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Understanding the Arabian Sea: Reflections on the 1994–1996 Arabian Sea Expedition

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Abstract

The Arabian Sea Expedition, now five years past its field observations, is at a stage when some of its dominant themes can be summarized. Of the large range of possible topics, five are considered here: (1) Is the Arabian Sea a source or sink for carbon dioxide?; (2) Is the Arabian Sea Mother Nature's iron experiment?; (3) Do grazing zooplankton control carbon flux to the seabed?; (4) Does the paleoceanographic record help us predict the ocean's response to climate change?; and (5) What are the predominant physical processes of the Arabian Sea? A short history of each issue and results from the field work of 1994–1996 are presented.

1. Introduction

The Arabian Sea Expedition, which developed initially within the framework of the US JGOFS program, began with the goal of understanding the processes controlling the time-varying fluxes of carbon and associated biogenic elements (SCOR, 1990). In a review of the state of our knowledge of the Arabian Sea, numerous scientific questions arose, all pertaining to aspects of the carbon and nitrogen cycles of this region, which experiences strong and seasonally varying forcing from the atmosphere (Smith et al., 1991). The combination of investigations focused on regional biogeochemistry (Arabian Sea Process Study) and regional physical oceanography (Forced Upper Ocean Dynamics Program) has provided unusually comprehensive and interdisciplinary knowledge of this remote and challenging ocean. A few of the questions about which we have gained substantial new insight over the past several years are summarized.

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2. Results and discussion

2.1. The carbon dioxide cycle: is the Arabian Sea a carbon source or sink?

The sea-surface carbon dioxide surveys conducted during most of the US JGOFS cruises in the Arabian Sea showed prominent seasonal and spatial variability. Seasonality was most pronounced in waters within 300 km of the coast where pCO_2 during upwelling was more than 260 μatm higher than in the same location during the non-upwelling seasons (Fig. 1; Goyet et al., 1998). Offshore, seasonal variation was less than 4 μatm. Spatial gradients were also steepest during the summer upwelling season (Southwest Monsoon). More widespread elevation of sea surface carbon dioxide concentrations were recorded during the Northeast (NE) Monsoon (winter), reflecting the broad-scale influence of cool, dry atmospheric conditions that lead to convection (Millero et al., 1998). Thus, upwelling during the Southwest (SW) Monsoon caused the most intense response and the largest concentration differentials at the sea surface, but mixing caused by convection during the NE Monsoon season caused the largest areal elevation of surface pCO_2.

Upwelling areas, in general, have potential to be both sources and sinks of carbon dioxide. They are sources on the short term when cold upwelled water warms and loses its carbon dioxide to the atmosphere and sinks on the longer term when carbon fixed biologically in the upwelling area is deposited at depth (Watson, 1994). The combination of outgassing and biological uptake can create an undersaturation in the surface layer, which is not restored for a matter of weeks or months (Lampitt et al., 1994). The Arabian Sea is an excellent example of the generalizations from the Watson (1994) model. Looking only at the concentrations of carbon dioxide in the surface

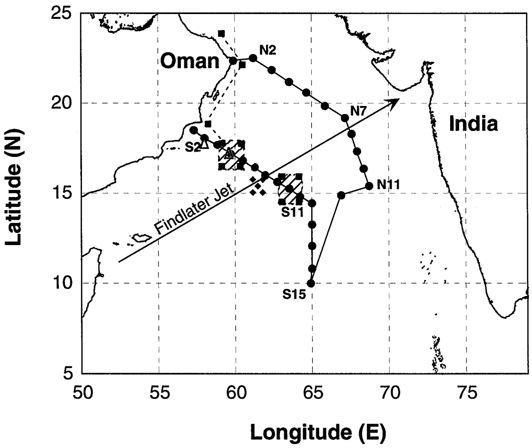

Fig. 1. Map of the study area for the Arabian Sea Expedition. The US JGOFS stations and cruise track are shown as solid circles and solid lines. Stations N1 through N11 comprise the north line and stations S1 through S15 comprise the south line. The Forced Upper Ocean Dynamics cruise track is shown in solid squares and dashed lines; it overlaps the US JGOFS track between stations S2 and S11. The hatched boxes are areas intensively surveyed by SeaSoar. Solid diamonds show the locations of moorings measuring currents, optical properties, temperature, salinity and oxygen concentration. The open triangles show the locations of sediment trap moorings MS1 and MS3. The grey shaded area depicts the general location of a major filament carrying upwelled water offshore.

layer, and the atmosphere adjacent the ocean's surface, Goyet et al. (1998) conclude that the Arabian Sea is a small (460 mmol C/m^2/yr) source of CO_2 to the atmosphere. However, if one considers the carbon captured by sediment traps under the upwelling area (Honjo et al., 1999), the conclusion is that the Arabian Sea is a sink (822 mmol C/m^2/yr). The areal extent of the Arabian Sea acting as a sink is approximately twice the size of the Arabian Sea acting as a source.

The efflux of carbon dioxide at the sea surface in July was roughly 313 mmol C/m^2/month (Goyet et al., 1998), largely because of upwelling, but the uptake of carbon in photosynthesis at that time was approximately 3750 mmol C/m^2/month (Barber et al., 2000). Adjusting this average by the average *f*-ratio measured in the upwelling area the previous month (Stations S1–S3; 0.15; Sambrotto, 2001) suggests that a minimum of 562 mmol C/m^2/month were exported to depth. The biological processes were removing two to ten times more carbon from the surface layer (0–30 m) than the physical processes during the SW Monsoon season. The culmination of the biological carbon transformations is reflected in the carbon content of the sediments of the margin of the Arabian Sea. The swath of carbon-rich sediments off Oman and Yemen is as broad as that off Peru and Central America, or any place else on the planet (Premuzic, 1979). The measured organic carbon content of the sediments, as much as 7.5% by weight (Shimmield et al., 1990), is as high as the organic carbon content of any other sediments worldwide (Premuzic, 1979).

Another possible sink is the dissolved organic carbon pool, measured in the US JGOFS program as total organic carbon. The upper layer (150 m) showed a seasonal variation of roughly 2.5 mol C/m^2 between the NE and SW Monsoons, with the early SW Monsoon having the highest concentrations (11.5 mol C/m^2; Hansell and Peltzer, 1998). The lowest concentrations were measured in the Fall Intermonsoon (FI) season over the entire study area (Hansell and Peltzer, 1998). A nearshore/offshore contrast of roughly 1 mol C/m^2 existed in the late NE Monsoon, Spring Intermonsoon (SI) and early SW Monsoon. There was no variability in deep-water concentrations. In contrast to the sediment trap collections (Honjo et al., 1999), TOC was higher offshore than nearshore when a difference existed (Hansell and Peltzer, 1998). The large size of the TOC pool makes it a major component of the carbon budget, but its heterogeneous nature and lack of clear definition of a refractory component hamper rigorous evaluation of its role as a sink.

In calculations of global carbon budgets, the Arabian Sea, due to its small area, is surpassed by the North Atlantic Ocean as a sink for carbon dioxide and by the Equatorial Pacific Ocean as a source of carbon dioxide (Goyet et al., 1998). For modeling and prediction purposes, it is essential that both the short-term carbon dioxide changes (air–sea exchange physically driven) and long-term carbon transformations (deposition at the seabed; storage as refractory dissolved organic carbon in the deep water) be evaluated and quantified.

2.2. *Primary productivity in the Arabian Sea: Mother Nature's iron experiment?*

The Indian Ocean receives up to 63% of terrestrial dust deposited in the oceans world-wide each year, and the frequency of observation of haze in the marine atmosphere is higher in the Arabian Sea during June–August than it is off northwestern Africa in December–February (Pye, 1987). More recent investigations, particularly the intensive observations of the Monsoon Experiment (MONEX) of the First Global Garp Experiment (FGGE) in the Arabian Sea region in 1979, have shown that most of the dust is delivered in June, July and August and that the source of dust is the Arabian Peninsula and Iraq (Ackerman and Cox, 1989; Sirocko, 1993). In spite of the lower tropospheric jet (Findlater Jet), which blows strongly during summer from Somalia across to India, dust is delivered to the Arabian Sea by shamal winds blowing down the Arabian peninsula and south of the mountains of Oman (Sirocko, 1993; Sirocko and Sarnthein, 1989). A recent attempt to quantify the global nature of aerosols over the ocean using the advanced very high resolution radiometer (AVHRR) showed a seasonality in the Arabian Sea consistent with earlier reports (Husar et al., 1997). The estimated aerosol optical thickness over the Arabian Sea in summer is the highest value anywhere on the planet, and the horizontal extent of this thick layer equals or exceeds that seen off west Africa (Husar et al., 1997).

