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Impact of sinking carbon flux on accumulation of deep-ocean carbon in the Northern Indian Ocean

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Abstract The export of carbon through the biological pump from the surface to the deep ocean has a direct influence on the removal of CO_2 from the atmosphere. This is because the carbon is sequestered for only a few days to months in the surface while the carbon removed from the surface to deep waters takes hundreds of years to re-enter the atmosphere. The highest dissolved inorganic carbon (DIC) is expected in the deep waters of the North Pacific due to longer age of waters. On contrary, the higher deep water DIC is found in the northern Indian Ocean than elsewhere in the World Oceans. The sinking fluxes of particulate organic (POC) and inorganic carbon ($CaCO_3$) are found to be the highest in the northern Indian Ocean. The rates of bacterial respiration, organic carbon regeneration and inorganic carbon dissolution are also found to be the highest in the northern Indian Ocean than elsewhere. A most efficient biological pump appears to be operating in the northern Indian Ocean that transports surface-derived organic/inorganic carbon to deeper layers where it is converted and stored for longer times in dissolved inorganic form.

Keywords Arabian Sea · Carbon cycling · Regeneration · Sinking carbon fluxes

Introduction

The export of carbon from the surface to the deep ocean as a part of the biological pump has a direct implication on the removal of carbon-dioxide (CO_2) from the atmosphere. The biological pump consists of primary and export productions, and sinking flux. The carbon absorbed in surface waters may be returned to the atmosphere in a few days to months whereas carbon exported from surface to deep waters requires several hundreds of years to return to the atmosphere. The exported organic carbon is largely respired by bacteria resulting in its release back to the water column as dissolved inorganic carbon (DIC). However, a small portion of organic carbon reaches the sediments, where a part of it supports benthic biological activity and the rest is buried over geological time scales.

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The Arabian Sea is one of the most productive regions in the World Oceans (Qasim 1982; Bhattathiri et al. 1996; Barber et al. 2001). The primary and export productions exhibit strong seasonal and spatial variability in the Arabian Sea (Bhattathiri et al. 1996; Buesseler et al. 1998; Barber et al. 2001). Changes in primary production are found to be reflected by sinking fluxes at ~500 m above the bottom within a few days in the Arabian Sea (Honjo et al. 1999; Lee et al. 1998). This suggests a tight coupling between primary production and sinking organic carbon (Honjo et al. 1995; 1999). On the other hand, decoupling between source and sink was also observed during relatively high organic carbon sinking fluxes, associated with high-latitude blooms or seasonal and episodic export pulses in lower latitudes (Buesseler 1998). Nevertheless, the release of DIC in the deep ocean through decomposition/dissolution of sinking carbon by bacteria and zooplankton is important in the context of sequestration of carbon as the deep waters have higher residence times that are regionally variable. For instance, the residence time of deep water (> 1500 m) in the North Pacific is ~600 years, while it is 250 and 275 years in the Indian and Atlantic Oceans, respectively (Stuvier et al. 1983). Thus, it is important to examine the fate of sinking carbon to understand the role of the oceanic biological pump in the global carbon cycle. In this manuscript, accumulation of DIC in the deep waters of the northern Indian Ocean in relation to the sinking fluxes of carbon is assessed to find the impact of sinking fluxes on the deep water DIC levels and storage.

Data and methods

The present study was carried out using Joint Global Ocean Flux Study (JGOFS), World Ocean Circulation Experiment (WOCE), and other data available from the northern Indian Ocean, which were collected under different international programs (Sarma 1998). WOCE data are obtained from the WOCE Hydrographic Program Office (http:// whpo.ucsd.edu/) whereas US-JGOFS data are downloaded from http://usjgofs.whoi.edu/. The accuracies of temperature, salinity, DIC and Titration Alkalinity (TA) measurements during WOCE and US-JGOFS are estimated to be ± 0.002°C, ± 0.003, ± 3.0 and ± 5 μmol kg^{-1}, respectively (Johnson et al. 1999; Lamb et al. 2002) whereas that for DIC and TA are ± 4 and ± 6.2 μmol kg^{-1}, respectively, for Indian-JGOFS data sets (Sarma 1998). The σ_3 is the potential density referenced to a depth of 3 km and was computed following the equations given by Fofonoff and Millard (1983). In averaging the DIC and TA concentrations in the deep waters, we considered data for waters with density (σ_3) levels between 41.446 and 41.488 that occur between > 2500 m and 3800 m. We have taken care to avoid the possible effects of benthic fluxes by considering the data only above 500 m from the sediments.

The in situ oxygen utilization rate (OUR) is calculated according to following equation (Feely et al. 2004a):

$$\mathrm{OUR} = \mathrm{AOU}/{}^{14}\mathrm{C\ \ Age} \tag{1}$$

Apparent Oxygen Utilization (AOU) is calculated using the measured oxygen concentrations and the solubility estimated following Garcia and Gordon (1997). The age of the water mass is based upon the ^{14}C levels. Similarly the in situ organic carbon remineralization rate (OCRR) is calculated using

$$\mathrm{OCRR} = \Delta\mathrm{Corg}/\mathrm{Age} = \mathrm{Rc:o} * \mathrm{AOU}/\mathrm{Age} \tag{2}$$

where Rc:o is the C:O Redfield ratio and ΔCorg change in organic carbon.

