998; Pondaven *et al.*, 1998, 2000b; Louanchi *et al.*, 2001]. The reason for such low silicate concentrations in summer is not understood. Using a complex 1-D biogeochemical model, Pondaven *et al.* [1998, 2000a] tried to reproduce the temporal nutrients cycles at the Kerfix station. Their model was able to simulate the seasonal variations but was unable to capture the interannual variability and to reproduce the low silicate concentrations observed in February 1994. Biogeochemical models depend strongly on the choice of biological parameters. For example, different values for half-saturation constant for silicate uptake,  $K_{\text{SI}}$ , lead to very different results: a high  $K_{\text{SI}}$  value (8  $\mu\text{mol kg}^{-1}$ ) reproduces relatively well the average seasonal cycle of nutrients but

cannot reach very low silicate concentrations. On the other hand, a low  $K_{\text{SI}}$  value (1.04  $\mu\text{mol kg}^{-1}$ ) leads to low silicate concentrations all the year (this low  $K_{\text{SI}}$  value was not retained for standard simulations, see Figure 5a of Pondaven *et al.* [1998]). In high-nutrient surface waters in the POOZ and SIZ in the Atlantic sector, an intermediate  $K_{\text{SI}}$  value (4  $\mu\text{mol kg}^{-1}$ ) was selected for modeling a phytoplankton bloom during spring 1992 [Lancelot *et al.*, 2000]. The exceptionally low silicate concentrations we observed in summer 1998 suggest that standard and temporally independent value for biological parameters like  $K_{\text{SI}}$  is not appropriate for reconstructing the interannual primary production, nutrients, and carbon cycle in the POOZ.

[22] As for silicates, but to a lesser degree, the nitrate distribution observed in summer 1998 (Figure 6b) presents some differences compared to previous observations in the same sector [Le Jehan and Treguer, 1983; Pondaven *et al.*, 2000b]. The nitrates were not homogeneous as generally observed in the POOZ. They increase southward from about 23  $\mu\text{mol kg}^{-1}$  south of 50°S to 26  $\mu\text{mol kg}^{-1}$  at 57°S. Like for silicates and DIC, a nitrate depletion was also observed at depth in summer 1998 (Figure 7). Compared to climatologies of nutrients in the Southern Indian Ocean [Conkright *et al.*, 1994; Louanchi and Najjar, 2000], both nitrates and silicates observed during summer 1998 represent a relative large-scale anomaly in the POOZ as it was found for SST,  $f_{\text{CO}_2}$ , and air-sea CO<sub>2</sub> fluxes.

## 5. Reconstructing the Summer $f_{\text{CO}_2}$ Anomaly in the Permanent Open Ocean Zone

[23] In the previous section we discussed in a qualitative way the origin of the large oceanic CO<sub>2</sub> sink observed in the POOZ in summer 1998. We now turn to quantify these processes using a simple model adapted from the work of Louanchi *et al.* [1996] and Metzl *et al.* [1998]. The model is first forced by monthly SST [Reynolds and Smith, 1994] and SeaWiFS chl *a* data for the year 1998 by climatological winds [Hellerman and Rosenstein, 1983], by salinity, and by mixed-layer depth. The last two fields were constructed from OISO data. The data obtained in January, August, and December 1998 are used to validate the seasonal cycle of DIC, nitrates, and  $f_{\text{CO}_2}$ . We then change the forcing of SST

**Table 1.** Average of Oceanic  $f_{\text{CO}_2}$ , SST, Calculated Air-Sea Flux, and SeaWiFS chl *a* Concentrations in the POOZ (50°S–56°S) and the SIZ (58°S–60°S) in January 1998 and January 2000 and Their Differences<sup>a</sup>

Property	Area	1998	2000	1998–2000
$f_{\text{CO}_2}$ , $\mu\text{atm}$	POOZ	341 $\pm$ 6.7	355.6 $\pm$ 4.5	–14.6
	SIZ	379.3 $\pm$ 6.1	341.4 $\pm$ 8.8	37.9
SST, deg	POOZ	4 $\pm$ 0.5	3.5 $\pm$ 0.5	0.5
	SIZ	2.3 $\pm$ 0.28	1.2 $\pm$ 0.2	1.1
Air-sea flux, $\text{mmol m}^{-2} \text{d}^{-1}$	POOZ	–3.2	–1.3	–1.9
	SIZ	3.2	–3.8	7
chl <i>a</i> , $\text{mg m}^{-3}$	POOZ	0.51 $\pm$ 0.07	0.38 $\pm$ 0.08	0.13
	SIZ	0.16 $\pm$ 0.08	0.27 $\pm$ 0.08	–0.11

<sup>a</sup>For fluxes, positive (negative) values represent an oceanic CO<sub>2</sub> source (sink).

**Table 2.** Average of Sea Surface DIC, Normalized DIC, and Nutrients in the POOZ (50°S–56°S) and in the SIZ (58°S–60°S) Measured in Summer 1998 and 2000 During OISO 1 and 4 Cruises<sup>a</sup>

Property	Area	1998	2000	1998–2000	Number of Points in 1998	Number of Points in 2000
DIC	POOZ	2124.5 ± 6.4	2130 ± 6.7	−5.5	71	76
	SIZ	2144.7 ± 2.1	2140.2 ± 4.4	4.5	20	23
NDIC	POOZ	2194.4 ± 5	2199.4 ± 5.2	−5	71	76
	SIZ	2220.4 ± 4.7	2218 ± 5.1	2.4	20	23
Nitrates	POOZ	24.3 ± 0.6	26.1 ± 0.5	−1.8	7	8
	SIZ	26.7 ± 0.5	24.5 ± 0.7	2.2	3	3
Silicates	POOZ	1.9 ± 0.3	12.6 ± 3.8	−10.7	7	8
	SIZ	25.5 ± 3.5	27.8 ± 1.3	−2.3	3	3

<sup>a</sup>Also listed is the difference 1998–2000 of the mean data in each region (all in  $\mu\text{mol kg}^{-1}$ ). Number of data points is specified in the last two columns.

and chl *a* to the monthly fields for the year 1999, and we analyze the differences in  $f_{\text{CO}_2}$  between the two experiments to detail the effect of SST and primary production on  $f_{\text{CO}_2}$  variations.

### 5.1. Model Description

[24] The model was previously applied in different regions for seasonal and interannual analysis [Louanchi *et al.*, 1996, 1999; Metzl *et al.*, 1998; Goyet *et al.*, 1998]. Here, we recall the basic equations and coefficients used for this application and specify some changes from previous studies. The model, a two-boxes version, takes into account the main processes controlling the sea surface  $f_{\text{CO}_2}$  cycle: the air-sea gas exchange, the biological activity, the thermodynamics, and the mixing between the surface and subsurface ocean layers. The temporal variation of  $f_{\text{CO}_2}$  is deduced from equation (1):

$$\delta f_{\text{CO}_2} / \delta t = (\delta f_{\text{CO}_2})_{\text{airsea}} + (\delta f_{\text{CO}_2})_{\text{bio}} + (\delta f_{\text{CO}_2})_{\text{therm}} + (\delta f_{\text{CO}_2})_{\text{TA}} + (\delta f_{\text{CO}_2})_{\text{mix}}. \quad (1)$$

