5-2007

The Sensitivity of the Southwest Monsoon Phytoplankton Bloom to Variations in Aeolian Iron Deposition over the Arabian Sea

Jerry D. Wiggert
Old Dominion University

Raghu G. Murtugudde

Follow this and additional works at: https://digitalcommons.odu.edu/ccpo_pubs

Part of the [Oceanography Commons](#)

Repository Citation

Wiggert, Jerry D. and Murtugudde, Raghu G., "The Sensitivity of the Southwest Monsoon Phytoplankton Bloom to Variations in Aeolian Iron Deposition over the Arabian Sea" (2007). *CCPO Publications*. 233.
https://digitalcommons.odu.edu/ccpo_pubs/233

Original Publication Citation

Wiggert, J. D., & Murtugudde, R. G. (2007). The sensitivity of the southwest monsoon phytoplankton bloom to variations in aeolian iron deposition over the Arabian Sea. *Journal of Geophysical Research: Oceans*, 112(C5), C05005. doi:10.1029/2006jc003514

The sensitivity of the southwest monsoon phytoplankton bloom to variations in aeolian iron deposition over the Arabian Sea

Jerry D. Wiggert¹ and Raghu G. Murtugudde²

Received 31 January 2006; revised 24 June 2006; accepted 15 August 2006; published 1 May 2007.

[1] A coupled, 3-D biophysical ocean general circulation model is used to investigate how aeolian iron deposition affects the Arabian Sea ecosystem. Two separate aeolian iron deposition fields, derived from the GISS and GOCART atmospheric transport models, have been applied as surface boundary conditions. The model results exhibit widespread biogeochemical sensitivity to the choice of deposition field. With GOCART deposition, SW Monsoon phytoplankton blooms in the western and central Arabian Sea are enhanced and exhibit greater realism. The central Arabian Sea bloom is supported by supplemental input of horizontally advected iron from a pool that undergoes a yearlong progression that begins in the Gulf of Oman, where the difference in aeolian iron enrichment between the two deposition fields is most prevalent. The GOCART-enhanced blooms result in a more pronounced shift toward netplankton, an increase in euphotic zone export flux of up to a 20% during the SW Monsoon and an additional annual biogenic export of 3.5 TgC. The potential ramifications of regional N-cycle alteration through stimulation of N₂-fixation that is promoted by significant aeolian mineral flux needs to be explored. The canonical thinking that the northern Arabian Sea is invariably iron replete is now being challenged by both our model results and recent observational studies. As well, our results indicate that Arabian Sea iron concentrations are strongly modulated by the specific nature of aeolian mineral deposition. Thus climate or land use influences on dust mobilization could exercise leading-order controls on regional biogeochemical variability, metabolic status and air-sea exchanges of CO₂.

Citation: Wiggert, J. D., and R. G. Murtugudde (2007), The sensitivity of the southwest monsoon phytoplankton bloom to variations in aeolian iron deposition over the Arabian Sea, *J. Geophys. Res.*, 112, C05005, doi:10.1029/2006JC003514.

1. Introduction

[2] The Arabian Sea's three land boundaries distinguish it from other tropical oceans. In addition to inhibiting thermocline ventilation, these arid terrestrial regions are prominent sources of windblown mineral dust that gets deposited over the Arabian Sea. The Mudug Plateau in Somalia, the Wahiba Sand Sea in Oman, the Thar Desert in Pakistan/India, and the Rajasthan Desert in India are the Arabian Sea's primary dust sources [Léon and Legrand, 2003; Pease *et al.*, 1998; Prospero *et al.*, 2002]. These sites' relative contributions to the region's mineral aerosol distribution depends on the state of the seasonally reversing monsoon winds and the additional contribution to dust deflation of more localized atmospheric phenomena such as the Shamal winds over the Arabian Peninsula or mountain/valley winds over northeast Somalia [Léon and Legrand, 2003; Pease *et al.*, 1998]. Over the Arabian Sea, recent observations reveal that the maximal mean atmospheric dust loads near the Gulf of