Equation 2 is sensitive to the choice of C:O ratio used. The average C:O ratio for deep waters (2000–4500 m) of the Arabian Sea is found to be 0.8 ± 0.03 by Hupe and Karstensen (2000) whereas it is found to be 0.77 ± 0.02 for the Bay of Bengal and 0.70 ± 0.04 for the south Indian Ocean (Li and Peng 2002). Though no significant spatial gradients in C:O ratios were found in the deep waters of the Arabian Sea (Hupe and Karstensen 2000) large latitudinal variations in C:O ratios (from 0.67 to 0.78) were found in the South Indian Ocean (Li and Peng 2002). Therefore, an average C:O ratio was derived based on Li and Peng (2002) for the entire south Indian Ocean.

The complication in the age calculations using ^{14}C activity arises from the separation of natural and bomb-produced ^{14}C. Rubin and Key (2002) proposed the potential alkalinity method to achieve the separation. However, they found anomalous scatter in the relationship between ^{14}C and potential alkalinity caused by data from the northern Indian Ocean (north of equator) and attributed that to the possible transportation of bomb radiocarbon, as carbonate particles from the surface ocean to the sediment and their subsequent dissolution and mixing. Gordon et al. (2002) found nutrient and carbon rich, oxygen poor benthic layer in the bottom 100 m of the central and western Bay of Bengal at depths between 3400 and 4000 m. A closer look at the potential alkalinity to ^{14}C relation in the Indian Ocean reveals that the observed anomalous points represent the southwestern Bay of Bengal and the Oman coastal regions. Therefore, a regression equation was developed for the northern Indian Ocean here, after removing these anomalous points (Fig. 1). The constructed regression equation is

$$\text{Natural } \Delta^{14}\text{C} = -87.1 - 0.72(\text{Palk} - 2320) \qquad (3)$$

where Palk is potential alkalinity, which can be estimated using Eq. 4 and the factor 2320 is the global average surface ocean potential alkalinity (according to Rubin and Key (2002))

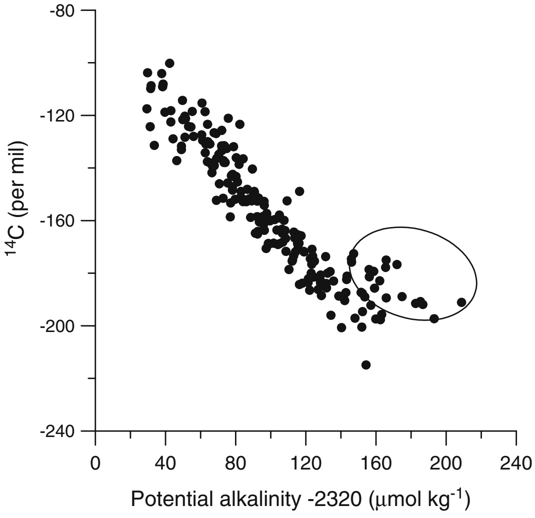

Fig. 1 Relationship between potential alkalinity (PALK-2320) to ^{14}C in the northern Indian Ocean. Anomalous points from the southwestern Arabian Sea and Oman coasts are circled and were excluded from the regression

$$\text{Palk} = (\text{Alkalinity} + \text{Nitrate}) \times 35/\text{Salinity} \qquad (4)$$

The rates of carbonate dissolution are computed using the equation given by Sabine et al. (2002)

$$\text{TA}* = 0.5(\text{NTA} - \text{NTA}_s) + 0.63(0.0941 \times \text{AOU}) \qquad (5)$$

where NTA is normalized total alkalinity (NTA = (TA × 35/S)), NTA$_s$ is normalized preformed alkalinity (following Sabine et al. (1999)), in which TA$_s$ is derived as a function of salinity, potential temperature (θ) and PO tracer of Broecker (1974) (PO $= \text{O}_2 + 170 \times \text{PO}_4$)

$$\text{TA}_s = 378.1 + 55.22S + 0.0716\text{PO} - 1.236\theta \qquad (6)$$

Results and discussion

Distribution of deep water DIC and TA in the World Ocean

Regional averages (Table 1) of deep-water (between density (σ3) levels of 41.446 and 41.488) DIC and TA concentrations (normalized to a salinity of 35.0, hereafter referred as NDIC and NTA, respectively) in the World Oceans are presented in Fig. 2. In the Pacific and Indian Oceans, deep water NDIC increased by 80 and 115 μ mol kg^{-1} from south of 60°S to north of 30°N (16–30°N), respectively whereas in the Atlantic, NDIC decreased by 55 μmol kg^{-1}. Similarly, NTA increased by 67 and 104 μmol kg^{-1} in the Pacific and Indian Oceans, respectively, but decreased by 18 μmol kg^{-1} in the Atlantic Ocean. As a water mass moves along an isopycnal surface its DIC content is expected to increase through decomposition of biogenic particles, such as POM, CaCO$_3$ etc., sinking from above. In general, for 1 mole of CaCO$_3$ dissolves ~15 moles of POM remineralizes (Sarmiento et al. 2002). Consequently, the highest TA and DIC should occur in the oldest deep waters. In addition to this, the path of the water mass from its source region and the initial concentrations of each parameter, when the water mass sinks, is also important. Figure 2 suggests that NDIC, NTA