The direction dust is being carried off the Arabian peninsula depends upon height above the earth. In the first kilometer, the layer is moving from Arabia to Pakistan, while in the 1–2 km layer, dust is moving into the Arabian Sea from the peninsula (Sirocko and Sarnthein, 1989). In the 2–4 km layer, dust is traveling from the Arabian peninsula into the open Arabian Sea and to Somalia. In spite of these complex trajectories, dust accumulates in the sediments of the Arabian Sea inshore of the Findlater Jet or north of approximately 20°N (Sirocko and Sarnthein, 1989).

How much of this dust aloft reaches the sea surface? Because of the potential delivery of iron-rich dust to the sea surface during the upwelling season (SW Monsoon), the Arabian Sea can be considered Mother Nature's iron experiment. Measurements of aerosols in the oceanic boundary layer measured on the ship at a height of 6 m above the bow of the R.V. *Thomas G. Thompson* (approximately 13 m above the sea surface) during the US JGOFS investigations revealed that suspended dust levels were highest in the NE Monsoon and Spring Intermonsoon (Pease et al., 1998; Tindale and Pease, 1999), the opposite seasonality of earlier reports (Pye, 1987). In the SW Monsoon of 1995, dust was delivered near the coast by shamals carrying material from the Wahiba Sands of coastal Oman (Pease et al., 1998); concentration of dust decreased with distance from the coast. Measurement of dust in the air near the sea surface in the SW Monsoon was complicated by rain and sea salt, both of which could strip dust out of that layer and convert it to wet deposition. Aerosol sea salt concentration was high during the SW Monsoon, suggesting a mechanism by which suspended dust levels in the atmosphere were reduced (Tindale and Pease, 1999). Measurements of aluminum and iron in surface seawater during the SW Monsoon showed that dust does get deposited; concentrations of aluminum doubled in the SW Monsoon compared with the NE Monsoon and Spring Intermonsoon, while iron increased by 30% in the SW Monsoon (Measures and Vink, 1999). Measures and Vink (1999) further demonstrated that although iron concentrations were high in the suboxic layer, this layer does not upwell and that the freshly upwelled oxic water from above this layer contained sufficient iron to allow full biological utilisation of the available nitrate. Thus, continued dust input is necessary in the upwelling area to maintain high rates of primary productivity.

The coastal zone color scanner (CZCS) archival composite images of sea-surface color, and the analyses of these by Banse and English (1994), suggested that a strong seasonal contrast in surface pigment concentrations, and therefore in primary productivity, would be observed in the Arabian Sea. The archival images also indicated enhanced primary productivity in a broad area off Oman during the SW Monsoon compared to a more restricted area during the NE Monsoon. Data from the International Indian Ocean Expedition in 1964–1966 showed a four-fold difference in primary productivity between the SW and NE Monsoons (Table 1; 15–20°N, 57–65°E) in the area within 900 km of the coast of Oman. New images of monthly surface ocean pigment concentration observed by the Sea-viewing Wide Field of view Sensor (SeaWiFS) show seasonal patterns that are

Table 1
Average daily primary productivity (mgC/m^2/day) for two seasons, SW Monsoon (May–October) and NE Monsoon (November–April)

Location	1964–1966[a]		1995[b]	
	NEM	SWM	NEM	SWM
15–20°N; 57–65°E	220	810	1210	1470
10–15°N; 65°E	310	130	940	1050
15–23°N; 60–65°E	800	620	1650	1080
15°N; 65–70°E	100	70	1220	1110

[a]Data from Krey and Babenerd (1976).
[b]Data of R. Barber and J. Marra, analyzed by A. Hilting.

not very different from the CZCS archive (Murtugudde et al., 1999). However, the measured rates of primary productivity for the region in 1995 were very different from those of the 1960s (Barber et al., 2001). In the area of upwelling off Oman, primary production during the SW Monsoon was nearly twice that measured in the 1960s, while during the NE Monsoon, primary production in 1995 was six times that of the 1960s (Table 1; 15–20°N, 57–65°E). In the less productive, more offshore, portion of the study area, primary production in the SW Monsoon was eight times that measured in the 1960s, and during the NE Monsoon, the differential was three times (Table 1; 10–15°N, 65°E). Along the north line nearer India (Fig. 1), primary production was two to sixteen times greater in 1995 than it was in the 1960s (Table 1; 15–23°N, 60–65°E and 15°N, 65–70°E).

In trying to ascertain why primary production apparently has increased so much in the last three decades, especially why the NE Monsoon supports such high primary production, there are at least three possibilities. First, the methods for handling seawater samples used to measure primary production have changed dramatically. In the 1960s possible contamination from the ship was not recognized, and the clean techniques used uniformly today were unknown. Second, the sampling tracks were not arranged to capture spatial variability and the times of cruises were not tied to the seasonality of the monsoon, resulting in considerable alias in the spatial and temporal averages. Finally, the autotrophic community structure may have changed. Certainly the cyanobacteria, which are prominent in the primary productivity story of the Arabian Sea over the past decade (Burkill et al., 1993; Veldhuis et al., 1997; Campbell et al., 1998; Tarran et al., 1999), were unknown in the 1960s. It may be possible to track species shifts in larger taxa more carefully because floristic studies were conducted in the 1960s and 1990s, but this comparison has not yet been done. The dominant finding was that the NE Monsoon was much more productive than had been known previously (Madhupratap and Parulekar, 1993; Madhupratap et al., 1996; Barber et al., 2001). The strong seasonal contrast seen in CZCS and SeaWiFS images, but not evident in the actual measurements of primary production in the 1990s, may be the result of aerosols being registered by CZCS as plant pigment during the SW Monsoon season.

The combination of iron delivered by airborne dust and upwelling of nitrate at the same time make the Arabian Sea "Mother Nature's" iron experiment. A comparison of primary production

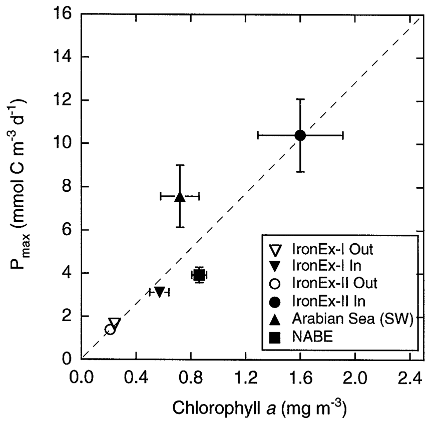

Fig. 2. The correlation of maximum primary productivity and chlorophyll *a* biomass measured in a variety of environments. The error bars for the IronEx I and II Out are obliterated by the symbols. These data provided by R. Barber.

achieved for a given concentration of chlorophyll *a*, when nitrogen is not limiting, among several study regions including areas of deliberate iron fertilization demonstrates that the Arabian Sea is "Mother Nature's" iron experiment (Fig. 2). Primary production per unit chlorophyll *a* is considerably higher (approximately double) in the Arabian Sea than in either of the controlled, iron-addition experiments conducted off the Galapagos Islands.