Oman during the Northeast Monsoon (NEM) and the Spring Intermonsoon (SIM) are 5–6 times higher than during the Southwest Monsoon (SWM) [Pease *et al.*, 1998; Tindale and Pease, 1999]. This observed timing of the seasonal maximum is contrary to remote sensing based characterizations, used in developing models of atmospheric dust transport, that identify maximum dust loads during the SWM [Chin *et al.*, 2002; Tegen, 2003]. Nevertheless, a pronounced meridional gradient is evident in the annual rate of atmospheric dust deposition [see Jickells *et al.*, 2005, Figure 2]; and while annual dust fluxes to the Arabian Sea are among the world's highest, the spatiotemporal variability of this deposition is considerable.

[3] A portion of the deposited dust's ferrous component is solubilized and becomes part of the oceanic pool of dissolved iron, which is an essential micronutrient for phytoplankton photosynthesis [Martin *et al.*, 1991; Sunda and Huntsman, 1997]. Because of the elevated dust loads observed over the Arabian Sea, the canonical view has been that phytoplankton growth there is not iron limited [e.g., Barber *et al.*, 2001]. However, recent model results suggest that the region's pelagic ecosystem is not uniformly iron replete [Wiggert *et al.*, 2006]. Rather, our solution showed that pronounced spatiotemporal variability in biologically available iron manifests in response to the annually reversing currents and the seasonally developing

¹Center for Coastal Physical Oceanography, Old Dominion University, Norfolk, Virginia, USA.

²Earth System Science Interdisciplinary Center, University of Maryland, College Park, Maryland, USA.

upwelling, lateral advection and Ekman pumping that are driven by monsoonal forcing. During the SWM this dynamic response is at its peak and the distribution of iron limited surface waters is most widespread [see Wiggert *et al.*, 2006, Figure 15]. Similar conditions develop in response to the buoyancy driven convective mixing of the NEM. For both monsoon periods the surface distribution of dissolved iron realized by the model results directly from the higher N:Fe ratio prescribed for subsurface waters, since as the upwelled or entrained nutrients are utilized iron is the first to reach limiting concentrations. The predisposition of upwelled waters toward iron limitation is well established at other locations [e.g., Jickells *et al.*, 2005; Measures and Vink, 1999; Wu and Boyle, 2002], such as the upwelling regions off both California and Peru where surface waters are iron replete only when augmented by shelf sediments or river discharges [Hutchins *et al.*, 1998; Hutchins *et al.*, 2002]. As more extensive and appropriate observational studies are performed [e.g., Bishop *et al.*, 2002; Johnson *et al.*, 2003; Neuer *et al.*, 2004; Wells, 2003] there are corresponding advances in our capability to explicitly simulate oceanic iron biogeochemistry and the synthetic perspectives provided by these models contribute to our continually evolving understanding of iron's role in oceanic ecosystems [Moore *et al.*, 2004; Parekh *et al.*, 2004; Weber *et al.*, 2005].

[4] The biological response to SWM-period forcing consists of prominent phytoplankton blooms off the coasts of Oman [Brock and McClain, 1992; Latasa and Bidigare, 1998], Somalia [Hitchcock *et al.*, 2000; Veldhuis *et al.*, 1997] and India [Lierheimer and Banse, 2002]. Additionally, as the SWM evolves, offshore phytoplankton blooms consistently appear in ocean color observations and a persistent research question over the past several decades has consisted of resolving the attendant nutrient source mechanism(s). Early studies contended that these offshore blooms were supported by nutrients entrained by Ekman pumping [Bauer *et al.*, 1991; Brock *et al.*, 1991]. This mechanism was subsequently downplayed in favor of mixing and entrainment as the primary source of nutrients that was supplemented by advection from the Omani and Somali coastal zones [McCreary *et al.*, 1996]. A third viewpoint, based on results from a higher-resolution model, promoted the horizontal advection mechanism and contended that mesoscale processes significantly facilitated this offshore transport of coastally upwelled, high-nutrient waters from the Omani coast [Keen *et al.*, 1997; Young and Kindle, 1994]. More recently, an eddy-permitting model was used to demonstrate that advection from both the Omani and Somali coasts are the primary source mechanisms of offshore nutrients, with the Omani upwelling supplying nutrients to the northern Arabian Sea and the Somali upwelling supplying the central Arabian Sea [Kawamiya, 2001]. These latter modeling results regarding the primary nutrient source regions that support the offshore phytoplankton blooms were largely corroborated by a recent analysis of in situ and satellite based observations of physical fields [Kumar *et al.*, 2001]. A comprehensive review of our evolving understanding of the role these nutrient source mechanisms play in defining Arabian Sea biogeochemical variability is given by Wiggert *et al.* [2005].