Table 1 The average (± standard deviation) of salinity normalized DIC, TA and AOU (μmol kg^{-1}) in different latitudinal belts between σ_3 levels of 41.446 and 41.488 in the world oceans

Region	Position	NDIC	NTA	AOU
Atlantic	> 60°N	2158 ± 1 (6)	2312 ± 1 (6)	48 ± 4 (11)
	46–60°N	2168 ± 12 (223)	2325 ± 13 (147)	64 ± 11 (378)
	31–45°N	2178 ± 13 (84)	2333 ± 14 (81)	75 ± 10 (157)
	16–30°N	2180 ± 11 (93)	2337 ± 10 (97)	81 ± 8 (244)
	0–15°N	2184 ± 8 (95)	2334 ± 9(71)	85 ± 8 (273)
	0–15°S	2199 ± 13 (462)	2356 ± 16 (248)	96 ± 12 (1291)
	16–30°S	2199 ± 12 (359)	2344 ± 11 (142)	93 ± 8 (1415)
	31–45°S	2212 ± 14 (108)	355 ± 6 (44)	107 ± 11 (382)
	46–60°S	2261 ± 12 (21)	–	144 ± 8 (100)
	>60°S	2235 ± 10 (12)	2368 ± 7 (43)	45 ± 13 (28)
Indian	>60°S	2266 ± 14 (63)	2371 ± 6 (52)	131 ± 14 (78)
	46–60°S	2262 ± 9 (105)	2370 ± 6 (90)	148 ± 11 (188)
	31–45°S	2266 ± 12 (100)	2384 ± 11 (100)	145 ± 10 (212)
	16–30°S	2301 ± 10 (285)	2410 ± 9 (289)	169 ± 5 (625)
	0–15°S	2317 ± 8 (349)	2427 ± 9 (348)	181 ± 4 (732)
Arabian Sea	0–15°N	2341 ± 6 (188)	2443 ± 7 (189)	194 ± 5 (530)
	16–30°N	2381 ± 5 (252)	2475 ± 8 (275)	226 ± 6 (358)
Bay of Bengal	0–15°N	2331 ± 6 (122)	2444 ± 8 (122)	197 ± 5 (277)
	16–30°N	2359 ± 2 (12)	2448 ± 4 (12)	207 ± 2 (13)
Pacific	>60°S	2269 ± 7 (194)	2369 ± 10 (3)	135 ± 13 (339)
	46–60°S	2280 ± 9 (173)	2384 ± 7 (10)	156 ± 7 (483)
	31–45°S	2310 ± 14 (248)	2398 ± 3 (22)	172 ± 8 (617)
	16–30°S	2321 ± 12 (196)	2421 ± 8 (140)	175 ± 7 (607)
	0–15°S	2340 ± 10 (96)	2437 ± 7 (59)	191 ± 11 (363)
	0–15°N	2347 ± 10 (215)	2450 ± 7 (151)	198 ± 7 (725)
	16–30°N	2349 ± 8 (673)	2436 ± 10 (464)	194 ± 8 (1128)
	31–45°N	2354 ± 9 (386)	2450 ± 12 (265)	202 ± 11 (1090)
	46–60°N	2355 ± 11 (157)	2448 ± 11 (125)	207 ± 8 (618)

The number of data points used for averaging is given in parenthesis. See material and methods for data sources

and AOU concentrations are almost similar at 60°S in the Indian and Pacific oceans but were lower by 3–44, 5–14, and 3–27 μmol kg^{-1}, respectively, in the Indian Ocean compared to that in the Pacific Ocean from north of 45°S to 10–15°N. However, there was a sharp increase in their values to the north of 15°N in the Arabian Sea and Bay of Bengal. Broecker (1991) suggested that the North Pacific deep waters contain the highest DIC and TA contents in the World Ocean, as these are the oldest waters (~600 years). In contrast, higher NDIC, NTA and AOU by 32 ± 9, 39 ± 12, and 32 ± 10 μmol kg^{-1}, respectively are found to occur in the Arabian Sea deep-water compared to that in the North Pacific along the similar latitudinal belts (15–30°N). Two potential mechanisms can enhance deep water NDIC in the Arabian Sea, of which the first is the residence time of deep waters and DIC release with time, and the second is the higher sinking fluxes of particulate carbon (both organic and inorganic) and their subsequent remineralization and dissolution at deeper depths.