[25] At each time step the model calculates the variations of nitrogen and DIC in the surface box following

$$\delta N / \delta t = (\delta N)_{\text{bio}} + (\delta N)_{\text{mix}} \quad (2)$$

and

$$\delta \text{DIC} / \delta t = (\delta \text{DIC})_{\text{bio}} + (\delta \text{DIC})_{\text{mix}} + (\delta \text{DIC})_{\text{airsea}}. \quad (3)$$

[26] In equation (3),  $(\delta \text{DIC})_{\text{airsea}}$  corresponds to the variation of surface DIC due to air-sea gas exchanges. The air-sea CO<sub>2</sub> flux,  $F_{\text{CO}_2}$ , is computed from the relationship:

$$F_{\text{CO}_2} = \text{kw.sol.}(f_{\text{CO}_2\text{air}} - f_{\text{CO}_2\text{sea}}), \quad (4)$$

where kw.sol. is the product of the CO<sub>2</sub> solubility [Weiss, 1974] and the gas transfer velocity, kw, depends on wind speed [Wanninkhof, 1992]. The seasonal variation of atmospheric CO<sub>2</sub> concentration, being small at temperate and high latitude in the Southern Hemisphere, enables us to use the annual concentration of 365 ppm when calculating  $f_{\text{CO}_2\text{air}}$  the fugacity of CO<sub>2</sub> in the air using polynomials from the work of Weiss and Price [1980].

[27] In equations (2) and (3),  $(\delta N)_{\text{mix}}$  and  $(\delta \text{DIC})_{\text{mix}}$  represent the exchange between the surface and subsurface waters; the exchange is active only when the mixed layer deepens [Peng *et al.*, 1987].

[28] In equation (2)  $(\delta N)_{\text{bio}}$  represents the variation of inorganic nitrogen due to the biological activity, which is derived from the chl *a* concentrations following the expression:

$$(\delta N)_{\text{bio}} = -(U - rQ)R_{\text{N/Chl}}, \quad (5)$$

where the growth rate  $U = U_M N / (N + K_m)$  with  $U_M$  and  $K_m$  being the maximal and half-saturation constant of the growth rate, respectively. All the phytoplankton losses like grazing and mortality are included in the term  $Q$ , which is also based on chl *a* variations following the equation:

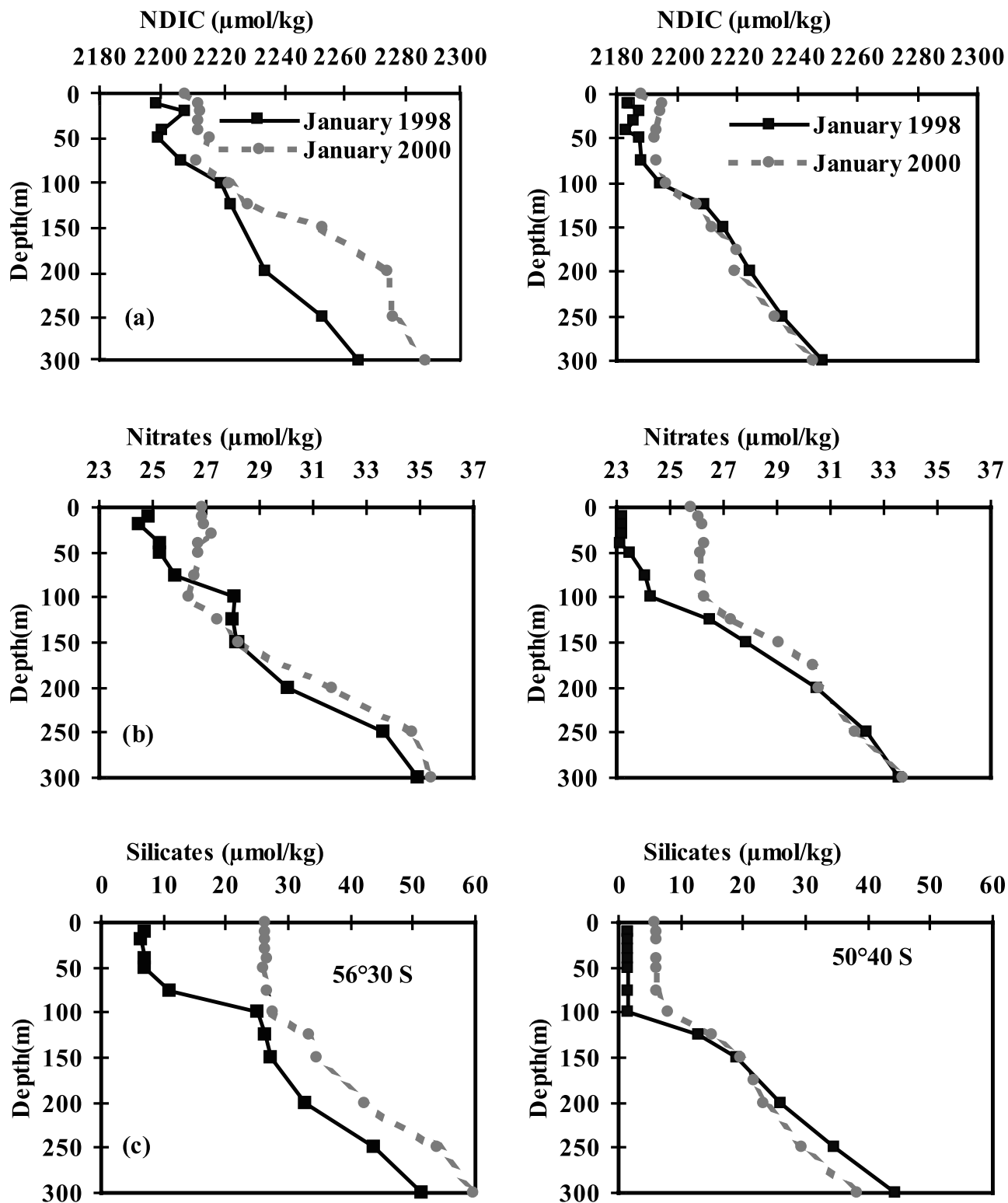
$$\delta \text{chl} / \delta t = (U - Q)\text{chl}. \quad (6)$$

[29] We derived  $(\delta \text{DIC})_{\text{bio}}$  from  $(\delta N)_{\text{bio}}$  with an additional term which takes into account the variation of DIC due to CaCO<sub>3</sub>, which was set to be proportional to the export production [Broecker and Peng, 1982]:

$$(\delta \text{DIC})_{\text{bio}} = R_{\text{C/N}}(\delta N)_{\text{bio}} - 0.2(1 - r)QR_{\text{C/Chl}}. \quad (7)$$