2.3. *Grazing Zooplankton in the Arabian Sea: do they control major vertical flux events?*

The patterns of flux and composition of material arriving in sediment traps moored in the Arabian Sea have been described comprehensively in an observational program that began in 1986 (Nair et al., 1989). In the first four years of observation, a peak in carbonate flux (coccolithophores and foraminifera, especially *Globigerina bulloides*) into traps was seen in early August, followed by a peak in biogenic opal flux (diatoms, particularly the genus *Rhizosolenia*) in early September (Haake et al., 1993). These peaks in flux occurred many weeks after the onset of upwelling-favorable winds in late May or early June. One of the important conclusions of this work has been that interannual variation in total flux in the region of upwelling off Oman is due to variation in flux during the SW Monsoon. The processes and sequence of events that occur during the SW Monsoon determine the geologic record of the northwestern Arabian Sea. The large flux of biogenic opal late in the upwelling season, observed repeatedly, is a characteristic that needs explanation.

Within the area off Oman where upwelling occurs and its effects are observed (Stations S1–S7; Fig. 1) during the SW Monsoon season, primary productivity showed little variation from one season to the next (approximately 1.0–1.5 gC/m^2/day or 83–125 mmol C/m^2/day, Barber et al., 2001). At the close of the SW Monsoon, however, particulate flux into the sediment traps beneath the area of upwelling was substantial: approximately 45–55 mg organic C/m^2/day (4–5 mmol C/m^2/day), 30–35 mg inorganic C/m^2/day, and total flux of 650–700 mg/m^2/day (Fig. 1; moorings MS1 and MS3, Honjo et al., 1999). The events that brought the largest particulate flux into the

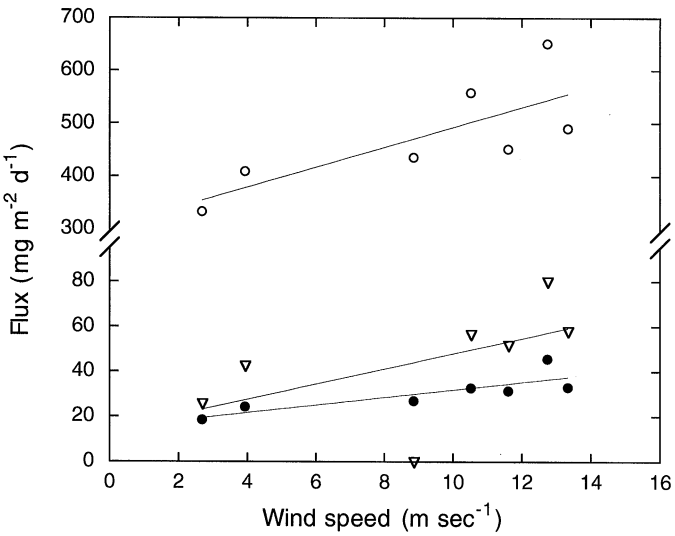

Fig. 3. Correlations of total flux (open circles), organic carbon flux (solid circles) and biogenic silica flux (open triangles) with wind speed at sediment trap mooring MS3, depth 1882 meters. There is a 68 day lag between the measurements of wind speed and trap contents. Note the break in the ordinate scale. Data are from Honjo et al. (1999) and Weller et al. (1998).

traps also brought the largest biogenic silica flux, suggesting that diatoms were a major component of the dominant flux events. Outside the area of upwelling, fluxes were uniformly low (4 mg organic C/m^2/day and 5 mg inorganic C/m^2/day; Honjo et al., 1999). The acceleration of flux of organic carbon and biogenic silica measured in the upwelling area off Oman between July 24 and September 30, 1995, was correlated with acceleration in wind speed measured at a nearby mooring (15°30′N, 61°30′E) between May 17 and July 24, 1995, but with a lag of $\sim$ 70 days (Fig. 3). This correlation shows that there is a substantial delay between the onset of upwelling-favorable winds and the peak flux of biogenic material to the seabed. Why is this so?

Although some flux events under the Findlater Jet seem to correlate with primary productivity in the upper layer (Honjo and Weller, 1997), the largest flux events, observed in the upwelling area near the end of the SW Monsoon, may be the result of the cessation of feeding by large-bodied zooplankton that selectively graze diatoms and dinoflagellates in the upwelling area. The succession in the phytoplankton community during the upwelling season was captured by the pigments measured as HPLC (Latasa and Bidigare, 1998) and in cell counts (Wood, unpublished data). Early in the upwelling season (SW Monsoon; June and July), the net phytoplankton community in the upwelling area was dominated by diatoms, particularly *Proboscia alata*, *Thalassionema nitzschoides* and *Chaetoceros* spp. (Wood, pers. comm.). In the middle of the upwelling season (SW Monsoon; July and August), of the total of the pigments fucoxanthin, peridinin, 19′-butanoyloxyfucozanthin, and 19′-hexanoyloxyfucoxanthin, fucoxanthin (diagnostic for diatoms) was an average of 49% in the upper 30 m at station S2 (Fig. 1) nearshore and 21% at station S6 (Fig. 1) just inshore of the Findlater Jet (overall average for stations S2–S6 was 30%). By late in the SW Monsoon (August–September), fucoxanthin at S2 was 51% of selected pigments in the upper 30 m, while at S6 it was 58% (overall average for stations S2–S6 was 50%). These pigment data suggest that from the middle to the late SW Monsoon, the area in which diatoms were predominant expanded from between the coast and 150 km offshore to between the coast and 550 km offshore. Biomass of autotrophic plankton in the mixed layer of the upwelling area late in the SW Monsoon was 55% diatoms, dominated by *Rhizosolenia* species (Garrison et al., 1998), a result also reported by Tarran et al. (1999). Although standing stocks of picophytoplankton are high in the SW Monsoon (Tarran et al., 1999; Campbell et al., 1998; Shalapyonok et al., 2001; Garrison et al., 2000), they are not included here because their direct contribution to particulate flux in the area of upwelling is small.

During August/September, grazing by *Calanoides carinatus* and *Eucalanus subtenuis*, two copepods that dominate the biomass of the upper 200 ms in the upwelling area, was predominantly on diatoms (55–56% of cells ingested by *C. carinatus*; 100% of cells ingested by *E. subtenuis*). The proportion of diatoms ingested was greater than the proportion of diatoms in the assemblage offered, suggesting that these two taxa were ingesting diatoms selectively in the upwelling area during the SW Monsoon. Furthermore, it is known that *C. carinatus* feeds particularly on diatoms in other upwelling areas (Schnack, 1982) and undergoes ontogenetic migration in which a late subadult population (copepodid stage V, CV) lives in a dormant state in deep layers during the non-upwelling season (Mensah, 1974; Binet and Suisse de St. Claire, 1975). In the northwestern Indian Ocean, *C. carinatus* CV is found at depths greater than 400 ms from December to May, and in the upper 100 ms abundantly in early June at the onset of upwelling (Fig. 4). Thus, these

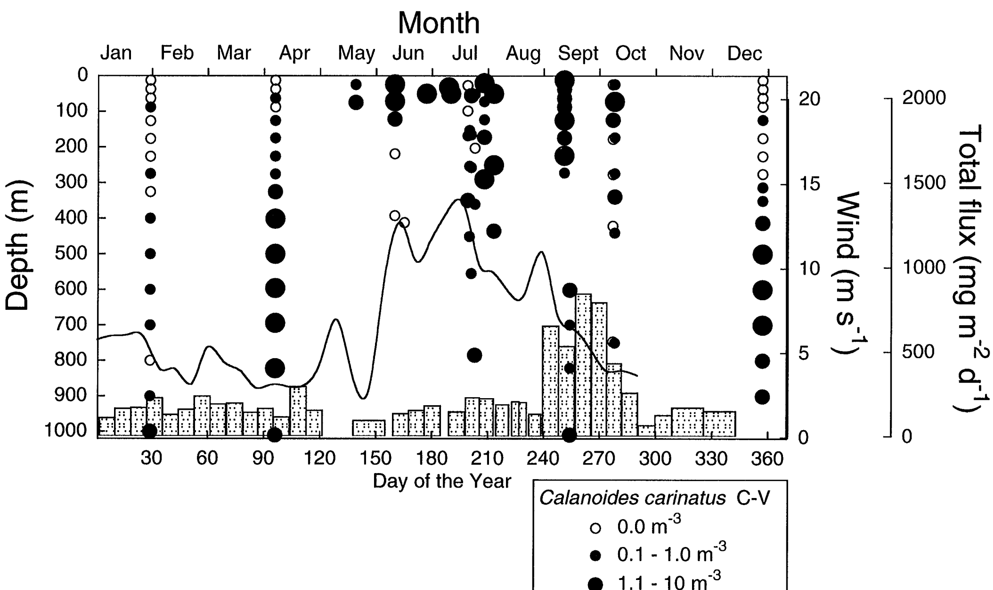

Fig. 4. Composite showing the annual cycle of the vertical distribution of the large-bodied grazer *Calanoides carinatus* collected off Somalia and Oman, the onset of upwelling-favorable winds at the central mooring (15°30′N, 61°30′E), and total mass flux into a sediment trap in the upwelling area off Oman (sediment trap MS3; 858 meters). Grazers are at the surface at the beginning of upwelling in late May and early June. Substantial flux into the sediment trap occurred only after the grazer began to leave the surface in early August. Somali data provided by M. Baars; June data for Oman derived from samples collected by M. Wood.