The DIC concentrations shall be proportional to the residence time/age of deeper waters, if there are minimal changes in sinking fluxes and decomposition/dissolution rates. Due to geographical setting the deep waters of the Northern Indian Ocean are expected to be older, as the oceanic connection exists only to the south, and hence might accumulate DIC. In order to examine such a possibility, ^{14}C activity is considered from selected locations in the Atlantic, Pacific, and Indian Oceans (Fig. 3). The ^{14}C activity is higher (~ –50 per mil) in the Atlantic but lower (~ –190 and –223 per mil) in the North Indian and North Pacific Oceans, respectively. This trend confirms that North Pacific deep waters are older followed by that of North Indian and Atlantic

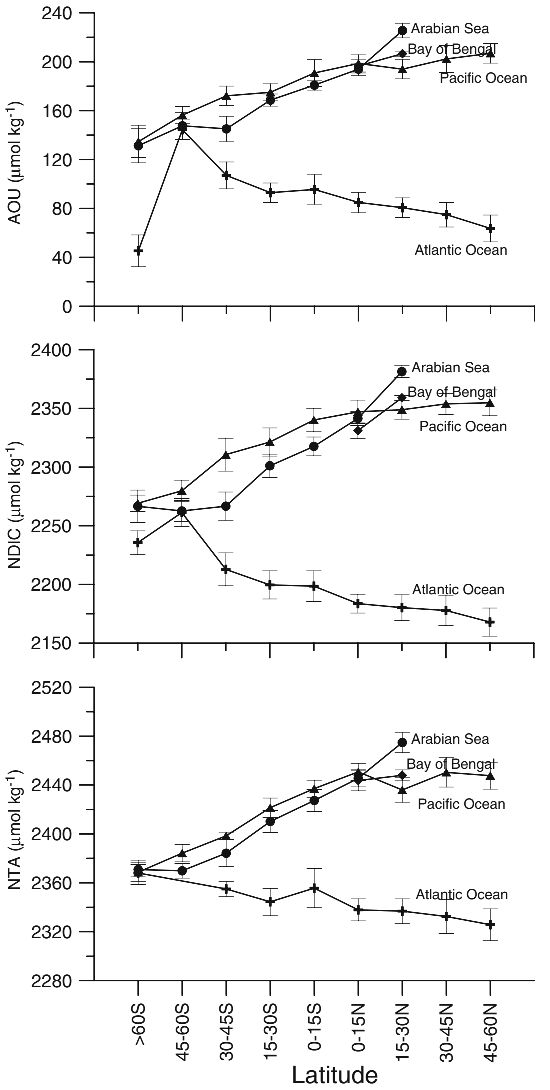

Fig. 2 Distributions of NDIC, NTA and AOU in the deep waters (between σ_3 levels of 41.446–41.488) in the Pacific (closed triangles), Atlantic (plus), Indian Ocean and Arabian Sea (closed circles) and Bay of Bengal (closed diamonds)

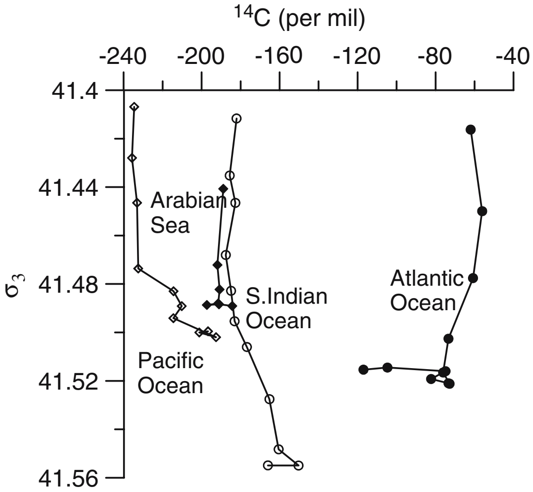

Fig. 3 Vertical distributions of ^{14}C (per mil) in deep waters at selected regions of the World Oceans. Pacific (34°N, 165°E; open diamonds), Atlantic (40°N, 52°W; closed circles), Arabian Sea (16.8°N, 72°E; closed diamonds), south Indian Ocean (18.5°S, 55°E; open circles)

Oceans. Broecker et al. (1988) found radiocarbon ages for deep waters of the North and South Atlantic to be 314 ± 153 and 534 ± 98 years, respectively, that are younger than those of the Indian Ocean (1260 ± 35 and 1252 ± 42 years in the north and south, respectively) and that the North Pacific waters are the oldest (1759 ± 176 years). Despite a low radiocarbon age of Northern Indian Ocean deep water its DIC is significantly higher than in the North Pacific (Fig. 2; Table 1). Though DIC increases because of the continuing remineralization with the aging of deep-water mass, higher accumulation of DIC in the North Indian than in the North Pacific Ocean is not accounted just by aging but requires invoking of sinking fluxes of carbon and their decomposition/dissolution in deep waters.