[5] In our earlier study of ecosystem variability in the Indian Ocean [Wiggert *et al.*, 2006], the aeolian iron flux

boundary condition was derived from the GOCART atmospheric transport model [Ginoux *et al.*, 2001]. Here, an alternative model solution is presented, which has been obtained by applying the aeolian flux condition obtained from the GISS model [Tegen and Fung, 1994]. These two aeolian flux boundary conditions are used to investigate the Arabian Sea ecosystem's sensitivity to the spatiotemporal distribution of aeolian iron deposition. The biophysical model, the forcing data and the ecosystem model formulation are described in the Methods section. The two oceanic model solutions are then described; the resulting biogeochemical responses to the two boundary conditions are contrasted; and the implications of their differences are considered.

2. Methods

[6] The physical component of the model is a reduced gravity, primitive equation, sigma coordinate scheme that has been extensively characterized and applied in previous studies of the Indian Ocean [Murtugudde and Busalacchi, 1999; Murtugudde *et al.*, 2000; Murtugudde *et al.*, 1996]. The model grid is $1/2^\circ$ (longitudinal) by $1/3^\circ$ (latitudinal). Principal features of this model include coupling to an advective atmospheric mixed layer [Seager *et al.*, 1995] and a hybrid vertical mixing scheme developed by Chen *et al.* [1994b] that combines both bulk [Kraus and Turner, 1967] and gradient [Price *et al.*, 1986] Richardson number stability criteria. The mixed layer model is a primitive equation model that explicitly computes thermocline evolution and mixed layer/thermocline interactions and details on its coupling to this OGCM (ocean general circulation model) are detailed in Chen *et al.* [1994a]. A minimum mixed layer depth of 5 m has been prescribed but otherwise mixed layer evolution is determined through a fully prognostic equation that balances surface TKE with the stratification below.

[7] Daily climatological forcing is supplied by surface momentum flux, which is based on daily winds from the NCEP reanalysis [Kalnay *et al.*, 1996]. The NCEP wind fields are also included in the determination of air-sea exchanges of latent and sensible heat, causing these boundary conditions to have daily climatological variability as well. All other boundary conditions are based on monthly climatological data and their specific application to the OGCM has been described elsewhere [Christian *et al.*, 2002a; Murtugudde and Busalacchi, 1999; Wiggert *et al.*, 2006]. Initial conditions for temperature and salinity were based on the climatological distribution [Antonov *et al.*, 1998; Boyer *et al.*, 1998]. At the bottom boundary, the initial conditions of temperature, salinity, nitrate and iron are maintained. Along the southern boundary of the model domain, salinity, temperature and layer thickness were relaxed to seasonal climatological conditions between 25°S and 30°S . The Indonesian Throughflow and riverine inputs were not included in these experiments. The climatological solutions of the fully coupled biophysical model discussed here were integrated for 16 years. The initial physical conditions for these 16-year runs were taken from a 30-year spin-up of the physics only OGCM.