large-bodied, diatom grazers were in the euphotic zone at the start of the upwelling season and potentially able to prevent an early diatom bloom by exerting predatory control on the diatoms. Later in the upwelling season (early August), the population of *C. carinatus* CVs begin to leave the euphotic zone when they have acquired the lipid reserves necessary for successful diapause at depth (Fig. 4). Grazing pressure on diatoms was thereby reduced, and biogenic silica flux increased in the traps in the upwelling area (Honjo et al., 1999). The largest biogenic silica fluxes occurred in September, following the collapse of upwelling-favorable winds on September 9, 1995 (Weller et al., 1998) and rapid heating of the surface layer (Smith et al., 1998a). *Calanoides carinatus* is never found at temperatures above 27°C (Smith et al., 1998b); the rapid heating of the sea surface in September drove the remaining euphotic zone population to deeper layers in a matter of a few days following collapse of the winds. Therefore, grazing pressure on diatoms was removed quickly over a large area; the diatoms used remaining nutrients and became senescent; senescent diatom cells aggregated and sank rapidly. This scenario is supported by the observation that lithogenic fluxes are high late in the SW Monsoon because the lithogenic fraction is transported with biogenic particles (Ramaswamy, 1993) and that in conditions of high grazing, traps contain diatoms and diatom frustules but no fecal pellets (Passow et al., 1993). Thus, it may be the release of grazing pressure (predation pressure) on diatoms, driven by collapse of the winds and increasing sea surface temperatures, that triggers the largest fluxes of carbon and silica to the seabed of the Arabian Sea in September each year.

2.4. The Paleoclimate Record in the Arabian Sea: any clues that help us predict the ocean's response to climate change?

Eight million years ago the Tibetan plateau began increasing in elevation, and within two million years it had risen an additional 1000–2500 m (Molnar et al., 1993). The massive size and great height of the Tibetan plateau interrupted the Hadley circulation and caused the climate of the region to change; monsoon winds strengthened and nearby land became arid (Molnar et al., 1993). This extensive, high landmass adjacent to a large ocean created the monsoon engine and upwelling began (Prell et al., 1992; Meyers and Dickens, 1992). The paleoclimate record shows that in the Arabian Sea, temperature decreased, nutrients increased, and the abundance of endemic upwelling species increased at this time (Prell et al., 1992) and biosiliceous sediments began to accumulate (Baldauf et al., 1992). The fundamental process of the region, from which much of the atmospheric forcing and circulation of the ocean derive, is the differential in heating between the Tibetan plateau/Indian subcontinent and the large volume of ocean surrounding it (Prell, 1984; Webster, 1987). The low-pressure area that develops over the Tibetan plateau during heating in spring and the high-pressure system over the ocean cause lower level tropospheric winds to flow northward from the equatorial zone toward the Tibetan plateau. The Coriolis effect causes these winds to be diverted first toward Africa near the equator and then toward Tibet at latitudes of approximately 10–20°N (Webster, 1987). The deflection of wind direction is further enhanced by the mountainous regions of Kenya, Ethiopia and Yemen (Knox, 1987; Krishnamurti, 1987). Because the mountains and Coriolis effect are unchanging, the timing of the onset of the lower tropospheric jet (Findlater Jet) depends on the timing of heating of the Tibetan plateau. Since this is a function of the annual circuit of earth around the sun, onset of the jet is a remarkably regular event. From 1901 to 1978, onset of the SW Monsoon over Kerala, India, varied only between May 31 and June 18 (Shukla, 1987). Records of onset of winds recorded by merchant vessels show that winds associated with the Findlater Jet begin within a few days of May 24th in 60% of the years examined (Fieux and Stommel, 1977). For a regional climate response, this is a very small temporal window.

We chose the Arabian Sea because the intensity and predictability of physical forcing made it an excellent place to look at possible climate change responses. Although the heating of the Tibetan plateau is a regular and predictable event, monsoon intensity is governed by the amount of snow and ice on the plateau. When snow and ice are deep on the plateau, sensible heat melts the snow and ice but does not heat the land. The result is a weak SW Monsoon because the contrast between the low-pressure system over Tibet and the high-pressure system over the Indian Ocean is small. In years of little snow and ice, sensible heat in spring and summer warms the land, and the contrast between land and ocean is large resulting in a strong SW Monsoon. Consequently, anomalies in rainfall over India and snow cover over the Tibetan plateau are inversely related (Shukla, 1987). In years of high snowfall anomaly in winter, the subsequent SW Monsoon is weak and picks up less moisture as it transits over the Arabian Sea, resulting in reduced rainfall over India. In the ten-year period, 1967–1977, the largest negative rainfall anomaly and the largest positive snowfall anomaly occurred in 1972 (Shukla, 1987), the year of the strongest El Niño in that ten-year period as well (Harrison and Larkin, 1998).

The paleoclimate record assembled from cores in the region and various modeling studies suggests that we may be able to predict some regional consequences of increased carbon dioxide in the atmosphere and climate warming. The region has a distinct foraminiferan assemblage (*Globigerina bulloides; Neogloboquardrina pachyderma*) associated with the upwelling response to the SW Monsoon (Prell and Curry, 1981; Ivanova, 1999). During the last interglacial (9000 years B.P.), solar radiation was 7% stronger during the SW Monsoon than it is today (Kutzbach, 1981), and upwelling off Oman influenced an area more the twice the size of the upwelling area today (Fig. 5; Prell and Streeter, 1982; Prell et al., 1990). Proportionally, the foraminifera indicative of upwelling (*Globigerina bulloides*) that reached the seabed during the last interglacial was twice that reaching the seabed today (Prell, 1984). The cores also contain more pollen of tropical montane plants (Prell and Van Campo, 1986) and larger grain size in the lithogenic fraction (Prell et al., 1992), both suggesting stronger winds and higher rainfall in the SW Monsoons of the last interglacial period. Modeling studies have indicated that increased solar radiation in the SW Monsoon yields a larger land-sea pressure gradient, resulting in stronger winds, more upwelling, and more precipitation (Prell and Kutzbach, 1987; Kutzbach, 1981; Prell et al., 1990). All of these model results are consistent with the distribution of paleoclimate indicators quantified from cores. In terms of factors influencing the strength of the SW Monsoon, models and the paleoclimate record agree that spring/summer solar radiation is dominant, followed closely, but on a different time scale, by changes in the height of the Tibetan plateau (Prell et al., 1992). The extent of the glacial ice sheets had less influence.