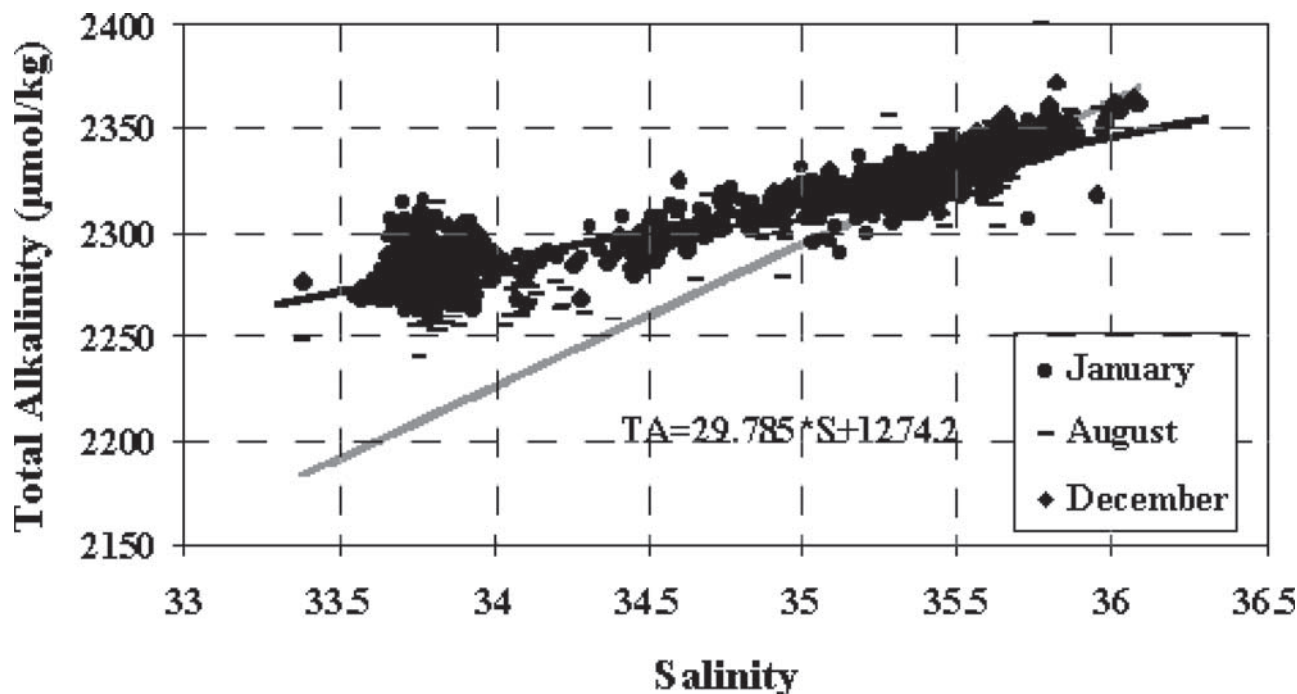
[30] In equations (5) and (7) the constants  $U_M$  (2 d<sup>−1</sup>),  $K_m$  (0.11  $\mu\text{mol kg}^{-1}$ ), and  $r$  (0.5) as well as Redfield ratios  $R_{\text{C/Chl}}$ ,  $R_{\text{C/N}}$  (40, 6) are those chosen by Louanchi *et al.* [1996]. Without any clear distinction about the phytoplankton assemblage in the studying area on both seasonal and interannual scales [Fiala *et al.*, 1998b; Kopczyńska *et al.*, 1998], we prefer to use relative standard values for  $U_M$ ,  $K_m$ ,  $r$ , and Redfield ratios (as used previously by Louanchi *et al.* [1996]) rather than tuning the parameters to obtain the best fit with the data.

[31] At each time step (1 day), when DIC is calculated, we need a TA value in order to calculate  $f_{\text{CO}_2}$ . In this application we are not seeking to simulate the TA cycle but instead we choose to derive the TA variations directly from the salinity using the TA/salinity relationship deduced from OISO observations. The TA/salinity relationship used in the model is shown in Figure 8 (TA = 29,785S + 1274.2). With no seasonality being clearly observed, it is possible to



**Figure 7.** (a) Profiles for the 300 upper meters of the water columns of NDIC and (b) nitrates and (c) silicates in  $\mu\text{mol kg}^{-1}$  in January 1998 (black squares) and January 2000 (shaded circles and dotted line) for two stations of the POOZ, (right) at the northern boundary, station 10 at 50°40'S, 68°5'E and (left) at the southern boundary, station 11 at 56°30'S, 63°E.





**Figure 8.** Relationship between total alkalinity and salinity at sea surface measured from OISO data during 1998 (solid circles for January, dashes for August, and diamonds for December),  $TA = 29.785S + 1274.2$ ,  $R^2 = 0.8709$ . For comparison, shading denotes the relationship from the work of *Millero et al.* [1998]:  $TA = 68.80S - 114$ .

use one equation to represent the annual TA cycle directly from salinity. We note that our calculated relation is quite different from the one calculated by *Millero et al.* [1998] for the Southern Indian Ocean, especially for salinity less than 34.5. This difference is largely due to the new TA observations we obtained in this sector.

[32] At each time step,  $f_{CO_2}$  in surface seawater is calculated from DIC and TA using the dissociation constants of carbonic acid published by *Goyet and Poisson* [1989]. Other sets of constants may be used but the same conclusions would be dressed for seasonal and interannual analysis.

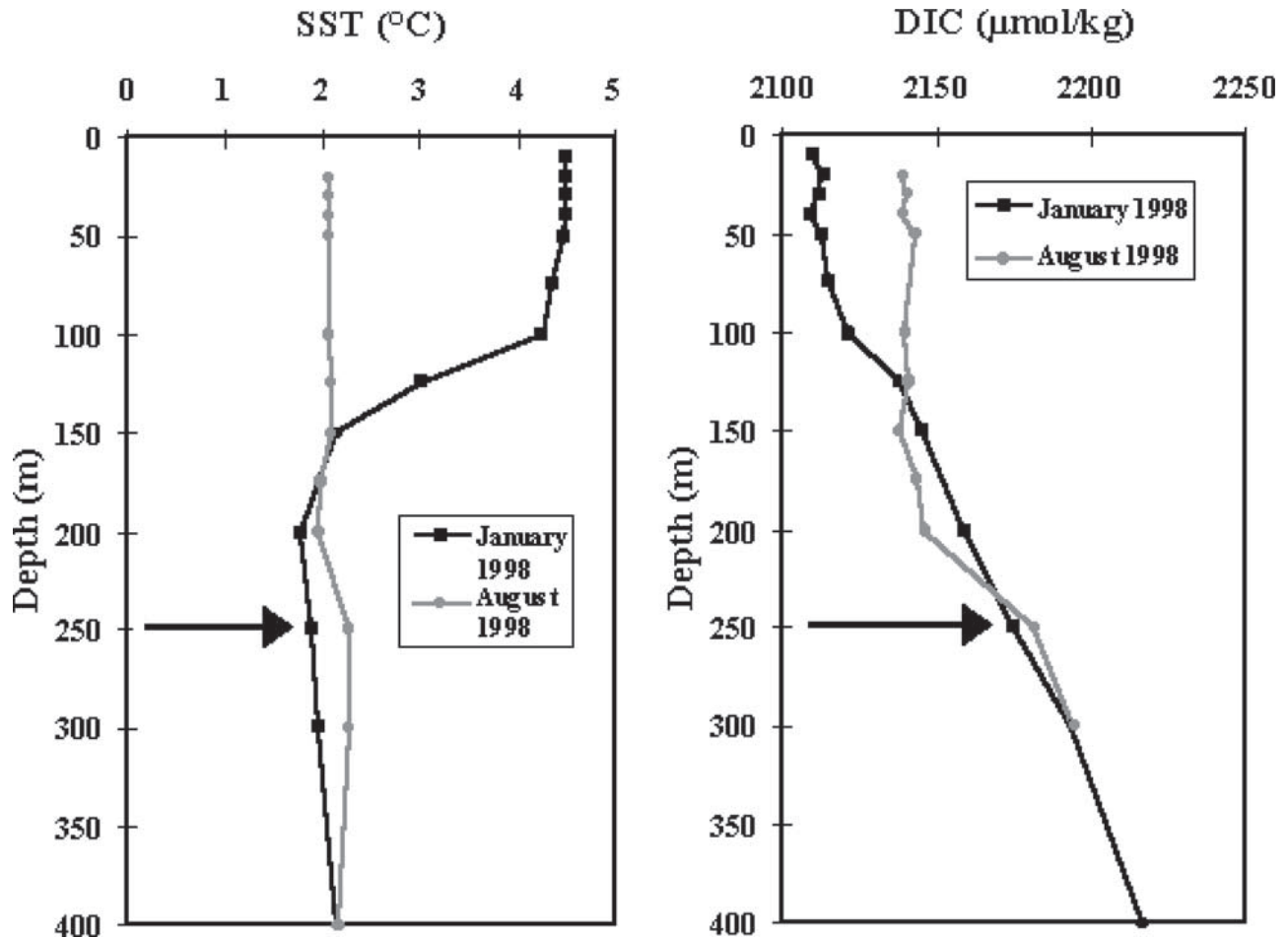
## 5.2. Validation of the Seasonal Biogeochemical Cycles