[8] The biological portion of the model is a nine-component oceanic ecosystem that has been fully coupled to the physical

OGCM and consists of a large and small size class for phytoplankton, zooplankton and detritus, as well as three phytoplankton nutrients (nitrate, ammonium and iron). Standard Michaelis-Menten nutrient uptake kinetics are applied, with the half-saturation coefficients of nitrate and iron uptake set at 0.4/0.8 μM and 25/120 pM, respectively, for the small/large phytoplankton size class. A comprehensive treatment of the ecosystem model and its explicit iron chemistry has been reported by *Christian et al.* [2002a, 2002b] and further details regarding this ecosystem model's application to the Indian Ocean can be found in the work of *Wiggert et al.* [2006]. Chlorophyll concentration is determined as a model diagnostic from a summation of the two phytoplankton size classes and an irradiance-dependent Chl:N ratio [*Geider et al.*, 1996] while modeled export flux time series shown herein are extrapolated from 400 m to the indicated depth by applying the Martin curve (i.e., $e(z) = e(400 \text{ m}) \cdot (400/z)^{0.858}$; *Martin et al.* [1987]).

[9] The initial nitrate distribution was taken from the NODC climatology [*Conkright et al.*, 1998]. The iron initial condition was based on this nitrate distribution using two constant N:Fe ratios, consisting of 2.5×10^5 in the upper 135 m and 3×10^5 at greater depth, with an iron solubility limit of 750 pM that reflects conditions in the IO [*Measures and Vink*, 1999; *Saager et al.*, 1989; *Takeda et al.*, 1995]. Losses of dissolved iron via particle scavenging occur at a maximum rate (10^{-4} d^{-1}) that is modified by the concentrations of iron and detrital material (complete details appear in *Christian et al.* [2002b]). The N:Fe ratio within the upper 135 m was chosen to be at the threshold between nitrogen and iron limitation for the netplankton. The subsurface condition was chosen to be slightly iron deficient relative to nitrogen and was based on N:Fe ratios reported for upwelled waters off the Arabian Peninsula [*Measures and Vink*, 1999]. These two initialization conditions allowed surface waters to evolve toward either nutritional state (i.e., N- or Fe-limited), depending on upwelling, atmospheric deposition and horizontal advection.

[10] Two separate climatological solutions were obtained for which the sole difference was which of two aeolian iron boundary conditions was applied. These two deposition boundary conditions were derived from the aeolian dust concentrations provided by the GOCART atmospheric transport model [*Ginoux et al.*, 2001] and the GISS atmospheric transport model [*Tegen and Fung*, 1994]. The primary distinction between these two transport models is in how sources of dust are identified. In the GOCART model there is no anthropogenic contribution and dust sources are associated with topographic depressions in arid or semiarid regions [*Ginoux et al.*, 2001], whereas in the GISS model all surfaces without tall vegetation are assumed to be potential sources of dust [*Tegen and Fung*, 1994]. The iron mass fraction in mineral dust (3.5%) and the solubility of aeolian iron in seawater (3%) were consistently applied to both dust fields. Furthermore, sources of iron in the present model configuration are limited to aeolian deposition and upwelling/mixing from below as no provision for contributions from riverine injection, sediment resuspension or elevated concentrations within suboxic waters has been included. Thus variations in modeled soluble iron distribution primarily relate to differences in how dust source regions are identified, the criterion for determining dust

emission to the atmosphere, the modeled atmospheric circulation, and how the aeolian iron flux from the two resulting deposition patterns are subsequently redistributed by the ocean model's circulation and biogeochemical components.

3. Results

[11] In the results that follow a number of model extraction sites and areal demarcations have been defined that will be foci for comparing the two model solutions and characterizing the differences in the mineral flux conditions. There are two sets of model extractions sites (AS2–AS4 and BR1–BR4), both of which are superimposed on Figure 1a. The first set was defined previously [*Wiggert et al.*, 2006] and are proximal to the location of three sediment traps deployed as part of the US JGOFS Arabian Sea Process Study [*Honjo et al.*, 1999]. Data from these traps were part of the comprehensive model-data comparison that was performed on the main run climatological solution that utilized the GOCART-derived deposition fields [*Wiggert et al.*, 2006]. This comparison made extensive use of the U. S. JGOFS measurements from the 1994–1995 Arabian Sea Process Study such as seasonally resolved in situ profiles of nutrients, particulate matter and Chla, production rate experiments, measurements of zooplankton biomass and HPLC-based characterizations of phytoplankton speciation. Herein, we forego repeating the expansive model validation that appears in the work of *Wiggert et al.* [2006] and rely on it as the foundation for extending the model's application to the current investigation. Of particular relevance to the analysis presented here are comparisons to Chla evolution within the upper water column and seasonal shifts in phytoplankton speciation at sites AS2–AS4. Unfortunately, only measurements of total dissolved iron have been made available from the JGOFS surveys, so comparison between those published values and the bioavailable (i.e., soluble) iron carried by the model was not attempted.