The combination of model results, paleoclimate records, and correlations among present day monsoon variables (i.e., rainfall, snow cover), which all fit together into a consistent paradigm, suggest that we can predict the outcome of climate warming in the region. Because warming would reduce the snow and ice cover on the Tibetan plateau, the pressure gradient between the plateau and the surrounding ocean would be increased, the SW Monsoon winds would increase, and upwelling would become both more vigorous and more widespread in the region. Increased upwelling is also associated with increased deposition at the seabed, a projection based upon

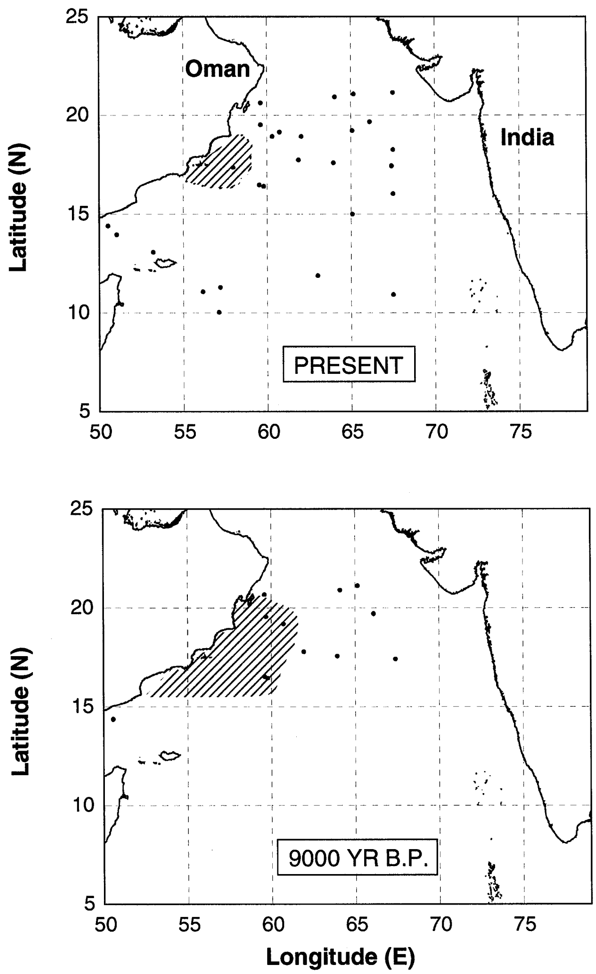

Fig. 5. Location and extent of upwelling areas (hatched) deduced from the presence of the foraminiferan *Globigerina bulloides* in sediment cores. Data in the lower panel are from the last interglacial period when a lack of snow and ice on the Tibetan plateau caused winds to be stronger. Redrawn from Prell et al. (1990).

observations from the last interglacial period ($\sim$ 9000 years B.P.). During the last glacial period, 18,000 years B.P., the SW Monsoon was weak (Kutzbach, 1981; Prell and Van Campo, 1986) or absent (Duplessy, 1982), and radiolaria and diatoms were absent (Prell et al., 1980). The large and quantifiable contrasts that characterize the Arabian Sea's response to the most recent major fluctuations (glacial/interglacial) in climate suggest that prediction of the ocean's response to climate change in this region is possible with a high degree of realism.

2.5. Physical Forcing in the Arabian Sea: open-ocean upwelling or another process?

Prior to 1995, our understanding of circulation in the northern Arabian Sea was based on very few field observations. The cruises of the R/V *Discovery* during the International Indian Ocean

Expedition (1963) revealed offshore extensions of cool water connected to the coastal upwelling zone (Bruce, 1974) and a subsurface, southward flowing current (countercurrent) near the coast during the SW Monsoon (Swallow in Smith et al., 1991). Analysis of the 1963 data, which showed upward vertical velocities 1000 km alongshore and 400 km offshore of Oman, led to the conclusion that both coastal upwelling and open-ocean upwelling due to strong wind stress curl accounted for this large upwelling area (Smith and Bottero, 1977). The concept of open-ocean upwelling, and its biological manifestations, was developed further by Bauer et al. (1991) and Brock et al. (1991). Hydrographic and acoustic Doppler current profiler data collected during the SW Monsoon of 1987 revealed an offshore-flowing filament connected to the area of upwelling near Ras Madraka, coastal upwelling from Ras Marbat to Ras al Hadd, and a strong front at Ras al Hadd (Elliot and Savidge, 1990). Finally, repeated XBT sections taken over many years off commercial vessels in the region showed numerous, persistent eddies from the Great Whirl off Somalia to the Socotra Eddy to a smaller eddy near Masirah Island (Bruce, 1979). Presciently, in 1991 Swallow stated, "… the distribution of upwelled water, after it leaves the immediate neighborhood of the upwelling region, is going to be strongly influenced by a succession of eddies. Those eddies identified by XBT tracks have vertical structure often extending to depths of 150–200 m … " (Smith et al., 1991).

The study of the annual cycle of physical forcing in 1994–1996 supplied evidence that cast doubt on the relative importance of open-ocean upwelling off the coast of Oman, showed the importance of eddies in all seasons, confirmed the dominant role of coastal upwelling and advection during the SW Monsoon season, and demonstrated the importance of convection during the NE Monsoon season.

2.5.1. *Open-ocean upwelling*

Because lateral variation in the exceptionally strong wind stress over the Arabian Sea during the SW Monsoon could cause vertical movement of water in an area where the CZCS recorded elevated plant pigment during the SWM, a conceptual model was developed that included open-ocean upwelling to explain the widespread distribution of high plant pigments extending from the coast of Oman to at least 300 km offshore (Brock et al., 1991; Bauer et al., 1991). Since this mechanism is fundamental to understanding the biogeochemical response of the region, it was investigated carefully in 1994–1996. Earlier studies of Ekman pumping, which used only wind stress curl to specify the vertical velocity, neglected the effect of the gradient of the Coriolis parameter, β, on Ekman pumping, a significant factor at these low latitudes (Lee et al., 2000; Fischer, 1997). The additional term that results from the 'beta'-effect is important in low latitude regions, especially during periods of strong zonal winds (Fischer, 1997; Lee et al., 2000), and can be large enough to alter the sign of the resulting vertical motions. During the SW Monsoon, this term acts to lift the pycnocline on both sides of the Findlater Jet, reinforcing upwelling on the inshore side and greatly reducing downwelling on the offshore side (Lee et al., 2000). However, Ekman pumping offshore of the coastal zone does not lift water to the sea surface; entrainment is an additional mechanism required to move nutrients into the euphotic zone. Calculations comparing the effects of Ekman pumping and wind-driven mixing (entrainment) in our study area, show entrainment is generally far more important than Ekman pumping in creating both the nutrient-enriched areas inshore of the Findlater Jet and the deep mixed layers offshore of the Findlater Jet during the SW Monsoon season (Lee et al., 2000).

A third mechanism, horizontal advection of coastally upwelled waters, shoals and freshens the mixed layer between the Findlater Jet and the coastal upwelling zone towards the end of the SW Monsoon season. This mechanism is necessary to explain observed changes in stratification and water-mass properties (Lee et al., 2000). Ekman pumping cannot produce this shoaling, because it is too weak to counter wind-driven mixed-layer deepening. Ekman pumping brings the pycnocline closer to the surface during the SW Monsoon, but wind mixing has to reach the pycnocline and redistribute nutrients into the upper layer before a plant response is possible. The depth of the mixed-layer showed little diel variability during the SW Monsoon because daytime heating was obliterated by wind mixing (Gardner et al., 1999). The upward mixing of nutrients introduced near the base of the mixed layer by Ekman pumping provides far less nutrient to the regional euphotic zone per unit time than does coastal upwelling and subsequent offshore transport. Therefore, the pigment response seen in the CZCS images in the region 100–600 km offshore is more likely the result of the offshore advection of nutrient-rich-waters upwelled at the coast than of Ekman pumping and subsequent entrainment. A recent modeling study reached similar conclusions (McCreary et al., 1996).