[33] To apply and validate the model in the POOZ, we select a location for which seasonal data are available for surface and water column. We thus select station OISO 10 (50°40'S–68°25'E), which is also the location of the Kerfix/JGOFS time series station where seasonal air-sea CO<sub>2</sub> fluxes have previously described for years 1992–1994 [*Jeandel et al.*, 1998; *Louanchi et al.*, 1999, 2001].

[34] The initial surface conditions are taken from observed DIC and nitrates in January 1998. Concentrations in the subsurface box are chosen at 250 m, below the winter mixed layer (Figure 9, see also Figure 7 for nitrates). At this depth, measured DIC at different seasons and years varies between 2170 and 2190  $\mu\text{mol kg}^{-1}$  and nitrates between 31.5 and 33  $\mu\text{mol kg}^{-1}$ . We tested the model with different subsurface DIC values according to the observations, but this did not affect the seasonal and interannual analysis. The

results presented below are those where DIC in subsurface box is set at 2190  $\mu\text{mol kg}^{-1}$ .

[35] The constraints of the model are presented in Figure 10. Mixed-layer depth has been constructed from OISO data and interpolated to follow the monthly mixed-layer climatology [*Levitus and Boyer*, 1994]. The mixed layer is minimum in February (50 m) and maximum in September (210 m) when temperature is low and winds are high. Salinity does not vary enough to create significant seasonal variation in TA and  $f_{CO_2}$ . Monthly chl *a* from SeaWiFS shows a clear maximum (0.5  $\text{mg m}^{-3}$ ) in January 1998. Results of the simulation for 1998 are shown in Figure 11 for nitrates, DIC, and  $f_{CO_2}$  (run noted R98). Compared to the observations in the mixed layer, the model reproduces relatively well the seasonal biogeochemical cycles. However, the seasonal amplitude in the simulation appears higher than observed. This is because the simulation creates a minimum of nitrates and DIC in February–March. Other biogeochemical models applied at Kerfix station also simulate a minimum of nutrients in March [e.g., *Pondaven et al.*, 1998]. The simulated nitrate concentration in end summer 1998 is below 20  $\mu\text{mol kg}^{-1}$ , which is low compared to observations obtained during the Kerfix program (nutrients' data available for years 1992–1994). Thus the R98 simulation represents a rather extreme anomaly in the summer nutrient concentrations in the POOZ, as previously discussed (section 4). As expected, the maximum nitrate concentrations are simulated during winter when mixed layer is deep. During this season, the simulated nitrate concentration was around 27  $\mu\text{mol kg}^{-1}$  as observed in August 1998.



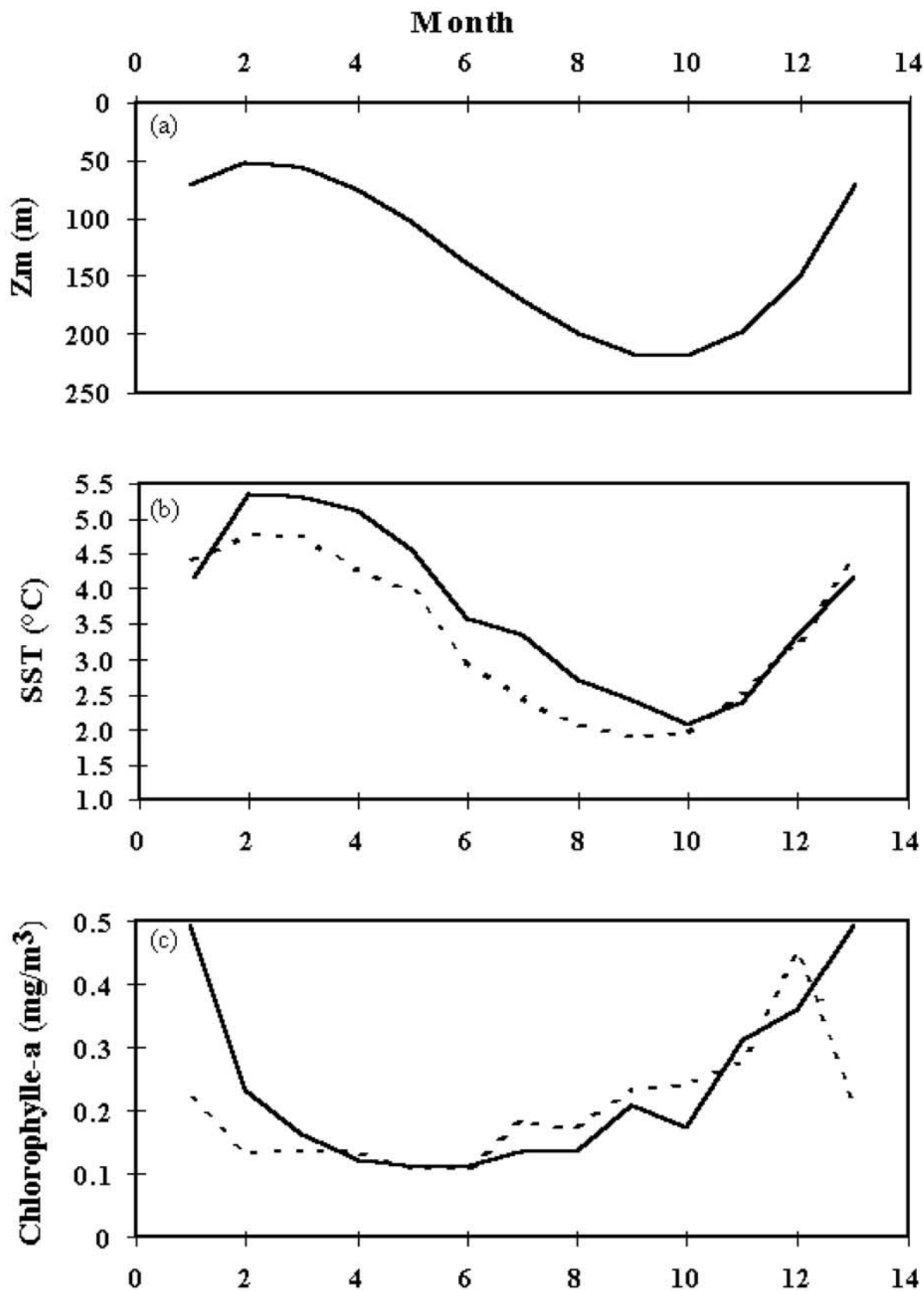
**Figure 9.** Profiles for station OISO-10 also known as KERFIX station (50°40'S, 68°5'E) of SST on the left and DIC on the right in summer 1998 in black and winter 1998 in gray. Arrow indicates the depth chosen for subsurface values.