The sinking flux of POC, measured at ~500 m above the bottom, is the highest in the Northern Indian Ocean (3–13.3 mg m^{-2} d^{-1}; Haake et al. 1993; Lee et al. 1998; Honjo et al. 1999; Ittekkot et al. 1991) compared to other regions in the World Oceans (Table 2). Sinking POC fluxes are found in the North Pacific to be (between 12°S and 50°N) 0.43–5.2 mg m^{-2} d^{-1} (Dymond and

Table 2 Sinking POC to the deep sea (> 2500 m) in the World Oceans (mgC m^{-2} d^{-1}) (source: Lutz et al. 2002)

Region	Sinking flux Minimum	Sinking flux Maximum
N. Atlantic	0.07	5.5
Eq. Atlantic	0.81	6.3
S. Atlantic	0.4	6.5
NC. Pacific	0.43	3.4
Eq. Pacific	0.7	5.2
S. Pacific	0.72	3.6
Arabian Sea	3.0	13.3
Bay of Bengal	5.6	7.3

Collier 1988; Kempe and Knack 1996; Honjo et al. 1995) and in the Atlantic (between 56°S and 75°N) 0.07–6.5 mg m^{-2} d^{-1} (Honjo 1980; Pudsey and King 1997; Usbeck 1999). Inorganic carbon fluxes were also higher in the northern Indian Ocean (4–13 mg m^{-2} d^{-1}) than elsewhere in the World Oceans (1.9–7.7 mg m^{-2} d^{-1}; Honjo et al. 1999). These results suggest higher rates of particulate carbon reaching the deep waters in the Northern Indian Ocean than elsewhere in the oceans.

Buesseler (1998) found higher *ThE* ratios (>10 to 50%; *ThE* is the ratio of primary production to POC export derived from ^{234}Th in the euphotic zone and at the base of the euphotic zone, respectively) of sinking fluxes at the base of the euphotic zone in the high-latitude, during blooms in the North Atlantic and in the Arabian Sea during SW monsoon. Such elevated *ThE* ratios were attributed to either active biological pump or time lag between phytoplankton bloom to zooplankton grazing. On the other hand, *ThE* ratio is generally low and range between 2 and 10% elsewhere in the oceans. The *p* and *s* ratios, defined as normalized sinking flux ratios of organic carbon with respect to primary and export productions, respectively, would help examine the efficiency of retention of sinking organic carbon, and the storage within the deep ocean (Lutz et al. 2002). The high *s* and *p* ratios in the Arabian Sea suggest that this region is one of the high storage efficiency regions in the World Oceans (Lutz et al. 2002). High storage efficiency is indicative of deep waters receiving a large fraction of sinking organic carbon. Francois et al. (2002) postulated that carbonate dominated low latitude regions have higher transfer efficiency to the twilight zone through export of tightly packaged refractory organic matter, which is less prone to hydrodynamic processes. On the contrary, diatom dominated regions export more labile organic matter susceptible to degradation in the mesopelagic zone. Klaas and Archer (2002) suggested that about 83% of the organic carbon flux to the seafloor can be accounted for by $CaCO_3$ ballast and the rest is associated with opal and clay. Since Arabian Sea has higher fraction of carbonate (14.0–42.3 g m^{-2} y^{-1}) than opal (2.5–15.7 g m^{-2} y^{-1}) in sinking fluxes (Nair et al. 1989; Honjo et al. 1999) the organic carbon reaching to the mesopelagic zone is expected to be more refractive in nature.

In the northern Indian Ocean, significant changes in deep water DIC are found in the northwestern (Arabian Sea) and northeastern (Bay of Bengal) Indian Ocean. The deep water DIC is higher by 22 ± 5 μmol kg^{-1} in the Arabian Sea than Bay of Bengal (16–30°N; Fig. 2, Table 1) despite more or less uniform (3.0–13.3 mgC m^{-2} d^{-1}; Table 2) sinking fluxes of POC in both the regions, except in the western Arabian Sea (13.3 mgC m^{-2} d^{-1}) where higher fluxes resulted from coastal upwelling. Naqvi et al. (1996) found respiration rates lower by 2–10 μmol m^{-3} d^{-1} in the Bay of Bengal in the upper 1500 m compared to that in the Arabian Sea. Despite higher sinking fluxes of POC in the Bay of Bengal, low deep water DIC than Arabian Sea might have resulted from scavenging of POC by mineral particles in to the sediments as explained by Ittekkot et al. (1991) and Kumar et al. (1998). The numbers of lithogenic particles are 6–10 times higher in the Bay of Bengal (23.4–76.6 mg m^{-2} d^{-1}; Ittekkot et al. 1991) than in the Arabian Sea (2.9–10.5 mg m^{-2} d^{-1}; Nair et al. 1989 and Honjo et al. 1999). Unger et al. (2005) found that higher amounts of hydrolysable carbohydrates (CHO) reach the deep Bay of Bengal (0.2 to > 2 mg m^{-2} d^{-1}) compared to that of other oceanic regions (0.1–1.1 mg m^{-2} d^{-1}). Similarly such high concentrations of CHO are also found in Subarctic Pacific (2.0 mg m^{-2} d^{-1}). This study further suggests that undecomposed and easily degradable organic carbon reaches the deep Bay of Bengal through scavenging by mineral ballast. This is consistent with the suggestion of Broecker et al. (1980) that large quantities of labile organic matter reach the deep sea floor in the Bay of Bengal leading to near-bottom anomalies.