2.5.2. *The eddy field*

The strong, directional winds of the SW Monsoon have captured our attention for millenia, and their effects on ship drift have contributed to our ideas that strong, directional currents characterize both the major monsoon seasons (SW Monsoon and NE Monsoon). However, sixteen months (1994–1995) of currents (20–400 m) measured by a ship-mounted acoustic Doppler current profiler (ADCP) found large temporal and spatial variability associated with eddies throughout the region from Ras al Hadd to Ras ash Sharbatat and from the coast of Oman to 1000 km offshore, regardless of season (Flagg and Kim, 1998). Seasonal mean currents were relatively weak and exhibited little large scale coherence or relation with wind speed or direction, except near shore and over the continental shelf. Inshore of the atmospheric Findlater Jet (0–600 km offshore) on the southern sampling line, annual mean velocities at all depths were dominated by a large ($\sim$ 300 km diameter), anticyclonic eddy particularly evident in the SW Monsoon and associated with filaments advecting offshore. Analysis of TOPEX/Poseidon sea level data also showed the dominance of eddy-like structures in our study area. The satellite data also showed that water upwelled along the coast during the SW Monsoon of 1995 extended offshore in a large filament or plume (Dickey et al., 1998). TOPEX/Poseidon data for other years show similar structures during each SW Monsoon season, but their position with respect to the coast of Oman varies (Manghnani et al., 1998). ADCP data also showed that more than 90% of the total kinetic energy in the region was due to the eddy field, apparently arising from instabilities associated with coastal squirts and jets, and not from the strong directional wind forcing of the region (Flagg and Kim, 1998). The intense eddy field may influence the paths of filaments as they flow offshore, while vertical motions associated with the eddies provides an additional mechanism for bringing nutrients to the surface layer. There is also substantial spatial variation in eddy kinetic energy with larger values in the west and south and a relative minimum over the central and eastern portion of the northern Arabian Sea (Kim et al., 2001).

2.5.3. *The Ras al Hadd Jet*

The only large-scale current response to wind forcing in the northern Arabian Sea is the Ras al Hadd Jet (Kim et al., 2001), which forms during the SW Monsoon and Fall Intermonsoon seasons

(Böhm et al., 1999). The Jet was observed in sea-surface temperature images (advanced very high-resolution radiometer, AVHRR), TOPEX/Poseidon sea level anomalies, and acoustic Doppler surface currents. The jet flows along the coast of Oman between Masirah Island and Ras al Hadd, where it turns offshore into the interior of the northeastern Arabian Sea; its seasonal maximum speed is 1 m/s (Böhm et al., 1999). During the SW Monsoon, the Jet advected cool, upwelled water into the northeastern Arabian Sea and exhibited a strong thermal front on its northern side separating waters originating in the Gulf of Oman from those arising from coastal upwelling. The Jet's thickness (150–400 m) suggests that its transport is significant, particularly because it carries biogenic and biogeochemical properties of the upwelling area into the interior of the Arabian Sea.

2.5.4. Convection during the NE Monsoon season

Mixing and nutrient infusion into the upper layer by convection during the NE Monsoon when cool, dry winds blow across our study area from the Indian subcontinent to Africa have been discussed for some time (Banse, 1984, 1987, 1994; Kumar and Prasad, 1996). The central mooring (15°30′N, 61°30′E) of the Arabian Sea Expedition showed that the heat budget in the NE Monsoon was one-dimensional (Fischer, 1997). The biogeochemical responses, however, generally have been thought to be localized (Banse, 1984, 1996) and short-lived (McCreary et al., 1996), with the exception of the northeastern Arabian Sea (Madhupratap et al., 1996). The combination of moorings and a cruise track covering a large area in 1994–1996 revealed that the mixed-layer deepening was evident throughout the study area in the NE Monsoon (Gardner et al., 1999) and existed from December to April (Dickey et al., 1998). The upper layer throughout the study area north of 15°N was enriched in nitrogen (Morrison et al., 1998), and primary productivity was an order of magnitude higher than had been reported from the International Indian Ocean Expedition in the 1960 s (Barber et al., 2001). The convection of the NE Monsoon has proved to be a broader scale phenomenon with greater longevity than was previously understood; the NE Monsoon has considerable impact on annual regional productivity.

3. Conclusion

The scientific questions for which the Arabian Sea provides provocative new insights are many. The five chosen for this analysis, (1) Is the Arabian Sea a source or sink for carbon dioxide?; (2) Is the Arabian Sea Mother Nature's iron experiment?; (3) Do grazing zooplankton control carbon flux to the seabed?; (4) Does the paleoceanographic record help us predict the ocean's response to climate change?; and (5) What are the predominant physical processes of the Arabian Sea?, are topics that have been debated and discussed often enough that new insights emerged. They are topics ready for summary at this time, and topics of interest to the author. This summary is not comprehensive; it does not review all the literature that has been published over that past decade by investigators from the Netherlands, Germany, the United Kingdom, India and Pakistan, who have put great effort into field programs and subsequent publications. The literature pertaining to the Arabian Sea has grown enormously in the past decade. Omissions also include considerable new knowledge we have gained, for example, on the nitrogen cycle, the microbial loop, and the oxygen minimum zone of the Arabian Sea. These topics are being reviewed by others; new topics will emerge as analysis of Arabian Sea data continues.

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References

Ackerman, S., Cox, S., 1989. Surface weather observations of atmospheric dust over the southeast summer monsoon region. Meteorology and Atmospheric Physics 41, 19–34.

Baldauf, J.G., Barron, J.A., Ehrmann, W.U., Hempel, P., Murray, D., 1992. Biosiliceous sedimentation patterns for the Indian Ocean during the last 45 million years. In: Duncan, R., Rea, D., Kidd, R., von Rad, U., Weissel, J. (Eds.), Geophysical Monograph 70, The Indian Ocean: a synthesis of results from the Ocean Drilling Program. American Geophysical Union, Washington, DC, pp. 335–349.

Banse, K., 1984. Overview of the hydrography and associated biological phenomena in the Arabian Sea, off Pakistan. In: Haq, B., Milliman, J. (Eds.), Marine Geology and Oceanography of Arabian Sea and Coastal Pakistan. Van Nostrand Reinhold Company, New York, pp. 271–303.

Banse, K., 1987. Seasonality of phytoplankton chlorophyll in the central and northern Arabian Sea. Deep-Sea Research 34, 713–723.

Banse, K., 1994. On the coupling of hydrography, phytoplankton, zooplankton, and settling organic particles offshore in the Arabian Sea. In: Lal, D. (Ed.), Biogeochemistry of the Arabian Sea. Indian Academy of Sciences, Bangalore, India, pp. 27–63.

Banse, K., English, D., 1994. Seasonality of coastal zone color scanner phytoplankton pigment in the offshore oceans. Journal of Geophysical Research 99, 7323–7345.

Barber, R.T, Marra, J., Bidigare, R.C., Codispoti, L.A., Halpern, D., Johnson, Z., Latasa, M., Goericke, R., Smith, S.L., 2001. Primary productivity and its regulation in the Arabian Sea during 1995. Deep-Sea Research II 48, 1127–1172.

Bauer, S., Hitchcock, G., Olson, D., 1991. Influence of monsoonally-forced Ekman dynamics upon the surface layer depth and plankton biomass distribution in the Arabian Sea. Deep-Sea Research 38, 531–553.

Binet, D., Suisse de Sainte Claire, E., 1975. Le copepode planktonique *Calanoides carinatus*. Repartition et cycle biologique au large de la Côte d'Ivoire. Cahiers, ORSTOM Série Oceanographie 13, 15–30.

Böhm, E., Morrison, J., Manghnani, V., Kim, H.-S., Flagg, C., 1999. The Ras al Hadd Jet: Remotely sensed and acoustic Doppler current profiler observationsin 1994–1995. Deep-Sea Research II 46, 1531–1549.

Brock, J., McClain, C., Luther, M., Hay, W., 1991. The phytoplankton bloom in the northwestern Arabian Sea during the Southwest Monsoon of 1997. Journal of Geophysical Research 96, 20623–20642.

Bruce, J., 1974. Some details of upwelling off the Somali and Arabian coasts. Journal of Marine Research 32, 419–423.

Bruce, J., 1979. Eddies off the Somali Coast during the Southwest Monsoon. Journal of Geophysical Research 84, 7742–7748.

Burkill, P., Leakey, R., Owens, N., Mantoura, R., 1993. *Synechococcus* and its importance to the microbial foodweb of the northwestern Indian Ocean. Deep-Sea Research II 40, 773–782.