This was also the winter nitrate concentration observed at Kerfix in years 1992–1994. As for nitrates, the simulated DIC and  $f_{\text{CO}_2}$  show a marked minimum in February–March and a maximum in winter (Figures 11b and 11c). Individual processes (each term in equation (1)) that are controlling the monthly variations of  $f_{\text{CO}_2}$  are presented in Figure 12. In summer, biological activity dominates the warming effect, whereas deep mixing is activated in April–August and dominates the effect of winter cooling. For most of the year, both air-sea CO<sub>2</sub> fluxes and alkalinity variations represent minor effects. Only during summer, when the potential sink is large, can the air-sea flux contribute to an increase of oceanic  $f_{\text{CO}_2}$  of around +5 to +10  $\mu\text{atm month}^{-1}$ . In comparison, biological activity contributes between –30 and –40  $\mu\text{atm month}^{-1}$  and warming affects  $f_{\text{CO}_2}$  by +10 to +20  $\mu\text{atm month}^{-1}$  (Figure 12).

[36] The simulated CO<sub>2</sub> sink is maximal in February–March (–10  $\text{mmol m}^{-2} \text{d}^{-1}$ ) when  $f_{\text{CO}_2}$  is low. The oceanic source is maximal in August (+5  $\text{mmol m}^{-2} \text{d}^{-1}$ ) when ocean  $f_{\text{CO}_2}$  is above equilibrium and winds are higher. The sum of monthly fluxes represents an annual CO<sub>2</sub> sink of 0.71  $\text{mol m}^{-2} \text{yr}^{-1}$ , much smaller than previous estimates up to 5  $\text{mol m}^{-2} \text{yr}^{-1}$  [Louanchi et al., 2001]. The seasonal

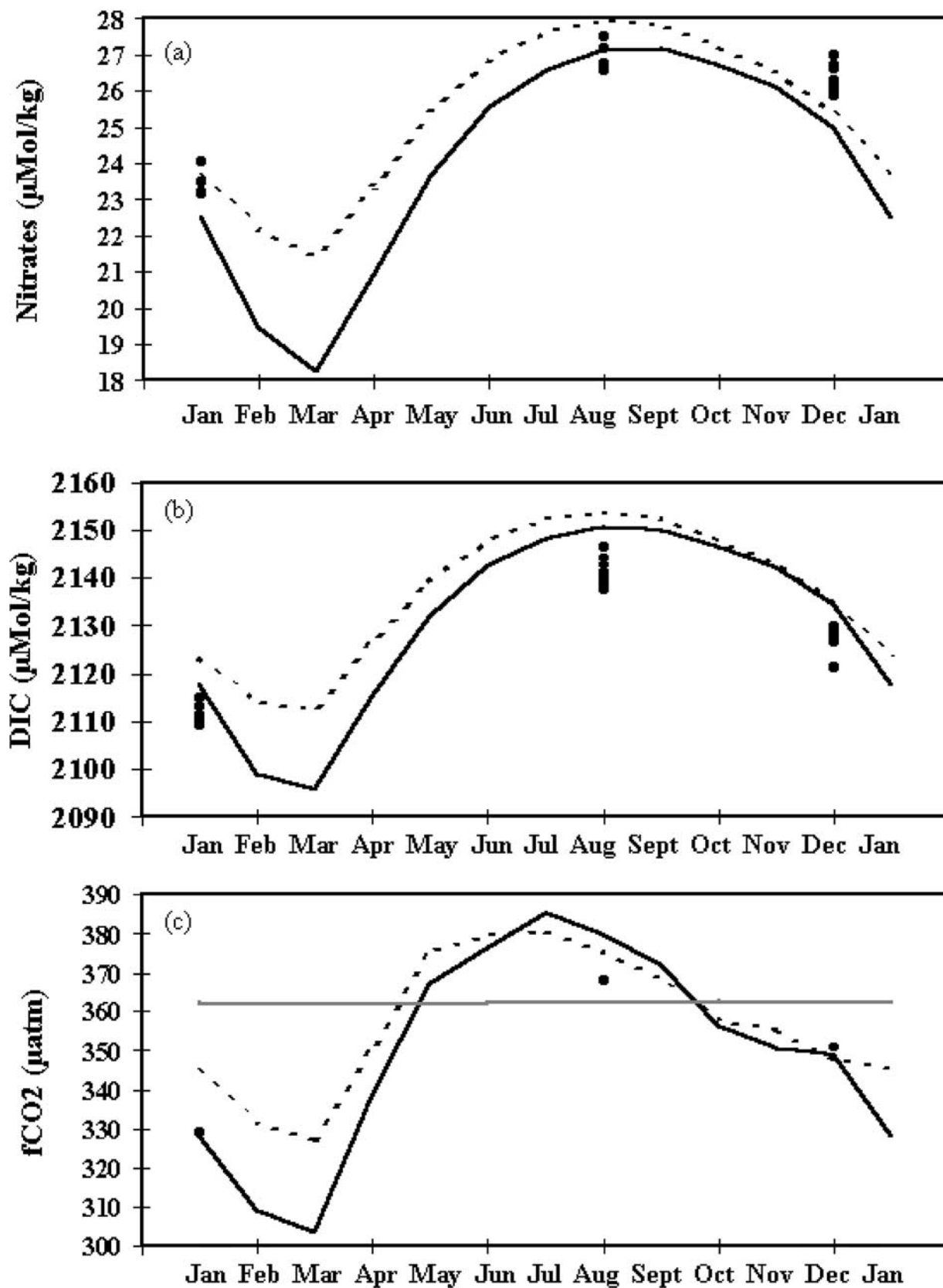
variations of sources and sinks of CO<sub>2</sub> in our simulation (and observations) are opposite to the  $f_{\text{CO}_2}$  cycle previously observed and simulated in the Indian Ocean POOZ at Kerfix station [Jeandel et al., 1998; Louanchi et al., 1999, 2001]. During the period 1991–1994, it was suggested that location was a permanent oceanic CO<sub>2</sub> sink, with the strongest sink during winter when winds are high and  $f_{\text{CO}_2}$  is low. Although we recognize that Kerfix station was sampled during a period when interannual temperature anomalies were low compared to 1998 (Figure 4), the difference in the  $f_{\text{CO}_2}$  seasonality cannot be interpreted as interannual variability. We suspect that contradiction is mostly related to DIC and TA corrections that must be applied to the original Kerfix data set. A comparison of deep DIC and TA measurements between Kerfix and OISO data indicates average difference of –35  $\mu\text{mol kg}^{-1}$  for DIC and –49  $\mu\text{mol kg}^{-1}$  for TA. Applying these corrections on the original Kerfix data of year 1993, the seasonal  $f_{\text{CO}_2}$  values calculated from DIC and TA would be about 348  $\mu\text{atm}$  in summer (a sink) and 360  $\mu\text{atm}$  in winter (a source).