Despite the minor differences in sinking carbon between the Arabian Sea and the Bay of Bengal the higher deep water DIC in the northern Indian Ocean could have resulted from decomposition (respiration) and dissolution of sinking particulate carbon. Upwelling in summer and convective mixing in winter pump nutrients into the surface waters and result in high primary production in this region (Madhupratap et al. 1996a; Barber et al. 2001). The mesozooplankton in the upper 300 m is supported by phytoplankton during the productive summer seasons but by the microbial loop during oligotrophic periods (Madhupratap et al. 1996b). Therefore, Madhupratap et al. (1996b) found minimal seasonal variability in the zooplankton biomass in the Arabian Sea. Several investigators observed a strong coupling between sinking organic carbon fluxes and bacterial abundance in the deep waters of the World Oceans (Patching and Eardly 1997; Nagata et al. 2000; Hansell and Ducklow 2003). The integrated bacterial production between 1000 and 4000 m is several folds higher in the Arabian Sea (96–360 (mean 240) mgC m^{-2} d^{-1}; Hansell and Ducklow 2003) than in the North Pacific (0.11–1.47 (mean 0.92) mgC m^{-2} d^{-1}; Nagata et al. 2000) based on Leucine incorporation method using similar carbon conversion factor. Koppelmann et al. (2000) also reported higher bacterial production in the deep waters of the Arabian Sea (0–300 mgC m^{-2} d^{-1} between 1000 and 3000 m and 0–70 mgC m^{-2} d^{-1} in 3000–4000 m). Assuming a bacterial growth efficiency of 20% (del giorgio and Cole 1988) in the deep waters of the North Indian and North Pacific Oceans, the bacterial respiration amounts to 404–1440 (mean 960) and 0.44–5.88 (mean 3.68) mgC m^{-2} d^{-1}, respectively. This strongly suggests that bacterial respiration rates are several folds higher in the Arabian Sea than elsewhere result in the accumulation of higher DIC. However, large uncertainties are involved in the conversion of bacterial abundance to rates of production and respiration due to wide range of conversion factors available. Therefore, we also estimated the rates of decomposition of organic carbon and dissolution of $CaCO_3$ and compared them with rates for other regions of the World Ocean in the following text. Further, though several studies have been made on the rates of organic carbon decomposition in deep waters of the Pacific and Atlantic Oceans, no such attempts have been made so far, to our knowledge, in the Arabian Sea and Bay of Bengal. Here, we estimate oxygen utilization and organic carbon decomposition rates using apparent oxygen utilization (AOU) due to remineralization of organic matter and age of water masses using Eqs. (1) and (2).

Figure 4 shows the mean rates of oxygen utilization, organic carbon decomposition and inorganic carbon dissolution in the Arabian Sea, Bay of Bengal (north of equator) and south Indian Ocean (0–30 °S) between σ_3 levels of 41.0–41.5. The OUR rates were higher in the Arabian Sea (0.114–0.236 ± 0.003 μmol kg^{-1} y^{-1}; $n = 357$) than in the Bay of Bengal (0.092–0.167 ± 0.002 μmol kg^{-1} y^{-1}; $n = 137$) and the south Indian Ocean (0.090–0.151 ± 0.004 μmol kg^{-1} y^{-1}; $n = 906$). Moreover, the OUR in the Arabian Sea is higher than elsewhere in the world oceans. For instance, it is 0.13 ± 0.002 μmol kg^{-1} y^{-1} in the North Pacific (2 km above bottom; Chen 1990), 0.03–0.14 in the eastern equatorial Pacific (Kroopnick 1974), 0.05–0.08 μmol kg^{-1} y^{-1} in the South Pacific (Wyrtki 1962; Craig 1971), 0.09–0.11, and 0.09–0.11 μmol kg^{-1} y^{-1} in the North (Arons