[12] The second set of model extraction sites (BR1–BR4) was chosen to characterize how the biogeochemical responses varied between the two model solutions, with the specific locations based on the clear differences in the July distribution of surface chlorophyll a (Chla) (Figures 1a and 1b). Superimposed on the GISS solution's Chla field (Figure 1b) are eight areal demarcations that will be used to compare local aeolian flux between the two boundary conditions. These equal area flux boxes have an offshore (relative to Oman) group (FB1–FB4) and a near shore group (FB5–FB8) and their boundary definitions are provided in Table 1. The numerical identifier of the four biogeochemical response sites (BR1–BR4) coincides with the flux box (FB1–FB4) in which a given extraction site is situated. The coordinates for sites BR1–BR4 are provided in the caption for Figure 1.

[13] In addition, three symbols or acronyms have been introduced with the intention that these would allow for a clearer and more succinct report. All three represent differences between the two aeolian flux experiments that are always defined as GOCART-GISS. The symbols are ΔExp (difference in export flux at 200 m), ΔAE (difference in surface layer aeolian iron enrichment) and ΔFe (difference in surface iron concentration). While they also appear

Figure 1. Modeled and observed July surface Chla (mg m^{-3}) for the Arabian Sea. (a) GOCART and (b) GISS are the two aeolian deposition cases while (c) is the climatological distribution observed by SeaWiFS. The eight boxes used to quantify the differences in surface iron enrichment (see Table 1 for box boundary definitions) are superimposed on the GISS Chla field (Figure 1b). The four time series extraction sites (BR1–BR4, black diamonds) are shown in both Figures 1a and 1b. The coordinates for the four extraction sites are BR1 (20°N , 66°E); BR2 (14°N , 62°E); BR3 (12.5°N , 64°E); and BR4 (11.5°N , 60°E). The three additional sites (AS2–AS4, red squares) shown on Figure 1a depict the location of three JGOFS sediment trap sites. Data from these traps are used in the model-data export flux comparison in Figure 5, and additional details regarding these trap sites are provided in its accompanying caption. The SeaWiFS climatology (Figure 1c) was created using monthly Level 3 Standard Mapped Image data from September 1997 through January 2002, which were obtained from the Goddard DAAC (<http://daac.gsfc.nasa.gov/>). Figure 1c reprinted, with modification, from Wiggert *et al.* [2005], with permission from Elsevier.

appropriately in the manuscript, the above definitions and descriptions have been collected here to simplify subsequent referral. Finally, oceanic features and geographical locations in and around the Arabian Sea mentioned within this report are shown in Figure 1a.

3.1. Distinct Ecosystem Responses in the GOCART and GISS Solutions

[14] A profound difference in Chla between the two model simulations appears in July. In the GOCART solution a broad tongue exceeding 0.8 mg m^{-3} extends across the

central AS between 11.5°N and 15°N (Figure 1a). In contrast, the GISS solution indicates somewhat elevated Chla (up to 0.6 mg m^{-3}) that lies along and northward of 15°N (Figure 1b), which is thus slightly farther north than its more pronounced GOCART counterpart. Less prominent distinctions between the two solutions are also evident. These consist of a more extensive coastal phytoplankton bloom off of the Arabian Peninsula in the GOCART result and somewhat higher concentrations north of 20°N in the GISS result. Alternatively, the blooms that extend around the northern limb of the Great Whirl (centered at 56°E , 8°N)