[37] The results obtained in the POOZ lead to a seasonal picture which is similar to the SAZ around 40°S–45°S [Metzl et al., 1999]. In the SAZ, low oceanic  $f_{\text{CO}_2}$  is mainly

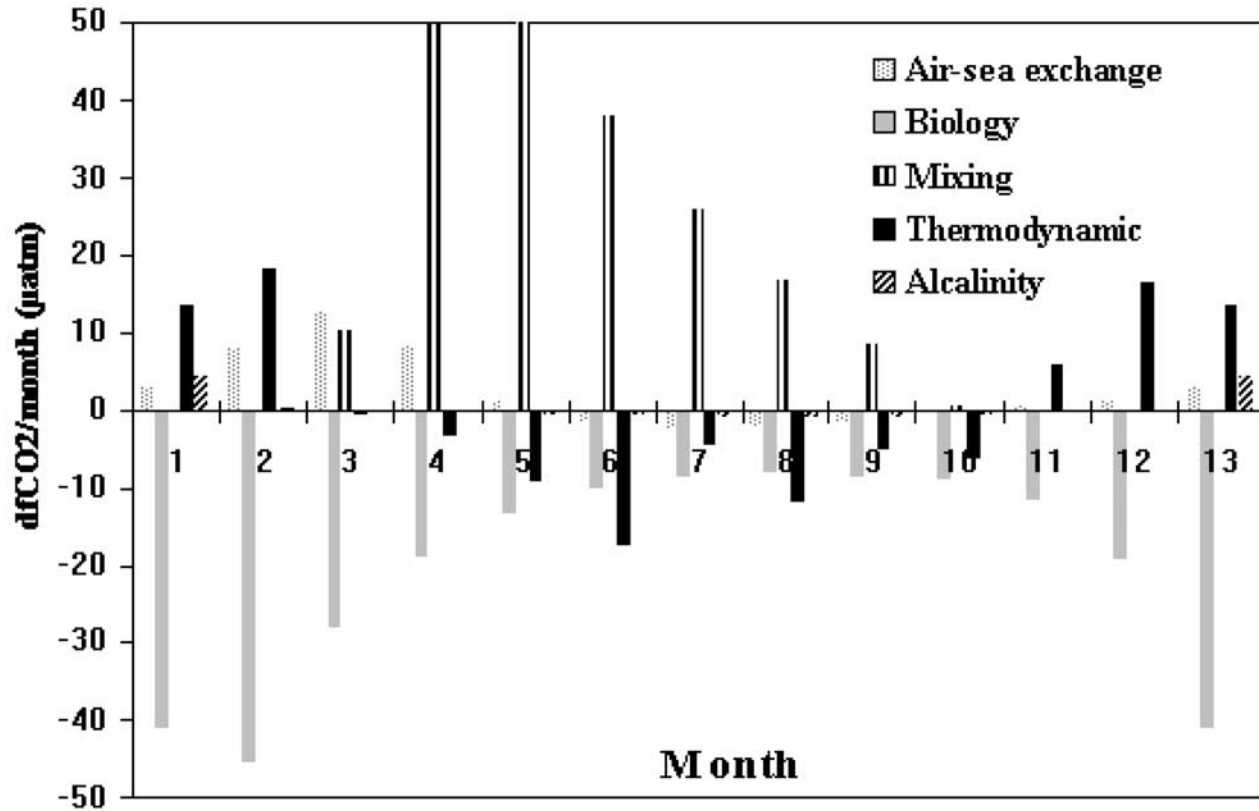


**Figure 10.** Monthly constraints used for the model applied at station 10 ( $50^{\circ}40'S$ ): (a) mixed-layer depth, (b) SST, and (c) chl *a* in 1998 (solid black line) and 1999 (dashed black line).





**Figure 11.** Simulated monthly evolution of (a) nitrates, (b) DIC, and (c) oceanic  $f\text{CO}_2$  for the year 1998 in solid black line and for the year 1999 in dashed black line and OISO data (dots) in the mixed layer at station 10. The solid gray line indicates the atmospheric  $f\text{CO}_2$ .



**Figure 12.** Monthly evolution of the processes (terms  $\delta f$  in equation (1)) involved in the evolution of the  $f_{\text{CO}_2}$  for the 1998 run.

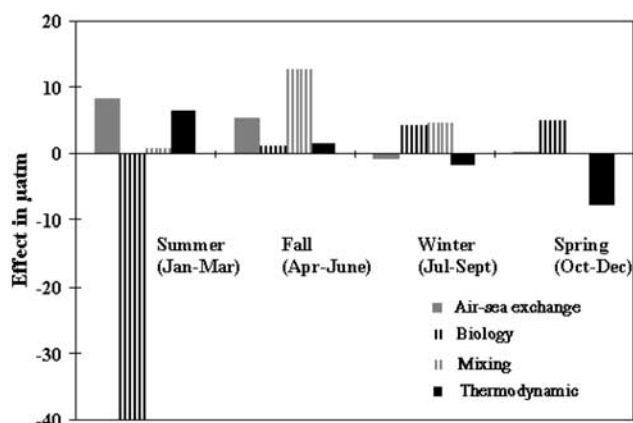
controlled by biological activity during summer and near-equilibrium or above atmospheric values in winter are associated with deep mixing with enriched DIC subsurface waters. For both SAZ and POOZ regions,  $f_{\text{CO}_2}$  and DIC seasonal variations are about 50  $\mu\text{atm}$  and 50  $\mu\text{mol kg}^{-1}$ , respectively. Observations about the same seasonal amplitude and phase of DIC in the POOZ and SAZ are coherent with the seasonal picture of the primary production (PP) deduced at large scale for the period October 1997–September 1998 from SeaWIFS data [Moore and Abbott, 2000]; in most of the high latitudes of the Southern Ocean, the maximum of the PP occurs in December–January. Although the annual PP is much higher in the SAZ, the seasonal variations are about the same in the SAZ and POOZ (Plate 7 of Moore and Abbott [2000]).

[38] The seasonality of the carbon dioxide system in the POOZ being relatively well reproduced, we now turn to analyze the interannual variations of  $f_{\text{CO}_2}$  and associated air-sea CO<sub>2</sub> fluxes.

### 5.3. Evaluating the Summer $f_{\text{CO}_2}$ Interannual Anomaly

[39] To evaluate the effect of warming and enhanced biological activity observed in summer 1998, we use the model described above in exactly the same configuration, at the same station, but we change the constraints of SST and satellite chl *a* to the data for the year 1999 (Figure 10).

[40] The new constraints show that chl *a* concentrations were higher in January–February 1998 and that SST was higher in 1998 from January to September 1998. For the MLD, as there are no monthly data during 1998–1999 we have to refer to dynamical models or previous long-term observations to test the sensitivity of our analysis. First, from our data, we have not detected significant variations of the MLD between January 1998 and January 2000. However, previous observations obtained at Kerfix station during 1990–1994 show that the interannual variability of the MLD could be significant ( $\pm 30$  m) [Louanchi *et al.*, 2001]. We also notice that although MLD can vary from year to year, the level of nutrient concentrations in winter remains relatively constant for the period 1992–1994. This could explain why the interannual variability of the winter MLD has very low impact on the interannual variability of the primary production in the POOZ. In this sector, the primary production is mostly dependent on anomalies during the timing of the bloom [Pondaven *et al.*, 2000a]. In addition, it has been suggested that the summer MLD does not significantly vary when warm anomalies, such as the one observed in 1998, do occur in the open ocean zone of the Southern Ocean [Markus, 1999]. Global ocean models suggest that the interannual variability of the mixed layer could be significant (20–30 m) in the Atlantic sector of the Southern Ocean but it appears to be lower, about 10 m, in the Indian sector [Le Quéré *et al.*, 2000].