Campbell, L., Landry, M., Constantinou, J., Nolla, H., Brown, S., Liu, H., Caron, D., 1998. Response of microbial community structure to environmental forcing in the Arabian Sea. Deep-Sea Research II 45, 2301–2325.

Dickey, T., Marra, J., Sigurdson, D., Weller, R., Kinkade, C., Zedler, S., Wiggert, J., Langdon, C., 1998. Seasonal variability of bio-optical and physical properties in the Arabian Sea: October 1994–October 1995. Deep-Sea Research II 45, 2001–2025.

Duplessy, J., 1982. Glacial to interglacial contrasts in the Northern Indian Ocean. Nature 295, 494–498.

Elliott, A., Savidge, G., 1990. Some features of the upwelling off Oman. Journal of Marine Research 48, 319–333.

Fieux, M., Stommel, H., 1977. Onset of the southwest monsoon over the Arabian Sea from marine reports of surface winds: structure and variability. Monthly Weather Review 105, 231–236.

Fischer, A., 1997. Arabian Sea mixed layer deepening during the monsoon: observations and dynamics. S.M. Thesis, Massachusetts Institute of Technology, Boston, 130pp.

Flagg, C., Kim, H.-S., 1998. Upper ocean currents in the northern Arabian Sea from shipboard ADCP measurements collected during 1994–1996 US JGOFS and ONR programs. Deep-Sea Research II 45, 1917–1959.

Gardner, W., Gundersen, J., Richardson, M., Walsh, I., 1999. The role of seasonal and diel changes in mixed-layer depth on carbon and chlorophyll distributions in the Arabian Sea. Deep-Sea Research II 46, 1833–1858.

Garrison, D., Gowing, M., Hughes, M., 1998. Nano- and microplankton in the northern Arabian Sea during the Southwest Monsoon, August–September 1995: A US-JGOFS study. Deep-Sea Research II 45, 2269–2299.

Garrison, D., Gowing, M., Hughes, M., Campbell, L., Caron, D., Dennett, M., Brown, S., Liu, H.-B., Azam, F., Steward, G., Ducklow, H., Smith, D., 2000. Microbial food web structure in the Arabian Sea: a US JGOFS study. Deep-Sea Research II 47, 1387–1422.

Goyet, C., Millero, F., O'Sullivan, D., Eischeid, G., McCue, S., Bellerby, R., 1998. Temporal variations of pCO_2 in surface seawater of the Arabian Sea in 1995. Deep-Sea Research I 45, 609–623.

Haake, B., Ittekkot, V., Rixen, T., Ramaswamy, V., Nair, R., Curry, W., 1993. Seasonality and interannual variability of particle fluxes to the deep Arabian Sea. Deep-Sea Research I 40, 1323–1344.

Hansell, D., Peltzer, E., 1998. Spatial and temporal variations of total organic carbon in the Arabian Sea. Deep-Sea Research II 45, 2171–2193.

Harrison, D., Larkin, N., 1998. El Niño-Southern Oscillation sea surface temperature and wind anomalies 1946–1993. Reviews of Geophysics 36, 353–399.

Honjo, S., Dymond, J., Prell, W., Ittekkot, V., 1999. Monsoon-controlled export fluxes to the interior of the Arabian Sea. Deep-Sea Research II 46, 1859–1902.

Honjo, S., Weller, R., 1997. Monsoon winds and carbon cycles in the Arabian Sea. Oceanus 40, 24–28.

Husar, R., Prospero, J., Stowe, L., 1997. Characterization of tropospheric aerosols over the oceans with the NOAA advanced very high resolution radiometer optical thickness oeprational product. Journal of Geophysical Research 102, 16889–16909.

Ivanova, E., 1999. Late Quaternary monsoon history and paleoproductivity of the western Arabian Sea. Ph.D. Thesis, Free University, Amsterdam, The Netherlands, 172 pp.

Kim, H.-S., Flagg, C.N., Howden, S.D., 2001. Northern Arabian Sea variability from TOPEX/Poseidon altimetry data, an extension of the US JGOFS/ONR shipboard ADCP study. Deep-Sea Research II 48, 1069–1096.

Knox, R., 1987. The Indian Ocean: interaction with the monsoon. In: Fein, J., Stephens, P. (Eds.), Monsoons. Wiley, New York, pp. 365–397.

Krey, J., Babenerd, B., 1976. Phytoplankton Production Atlas of the International Indian Ocean Expedition. Intergovernmental Oceanographic Commission, Paris, 70pp.

Krishnamurti, T., 1987. Monsoon models. In: Fein, J., Stephens, P. (Eds.), Monsoons. Wiley, New York, pp. 467–522.

Kumar, S., Prasad, T., 1996. Winter cooling in the northern Arabian Sea. Current Science 77, 834–841.

Kutzbach, J., 1981. Monsoon climate of the early Holocene: Climate experiment with the Earth's orbital parameters for 9000 years ago. Science 214, 59–61.

Lampitt, R., Fahrbach, E., Schneider, B., Schneider, R., Smith, S., Toggweiler, J., Watson, A., Zeitzschel, B., 1994. Group Report: Does upwelling have a significant influence on the global carbon cycle. In: Summerhayes, C., Emeis, K.-C., Angel, M., Smith, R., Zeitzschel, B. (Eds.), Upwelling in the Ocean. Wiley, Chichester, pp. 383–404.

Latasa, M., Bidigare, R., 1998. A comparison of phytoplankton populations of the Arabian Sea during the Spring Intermonsoon and Southwest Monsoon of 1995 as described by HPLC-analyzed pigments. Deep-Sea Research II 45, 2133–2170.

Lee, C., Jones, B., Brink, K., Fischer, A., 2000. The upper-ocean response to monsoonal forcing in the Arabian Sea: seasonal and spatial variability. Deep-Sea Research II 47, 1177–1226.

Madhupratap, M., Parulekar, A., 1993. Biological processes of the Northern Indian Ocean. In: Ittekot, V., Nair, R. (Eds.), Monsoon Biogeochemistry. Geology-Paleontology Institute, University of Hamburg, Hamburg, Germany, pp. 51–72.

Madhupratap, M., Prasanna-Kumar, S., Bhattathiri, P., Dileep-Kumar, M., Raghukumar, S., Nair, K., Ramaiah, N., 1996. Mechanism of the biological response to winter cooling in the northeastern Arabian Sea. Nature 384, 549–552.

Manghnani, V., Morrison, J., Hopkins, T., Böhm, E., 1998. Advection of upwelled waters in the form of plumes off Oman during the Southwest Monsoon. Deep-Sea Research II 45, 2027–2052.

McCreary J.P. Jr., Kohler, K., Hood, R., Olson, D., 1996. A four-component ecosystem model of biological activity in the Arabian Sea. Progress in Oceanography 37, 193–240.

Measures, C., Vink, S., 1999. Seasonal variations in the distribution of Fe and Al in the surface waters of the Arabian Sea. Deep-Sea Research II 46, 1597–1622.

Mensah, M., 1974. The reproduction and feeding of the marine copepod *Calanoides carinatus* (Krøyer) in Ghanaian waters. Ghana Journal of Science 14, 167–191.

Meyers, P.A., Dickens, G.R., 1992. Accumulations of organic matter in sediments of the Indian Ocean: a synthesis of results from scientific deep sea drilling. In: Duncan, R., Rea, D., Kidd, R., von Rad, U., Weissel, J. (Eds.), Geophysical Monograph 70, The Indian Ocean: a synthesis of results from the Ocean Drilling Program. American Geophysical Union, Washington, DC, pp. 295–309.

Millero, F., Degler, E., O'Sullivan, D., Goyet, C., Eischeid, G., 1998. The carbon dioxide system in the Arabian Sea. Deep-Sea Research II 45, 2225–2252.

Molnar, P., England, P., Martinod, T., 1993. Mantle dynamics, uplift of the Tibetan plateau, and the Indian monsoon. Reviews of Geophysics 31, 357–396.

Morrison, J., Codispoti, L., Gaurin, S., Jones, B., Manghnani, V., Zheng, Z., 1998. Seasonal variation of hydrographic and nutrient fields during the US JGOFS Arabian Sea Process Study. Deep-Sea Research II 45, 2053–2101.