Table 1. Rates of Aeolian Iron Enrichment for Surface Waters Within the Eight Aeolian Flux Boxes (FB1–FB8) in the Arabian Sea (Figure 1b) for the Two Atmospheric Transport Models (GOCART and GISS)^a

Location	Maximum Monthly Rate		Maximum Δ	Annual
	GOCART	GISS	Δ AE	Δ AE
Flux Box 1 (65.0–68.0E,18.0–22.0N)	11.6 [MAY]	28.0 [JUL]	(–19.6 [JUL])	0.4
Flux Box 2 (58.5–62.5E,12.5–15.5N)	5.9 [JUN]	3.1 [JUL]	3.9 [MAY]	20.0
Flux Box 3 (62.5–66.5E,11.0–14.0N)	4.1 [JUN]	3.9 [JUL]	2.6 [MAY]	4.6
Flux Box 4 (58.0–62.0E,9.0–12.0N)	3.1 [MAY]	1.9 [JUN]	1.9 [MAY]	6.8
Flux Box 5 (60.5–63.5E,19.5–23.5N)	31.0 [JUN]	7.5 [JUL]	27.5 [JUN]	165.0
Flux Box 6 (59.5–63.5E,16.0–19.0N)	11.3 [MAY]	5.7 [JUL]	9.7 [MAY]	43.1
Flux Box 7 (54.0–58.0E,13.0–16.0N)	14.9 [JUN]	5.0 [JUL]	11.7 [JUN]	49.5
Flux Box 8 (52.0–56.0E,9.0–12.0N)	15.1 [JUL]	3.0 [MAY]	12.5 [JUL]	56.3

^aThe coordinates that prescribe the boundaries of each box are provided in column 1. The maximum monthly rate, with the month noted in brackets, is provided for each deposition field in columns 2 and 3. The fourth column contains the maximum monthly difference in aeolian enrichment (Δ AE) between the two models. For all but one aeolian flux box (denoted by parentheses), the enrichment rate derived from the GOCART model is larger in magnitude. The fifth column contains the annual Δ AE for each box. Units are $\text{nmol Fe m}^{-3} \text{ d}^{-1}$.

and off the southwest coast of India are essentially identical in the two solutions. The July climatological Chl*a* distribution obtained from SeaWiFS data shows an extensive offshore maximum that is similar in spatial extent to that appearing in the GOCART result, though the observed concentrations are somewhat higher and east of 62°E extend to the northeast whereas in the model they extend due east (Figures 1a and 1c). The blackened areas in the July image indicate no available unflagged data in the standard SeaWiFS SMI product over the 4 years of observations that went into the climatological distribution (Figure 1c); the monthly climatologies based on 8 years of data that are now available from the NASA OceanColor Web (<http://oceancolor.gsfc.nasa.gov/>) show this gap as well. This is due to the extensive cloudiness and aerosols that adversely affect the successful retrieval of remotely sensed chlorophyll at this time. The clear difference in Chl*a* between the SeaWiFS data and the model results could relate to the transformation of phytoplankton biomass to pigments applied in the model or to the satellite retrievals being adversely affected by the absorbing aerosols just noted; further discourse on both possibilities has appeared elsewhere [Wiggert *et al.*, 2005; Wiggert *et al.*, 2006].

[15] Four time series extraction locations were chosen to compare the seasonality of the model solutions to the SeaWiFS climatology. The black diamonds in Figures 1a and 1b indicate the locations of these biogeochemical response sites (BR1–BR4). The extracted time series of surface Chl*a* from the two solutions are shown in Figure 2. At the northernmost site (BR1) the two models are quite similar, as the timing of the semiannual phytoplankton blooms is consistent while the maximum summertime concentration is $\sim 20\%$ greater in the GISS result (Figure 2a). Both sites within the GOCART experiment's offshore bloom (BR2 and BR3) show that this solution better captures the observed magnitude and timing of the peak summertime Chl*a* values (Figures 2b and 2c). At BR3 in particular, the July appearance of the SWM phytoplankton bloom is replicated in the GOCART result whereas this summer bloom is weaker and occurs later in the GISS result (Figure 2c). Site BR4 lies to the south of the offshore bloom and the extracted Chl*a* time series exhibit little difference between the two solutions (Figure 2d). The only notable dissimilarity is the magnitude of the July peak in Chl*a*, with

the GOCART result showing better agreement with the SeaWiFS climatology.