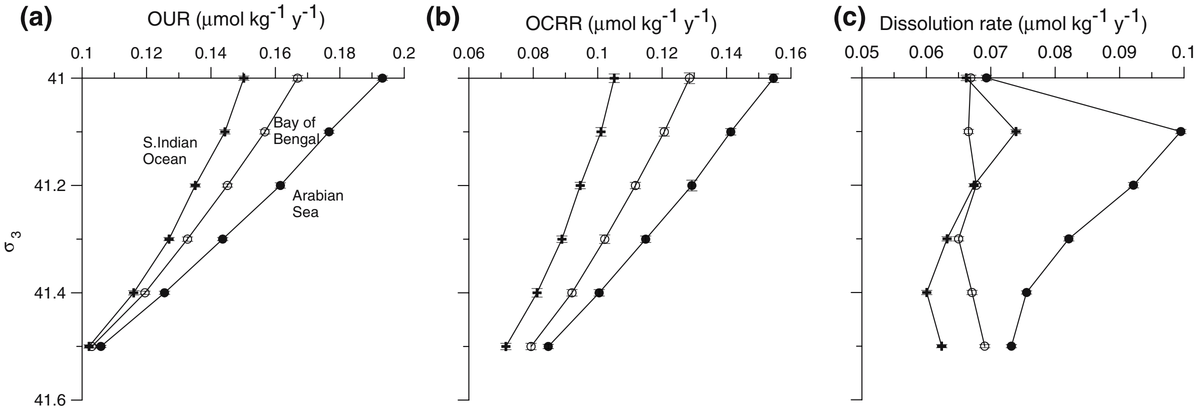

Fig. 4 Vertical distributions of (**a**) oxygen utilization rate (OUR), (**b**) organic carbon remineralization rate (OCRR) and (**c**) carbonate dissolution rates in the Arabian Sea (closed circles), Bay of Bengal (open circles) and southern Indian Ocean (plus) between σ3 levels of 41.0–41.5. Error bars are smaller than the symbols used

and Stommel 1967) and South Atlantic (Wright 1969). Similarly OCRR are also higher in the Arabian Sea between σ_3 levels of 41.0–41.5 (mean of 0.126 ± 0.01 μmol kg^{-1} y^{-1}) compared to that in the Bay of Bengal (0.093 ± 0.02 μmol kg^{-1} y^{-1}) and the southern Indian Ocean (0.081 ± 0.02 μmol kg^{-1} y^{-1}). These rates steadily decreased from σ_3 levels of 41.0 with depth in all three basins. OCRR in the Arabian Sea are again higher than the same found elsewhere. For instance it was 0.107 μmol kg^{-1} y^{-1} in the North Pacific (Chen 1990), 0.05–0.073 μmol kg^{-1} y^{-1} in the Atlantic (Arons and Stommel 1967; Wright 1969). Similarly the calcium carbonate dissolution rates were also higher between σ_3 levels of 41.0–41.5 in the Arabian Sea (~0.081 ± 0.004 μmol kg^{-1} y^{-1}; n = 357) than that of the Bay of Bengal (0.067 ± 0.002 μmol kg^{-1} y^{-1}; n = 137) and the South Indian Ocean (~0.064 ± 0.003 μmol kg^{-1} y^{-1}; n = 906). The estimated dissolution rates are consistent with the recent estimates by Sabine et al. (2002). The slight difference (by 0.02 μmol kg^{-1} y^{-1}) between these two estimates is due to categorization of south and northern Indian Oceans as Sabine et al. (2002) separated north and south Indian Ocean using 20°S as the boundary whereas we used the Equator in the present study. The summary of the rates of decomposition and dissolution in the world oceans are given in Table 3. The mean dissolution rate in the Arabian Sea (~0.081 μmol kg^{-1} y^{-1}) is higher than that found in the North Pacific (~0.051 μmol kg^{-1} y^{-1}; Feely et al. 2002a). However these rates are much smaller than North Atlantic (0.3–0.75 μmol kg^{-1} y^{-1}; Feely et al. 2004b) due to existence of aragonite saturation horizon in the deep waters (> 2500 m) in the North Atlantic whereas it is < 500 m in the North Pacific and Northern Indian Ocean (Feely et al. 2004b). Several investigators (for instance, Li et al. 1969; Tsunogai et al. 1973; Chen 1983; Chen 1990 and Feely et al. 2002a) estimated dissolution rates of skeletal material in the Pacific that mostly ranged from 0.03 to 0.06 μmol kg^{-1} y^{-1}. Both organic carbon decomposition and inorganic carbon dissolution resulted in increase in deep water DIC by 0.2 and 0.15 μmol kg^{-1} y^{-1} in the Arabian Sea and Bay of Bengal, respectively. Based on the residence time of waters, between 1500 m and bottom, in the Indian Ocean (250 years; Stuvier et al. 1983), DIC would increase by ~50 ± 5 in the Arabian Sea and by 37 ± 4 μmol kg^{-1} in the Bay of Bengal. The increase in DIC, between σ_3 levels of 41.446 and 41.488, from 0 to 15°N (2341 ± 6 μmol kg^{-1}) to the Arabian Sea (15–30°N; 2381 ± 5 μmol kg^{-1}) and 2331 ± 6 μmol kg^{-1} (0–15°N) to the Bay of Bengal (2359 ± 2 μmol kg^{-1}) amounts to 40 ± 8 and 28 ± 6 μmol kg^{-1}, respectively, which is consistent with that estimated using dissolution and decomposition rates.