**Figure 13.** Differences of the processes involved in the  $f_{\text{CO}_2}$  evolution between the run with 1998 constraints and the one with 1999 constraints.

According to these results, sensitivity experiments have been performed by reducing or increasing the depth of the mixed layer in our model. For example, in one test, the winter MLD has been increased up to 270 m. With such extreme conditions, the model (not shown) calculates unrealistic high  $f_{\text{CO}_2}$ , nitrates, and DIC concentrations during winter, but the summer DIC concentrations and  $f_{\text{CO}_2}$  values are not affected compared to the standard run (using the constraints MLD described in previous section). This is because when MLD is deep, the winter oceanic CO<sub>2</sub> source is larger, and the loss of CO<sub>2</sub> by air-sea exchange from winter to summer compensates the effect of winter mixing that supplies too high DIC and  $f_{\text{CO}_2}$ . Based on the sensitivity analysis, the conclusions concerning the inter-annual variability are not significantly different when changing the MLD, and results presented below correspond to the simulation without changing the mixed layer between the two runs for year 1998 and 1999.

[41] The main differences between R98 and R99 occur mainly during austral summer as expected. Results of the model with 1999 forcing show that nitrate concentration is minimum in March with a value around 22  $\mu\text{mol kg}^{-1}$  (Figure 11), close to what has been observed at Kerfix time series stations during summer in 1992–1994 [Pondaven *et al.*, 2000b; Louanchi *et al.*, 2001]. The simulated  $f_{\text{CO}_2}$  minimum of 330  $\mu\text{atm}$  in February–March for year 1999 corresponds well to the measurements made during end summer 1993 in the Indian POOZ [Robertson and Watson, 1995]. The differences between the two simulations are not constant. For nitrates it varies from 1.5  $\mu\text{mol kg}^{-1}$  in January to 3.2  $\mu\text{mol kg}^{-1}$  in March. For DIC, summer concentrations for year 1999 are also higher, by 6  $\mu\text{mol kg}^{-1}$  in January to 16  $\mu\text{mol kg}^{-1}$  in March. The differences for the month of January are relatively close to what was observed on average in the POOZ between a warm year and normal year (Table 2).

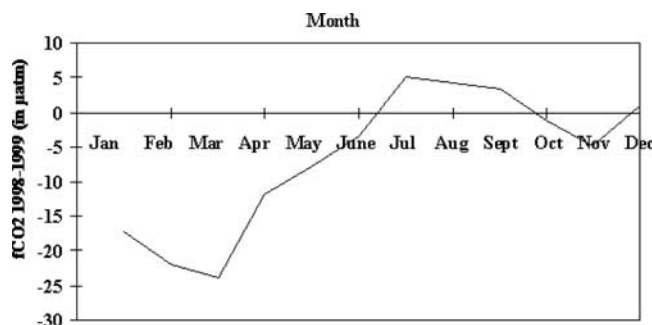
[42] The net primary production (NPP) of carbon can be deduced from the term  $(\delta\text{DIC})_{\text{bio}}$  (equations (3) and (7)). Assuming a depth of 50 m for an effective NPP, we obtain a total of 4.1  $\text{mol C m}^{-2} \text{yr}^{-1}$  (49.1  $\text{g C m}^{-2} \text{yr}^{-1}$ ) in 1999

and 5.0  $\text{mol C m}^{-2} \text{yr}^{-1}$  (60.9  $\text{g C m}^{-2} \text{yr}^{-1}$ ) in 1998. The model suggests that during the warm year 1998, the annual NPP is thus increased by 20%. The maximum NPP occurred in February 1998 and the minimum NPP occurred in August 1998. Although this is a local estimation, this seasonal amplitude (10  $\text{g C m}^{-2} \text{month}^{-1}$ ) is surprisingly close to the large-scale average values calculated by Moore and Abbott [2000] for the entire POOZ.

[43] We have not observed significant variations in salinity during OISO cruises conducted in 1998 and 2000. Therefore the total alkalinity, which is derived from salinity in our model, does not vary in both runs. The variations of  $f_{\text{CO}_2}$  are thus only controlled by changes in SST and DIC. The annual  $f_{\text{CO}_2}$  cycle for year 1999 follows the results of 1998 with higher values in summer. The changing contribution of each process can be deduced by calculating the difference of each term in equation (1) between the two scenarios. This is presented in Figure 13 where the differences have been summed for a seasonal view. The monthly  $f_{\text{CO}_2}$  differences are shown in Figure 14. Clearly, the most important change for the period January–March is related to the biological activity which represents 72% of the difference of  $f_{\text{CO}_2}$  (–40  $\mu\text{atm}$ ). It is balanced by an increase of  $f_{\text{CO}_2}$  due to air-sea exchanges (8  $\mu\text{atm}$  added because the sink is more pronounced in 1998) and by warming (+7  $\mu\text{atm}$ ). During the period April–June, although the mixed layer is the same in the two experiments, the mixing effect dominates the differences because it supplies more subsurface DIC and nutrients (the mixing between surface and subsurface water is proportional to the gradient between the surface and subsurface boxes). Overall, the oceanic CO<sub>2</sub> sink is smaller in 1999 simulation. The annual flux is  $-0.71 \text{ mol m}^{-2} \text{yr}^{-1}$  for 1998 against  $-0.26 \text{ mol m}^{-2} \text{yr}^{-1}$  in 1999. The summer anomaly in 1998 explains most of this difference.

## 6. Concluding Remarks

[44] In this paper we studied the variability of air-sea  $f_{\text{CO}_2}$  fluxes in five main oceanic regions in the Indian Ocean from 20°S to 60°S for years 1998 and 2000. For all sectors, it was observed that the ocean surface was warmer in 1998 than in 2000, a large-scale climatic signal associated with a change in the distribution on the oceanic  $f_{\text{CO}_2}$  and air-sea CO<sub>2</sub> fluxes.



**Figure 14.** Simulated  $f_{\text{CO}_2}$  difference between the run with 1998 constraints and the one with 1999 constraints.



[45] In the SBT western Indian Ocean, we observed an  $f_{\text{CO}_2}$  increase of 20  $\mu\text{atm}$  in summer 1998 relative to year 2000, a variation due to the warming likely governed by the subtropical dipole pattern in the South Indian Gyre. An important consequence of this warm anomaly in January 1998 is that the subtropical ocean was a CO<sub>2</sub> source, whereas it was previously observed, extrapolated, or simulated that the SBT region was a small sink or near equilibrium during this season [Poisson *et al.*, 1993; Metzl *et al.*, 1995, 1998; Louanchi *et al.*, 1996; Sabine *et al.*, 2000; Takahashi *et al.*, 2002]. An integration of the observed air-sea CO<sub>2</sub> flux anomaly in the SBT ocean in 1998 represents a source of 0.02 Gt C (for 3 months) over the region 25°S–37°S, whereas the estimate for 2000 is a sink of –0.01 Gt C. The difference of 0.03 Gt C applied for the SBT Indian Ocean may be significant to interpret the regional interannual variability of atmospheric CO<sub>2</sub> data in that sector, but the value is small compared to the uncertainties attached to interannual air-sea fluxes currently estimated at global scale from oceanic data and models ( $\pm 0.1$  to  $\pm 0.5$  Gt C yr<sup>–1</sup>) and from atmospheric data and models ( $\pm 0.5$  to  $\pm 2$  Gt C yr<sup>–1</sup>).