Murtugudde, R., Signorini, S., Christian, J., Busalacchi, A., McClain, C., Picaut, J., 1999. Ocean color variability of the tropical Indo-Pacific basin observed by SeaWiFS during 1997–1998. Journal of Geophysical Research 104, 18351–18366.

Nair, R., Ittekkot, V., Manganini, S., Ramaswamy, V., Haake, B., Degens, E., Desai, B., Honjo, S., 1989. Increased particle flux to the deep ocean related to monsoons. Nature 338, 749–751.

Passow, U., Peinert, R., Zeitzschel, B., 1993. Distribution and sedimentation of organic matter during the inter-monsoon period off Oman (West Arabian Sea). Deep-Sea Research II 40, 833–849.

Pease, P., Tchakerian, V., Tindale, N., 1998. Aerosols over the Arabian Sea: Geochemistry and source areas for aeolian desert dust. Journal of Arid Environments 39, 477–496.

Prell, W., 1984. Variation of monsoonal upwelling: a response to changing solar radiation. In: Hansen, J., Takahashi, T. (Eds.), Climate Processes and Climate Sensitivity. American Geophysical Union, Washington, DC, pp. 48–57.

Prell, W., Curry, W., 1981. Faunal and isotopic indices of the Southwest Monsoon upwelling, Arabian Sea. Oceanologica Acta 4, 91–98.

Prell, W., Hutson, W., Williams, D., Bé, A., Geitzenauer, K., Molfino, B., 1980. Surface circulation of the Indian Ocean during the last glacial maximum, approximately 18,000 yr B.P. Quaternary Research 14, 309–336.

Prell, W., Kutzbach, J., 1987. Monsoon variability over the past 150,000 years. Journal of Geophysical Research 92, 8411–8425.

Prell, W., Marvil, R., Luther, M., 1990. Variability in upwelling fields in the northwestern Indian Ocean. 2. Data-model comparison at 9000 years B.P. Paleoceanography 5, 447–457.

Prell, W., Murray, D., Clemens, S., Anderson, D., 1992. Evolution and variability of the Indian Ocean summer monsoon: evidence from the western Arabian Sea drilling program. In: Duncan, R., Rea, D., Kidd, R., von Rad, U., Weissel, J. (Eds.), Geophysical Monograph 70, The Indian Ocean: a synthesis of results from the Ocean Drilling Program. American Geophysical Union, Washington, DC, pp. 447–469.

Prell, W., Streeter, H., 1982. Temporal and spatial patterns of monsoonal upwelling along Arabia: A modern analogue for the interpretation of Quaternary SST anomalies. Journal of Marine Research 40, 143–155.

Prell, W., Van Campo, E., 1986. Coherent response of Arabian Sea upwelling and pollen transport to late Quaternary monsoonal winds. Nature 323, 526–528.

Premuzic, E., 1979. Organic carbon and nitrogen in the top sediments of world oceans and seas: distribution and relationship to bottom topography. Brookhaven National Laboratory, Informal Report, Upton, NY, 100pp.

Pye, K., 1987. In: Aeolian Dust and Dust Deposits. Academic Press, London, pp. 76–80.

Ramaswamy, V., 1993. Lithogenic fluxes to the northern Indian Ocean an – overview. In: Ittekot, V., Nair, R. (Eds.), Monsoon Biogeochemistry. Geology Paleontology Institute, University of Hamburg, Hamburg, Germany, pp. 97–111.

Sambrotto, R., 2001. Nitrogen production in the northern Arabian Sea during the Spring, Intermonsoon and Southwest Monsoon seasons. Deep-Sea Research II 48, 1173–1198.

Schnack, S., 1982. Feeding habits of *Calanoides carinatus* (Krøyer) in the upwelling area off northwest Africa. Rapports et proces-verbaux des reunions, Conseil permanent international pour l'exploration de la Mer 180, 303–306.

SCOR, 1990. The Joint Global Ocean Flux Study. JGOFS Report Number 5, 61pp.

Shalapyonok, A., Olson, R.J., Shalapyonok, L.S., 2001. Arabian Sea phytoplankton during Southwest and Northeast Monsoons 1995: composition, size structure and biomass from individual cell properties measured by flow cytometry. Deep-Sea Research II 48, 1231–1261.

Shimmield, G.B., Price, N.B., Pedersen, T.F., 1990. The influence of hydrography, bathymetry and productivity on sediment type and composition of the Oman Margin and in the northwest Arabian Sea. In: Robertson, A., Searle, M., Ries, A. (Eds.), The Geology and Tectonics of the Oman Region. Geological Society Special Publication 49, 759–769.

Shukla, J., 1987. Interannual variability of monsoons. In: Fein, J., Stephens, P. (Eds.), Monsoons. Wiley, New York, pp. 399–463.

Sirocko, F., 1993. Aerosol dust in late quaternary sediments of the northern Indian Ocean. In: Ittekot, V., Nair, R. (Eds.), Monsoon Biogeochemistry. Geology-Paleontology Institute, University of Hamburg, Hamburg, Germany, pp. 185–192.

Sirocko, F., Sarnthein, M., 1989. Wind-borne deposits in the northwestern Indian Ocean: Record of holocene sediments versus modern satellite data. In: Leinen, M., Sarnthein, M. (Eds.), Paleoclimatology and Paleometeorology: Modern and Past Patterns of Global Atmospheric Transport. Kluwer Academic Publishers, Dordrecht, pp. 401–433.

Smith, S., Banse, K., Cochran, J., Codispoti, L., Ducklow, H., Luther, M., Olson, D., Peterson, W., Prell, W., Surgi, N., Swallow, J., Wishner, K., 1991. US JGOFS: Arabian Sea Process Study, US JGOFS Planning Report Number 13, Woods Hole Oceanographic Institution, Woods Hole, Massachusetts, 164 pp.

Smith, R., Bottero, J., 1977. On upwelling in the Arabian Sea. In: Angel, M. (Ed.), A Voyage of Discovery. Pergamon Press, Oxford, United Kingdom, pp. 291–304.

Smith, S., Codispoti, L., Morrison, J., Barber, R., 1998a. The 1994–1996 Arabian Sea Expedition: an integrated, interdisciplinary investigation of the response of the northwestern Indian Ocean to monsoonal forcing. Deep-Sea Research II 45, 1905–1916.

Smith, S., Roman, M., Prusova, I., Wishner, K., Gowing, M., Codispoti, L., Barber, R., Marra, J., Flagg, C., 1998b. Seasonal response of zooplankton to monsoonal reversals in the Arabian Sea. Deep-Sea Research II 45, 2369–2404.

Tarran, G., Burkill, P., Edwards, E., Woodward, E., 1999. Phytoplankton community structure in the Arabian Sea during and after the SW monsoon, 1994. Deep-Sea Research II 46, 655–676.

Tindale, N., Pease, P., 1999. Aerosols over the Arabian Sea: atmospheric transport pathways and concentrations of dust and sea salt. Deep-Sea Research II 46, 1577–1596.

Veldhuis, M., Kraay, G., van Bleijswijk, J., Baars, M., 1997. Seasonal and spatial variability in phytoplankton biomass, productivity and growth in the northwestern Indian Ocean: the southwest and northeast monsoon. Deep-Sea Research I 44, 425–449.

Watson, A., 1994. Are upwelling zones sources of sinks of CO_2? In: Summerhayes, C., Emeis, K.-C., Angel, M., Smith, R., Zeitzschel, B. (Eds.), Upwelling in the Ocean. Wiley, Chichester, United Kingdom, pp. 321–336.

Webster, P., 1987. The elementary monsoon. In: Fein, J., Stephens, P. (Eds.), Monsoons. Wiley, New York, pp. 3–32.

Weller, R., Baumgartner, M., Josey, S., Fischer, A., Kindle, J., 1998. Atmospheric forcing in the Arabian Sea during 1994–1995: observations and comparisons with climatology and models. Deep-Sea Research II 45, 1961–2000.