Further evidence in support of that in situ decomposition and dissolution of sinking particulate carbon resulted in deep water DIC increase in the Arabian Sea comes from the loss of sinking organic carbon between ~2000 and ~3500 m that ranged from 0.04 to 0.34 mmolC m^{-2} d^{-1} (Table 4). In order to remove the rate term, we multiplied the same with residence time of waters (250 years; Stuvier et al. 1983). This resulted in an increase in DIC between 2000 and 3500 m by ~2–27 μmol kg^{-1} due to decomposition of sinking carbon. Similarly loss of $CaCO_3$ fluxes between the same depths, mentioned above, led to a DIC

Table 3 Rates of oxygen utilization (OUR), organic carbon regeneration (OCRR) and carbonate dissolution (CDR) in the different regions of the World Oceans

Region	OUR	OCRR	Reference	CDR	Reference
N. Pacific	0.13	0.107	Chen (1990)	0.01–0.05	Feely et al. 2002
S. Pacific	0.05–0.08	0.041–0.066	Wyrtki (1962); (Craig 1971)	0.04–0.06	Feely et al. 2002
E.Eq.Pacific	0.03–0.14	0.024–0.112	Kroopnicks (1974)	0.005–0.014	Honjo et al. (1995)
Atlantic	0.09–0.11	0.05–0.073	Arons and Stommel (1967); Wright (1969)	0.3–0.75	Feely et al. (2004a, b)
Arabian Sea	0.12–0.19	0.09–0.154	This study	0.07–0.1	This study
Bay of Bengal	0.1–0.16	0.08–0.13	This study	0.066–0.068	This study
S. I. Ocean	0.1–0.15	0.07–0.10	This study	0.058–0.073	This study

All rates are given in (μmol kg^{-1} y^{-1})

Table 4 Increase in DIC due to loss of sinking carbon based on sediment trap data in the Arabian Sea

Trap position	Loss of POC flux from shallow to deep trap (mmol m^{-2} d^{-1})[a]	Depth difference between deep to shallow trap (m)	Increase in DIC due to decomposition of sinking POC (μmol kg^{-1})	Loss of PIC flux from shallow to deep trap (mmol m^{-2} d^{-1})[b]	Increase in DIC due to dissolution of sinking PIC (μmol kg^{-1})
17.2°N, 58.4°E	0.34	1165	26.63	0.15	12.5
17.1°N, 59.3°E	0.30	1026	26.68	0.06	5.0
15.2°N, 61.3°E	0.19	1262	13.52	0.05	3.8
10.0°N, 65.0°E	0.04	1552	2.35	0.07	4.0

Source: Lee et al. (1998)[a] and Honjo et al. (1999)[b]

increase of 4–12.5 μmol kg^{-1} in the Arabian Sea. Thus these two processes together increase deep water DIC by ~6–39.5 μmol kg^{-1} in the Arabian Sea. This is in good agreement with the increase in DIC found in the Arabian Sea at depths below 2000 m (40 ± 8 μmol kg^{-1}) relative to South Indian Ocean and also based on decomposition and dissolution rates (50 ± 5 μmol kg^{-1}). This strongly suggests that the increase in DIC in the deep waters of the Arabian Sea is driven by higher rates of decomposition of organic matter and dissolution of skeletal material. Therefore, ~60% of deep water DIC increase is due to decomposition of organic matter and 40% because of dissolution in both the Arabian Sea and Bay of Bengal. Li et al. (1969) found that only about 20% of increase in DIC is derived from inorganic carbon dissolution in the Pacific and Atlantic Oceans. Several other investigators found inorganic carbon dissolution in the deep waters of the Pacific to be in the range 12–32% (for example Chen et al. 1982; Chen 1990; Edmond 1974; Tsunogai 1972).

The Arabian Sea is found to be a strong source of CO_2 to the atmosphere with a similar magnitude per unit area compared to that of eastern equatorial Pacific (Goyet et al. 1998; Sarma et al. 1998; Takahashi et al. 2002; Sarma 2003). The annual emission of CO_2 to the atmosphere is estimated to be 90 TgC (Sarma 2003). On the other hand, the net community production (NCP), which is also a measure of export production, is estimated to be 200–240 TgC y^{-1} in the photic zone of the Arabian Sea (Sarma 2004). Unlike in the eastern equatorial Pacific, where export production is almost equal to the CO_2 fluxes to the atmosphere (Feely et al. 2002b; Le Borgne et al. 2002), the Arabian Sea pumps carbon into the deep waters which is nearly double to that being pumped into the atmosphere. Although much of the biologically produced organic carbon remineralizes in the subsurface waters, a significant fraction reaches the deep waters through sinking where it continues to get remineralized and is stored as inorganic carbon for longer times. It would be interesting to examine how this mechanism would be influenced due to changes in circulation pattern or plankton community structure, which ultimately change the sinking flux patterns. Such changes might be forced by climate change (Schulte et al. 1999).

Conclusions

Accumulation of DIC in deep waters is found to be the highest in the North Indian Ocean compared to that elsewhere in the world oceans. The deep waters of the Arabian Sea receive higher amounts of both organic and inorganic carbon through sinking from the euphotic zone. Their subsequent decomposition and dissolution resulted in the accumulation of DIC (6–39.5 μmol kg^{-1}) in deep waters. Although this region acts as a source for atmospheric CO_2 the most efficient biological pump here facilitates the storage of a significant amount of carbon for longer times in in dissolved inorganic form.

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