[46] During both years, 1998 and 2000, a large CO<sub>2</sub> sink was observed in the Sub-Antarctic Zone of the Indian Ocean, the strongest sink being always occurs at the fronts as was previously reported in this region [Poisson *et al.*, 1993; Metzl *et al.*, 1995, 1999]. In the frontal zone, although the surface ocean was warmer on average in summer 1998 (about +1°C), we did not detect a clear change in the  $f_{\text{CO}_2}$ . The reason for this may be due to the competing effect of thermodynamical and biological processes and variations in the position of the fronts which is difficult to track when comparing only 2 years.

[47] South of the Polar Front the interannual variability is clearly revealed by the data. First, the sea surface temperature shows a dramatic warming at high latitudes in 1998. This has a direct impact on the  $f_{\text{CO}_2}$  values in the SIZ (south of 58°S), which was acting as an oceanic CO<sub>2</sub> source in 1998 and CO<sub>2</sub> sink in 2000. However, in the POOZ (50°S–57°S), we found the opposite scenario, the ocean being a more important CO<sub>2</sub> sink during the warm period. The analysis of biogeochemical data and the results of a semiprognostic model both demonstrate that biological activity was enhanced in the POOZ during 1998. The low oceanic  $f_{\text{CO}_2}$  in summer 1998 is coupled to very low silicate concentrations over a large domain (50°S–57°S), suggesting that when specific external conditions occur, primary production can lower silicates by a much larger amount than previously observed in the POOZ. For external forcing, we notice that the dramatic change of hydrological and biogeochemical properties in summer 1998 occurs within a year or so of the marked 1997 ENSO event. A possible explanation of the biogeochemical and air-sea CO<sub>2</sub> flux anomalies in 1998 could be related to this event, which could impact a significant change in cloud cover over the Southern Ocean [Li, 2000]. The reduction of clouds would increase the available light and stimulate the phytoplankton bloom observed in summer 1998, decreasing silicates, inorganic carbon, and  $f_{\text{CO}_2}$ . Sensitivity-modeling studies [Markus, 1999] showed that a 5% decrease in cloud cover

may explain interannual SST variations up to 0.5°C during austral summer only, which is the season of maximum primary production in high latitudes. The contrasting results obtained in the POOZ and the SIZ of the Indian Antarctic sector described in this paper also suggest that the interannual change of the primary production is not homogeneous, likely because phytoplankton species are different in these regions [Fiala *et al.*, 1998b]. This highlights the need of introducing several phytoplanktonic species in biological models using adapted clouds and light forcing in order to study the response of different oceanic provinces when the atmospheric forcing is varying or when the climate is changing.

[48] Our observations in the high southern latitudes confirm what has been observed in high latitudes of the Northern Oceans. For example, based on data obtained by different groups, Murphy *et al.* [1998] concluded that large interannual variations of air-sea CO<sub>2</sub> fluxes occurred during the summer seasons (August–September 1985–1987) in the subarctic North Pacific and that variations are mostly attributed to photosynthesis. At high latitudes, in both hemispheres, it thus appears that the primary productivity is a key process to explain the natural variability of the air-sea CO<sub>2</sub> fluxes during summer months.

[49] The results presented in this paper would not be revealed without maintaining repeated oceanic observations (for the Southern Ocean, however, the use of VOS and merchant ships is not appropriate for obvious reasons). Long-term biogeochemical observations in the oceans are highly recommended to validate global models as well as to detect changes in oceanic properties in the context of climate change and carbon/climate coupling studies. Since the preindustrial period, we know that the surface temperature of the Earth is increasing. A recent synthesis of the data collected in the oceans showed also that the heat content in the ocean is increasing [Levitus *et al.*, 2000], and it has been later suggested that this warming is related to the increase of greenhouse gases [Levitus *et al.*, 2001]. For the next century, the increasing emissions of carbon dioxide in the atmosphere would tend to increase the ocean surface by one to several degrees. Although the warm anomaly we observed in summer 1998 in southern Indian Ocean represents an episodic event at decadal and planetary scales, observing such dramatic change of biogeochemical properties including carbonates system may help us to understand what could be the future oceanic  $f_{\text{CO}_2}$  distribution and air-sea CO<sub>2</sub> fluxes in a warmer climate.

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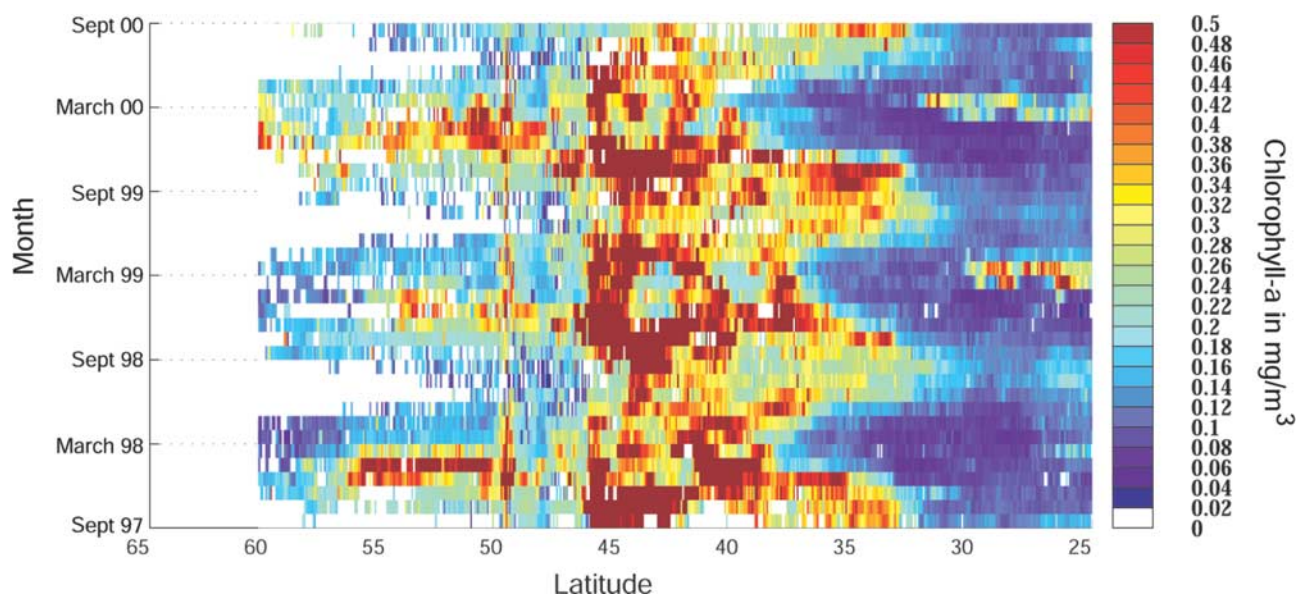
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**Figure 3.** Evolution along the western track of OISO cruises (from La Réunion to 60°S) of the monthly chl *a* concentration in  $\text{mg m}^{-3}$  (SeaWiFS level 3 binned data) from September 1997 to September 2000. Blank indicates no data.