[16] The temporal evolution of Chl*a* through the upper 120 m is shown for the GOCART and GISS solutions at BR1 and BR2 (Figure 3). Corresponding figures for the other two sites are not included as the differences between the two solutions at BR3 are largely similar to those at BR2, while at BR4 there is little to distinguish the two solutions. Superimposed on the Chl*a* distributions are time series of mixed layer depth (thick black line) and depth of the 0.2 μM nitrate isopleth (thick white line). The superimposed contour lines (thin black lines) in Figure 3 illustrate the relative contributions of picoplankton (P_S) and netplankton (P_L) to total phytoplankton biomass. This “P-ratio” ($P_R \equiv P_S/(P_S + P_L)$) is used to illustrate phytoplankton speciation shifts in the solution. For $P_R > 0.5$, picoplankton is the predominant component of phytoplankton biomass; for $P_R < 0.5$ netplankton dominate. Thus increased export of organic matter from surface waters is associated with low values of P_R that indicate a transition toward the larger size class. The temporal evolution of dissolved iron at these two sites provides further insight into why distinct ecosystem responses manifest between the two solutions (Figure 4). Superimposed on these iron distributions are the 0.2 μM (thin solid line) and 0.6 μM (thick solid line) nitrate isopleths. In addition, the 60 pM (thin dashed line) and 120 pM (thick dashed line) dissolved iron isopleths are shown. Interestingly, at or below 80 m differences in the seasonal evolution of iron concentration in the two solutions are minimal, yet profound differences in near-surface availability are evident (Figure 4).

[17] At BR1, the intensity of the deep chlorophyll maximum (DCM) is the most prominent distinction between the two cases (Figures 3a and 3b). While the depth of the DCM is consistent and its seasonal evolution is similar, Chl*a* in the GOCART solution is higher by 0.1–0.2 mg m^{-3} from March through December. Additionally, the shift toward netplankton ($P_R < 0.35$) that coincides with the maximal DCM concentrations occurs 2 months earlier in the GOCART solution (Figures 3a and 3b). In both solutions, the top of the DCM is closely aligned with the 0.2 μM nitrate isopleth. The clearest difference in this isopleth's evolution occurs during the Fall Intermonsoon (FIM) when it resides at 55 m in the GOCART solution and shoals from 50 m to 35 m in the GISS solution. Both the deeper

Figure 2. Time series of surface Chl *a* (mg m⁻³) from the GOCART (solid line), GISS (dashed line), and SeaWiFS climatology (crosses) at the four extraction sites (BR1–BR4).

penetration of this isopleth during the FIM and the more pronounced shift toward netplankton during the latter half of the SWM in the GOCART solution suggest stronger iron limitation at this location in the GISS solution. Indeed, near-surface iron concentrations in the GOCART solution remain above 100 pM and exceed 200 pM from June into November whereas in the GISS solution they do not exceed 60 pM from May through December (Figures 4a and 4b). During the SIM and early SWM the nitrate isopleth is consistently up to 5 m shallower in the GISS solution, because of lower iron concentrations (Figures 3a, 3b, 4a, and 4b). Thus the August entrainment bloom, which coincides with the mixed layer penetrating below the 0.2 μ M nitrate isopleth and a release of iron limitation by local deposition (see section 3.3), is stronger in the GISS solution because of a somewhat larger subsurface nitrogen pool.

[18] The most prominent differences between the two solutions at BR2 are notably higher Chl *a* in the DCM during the SIM and in the surface bloom that manifests in July in the GOCART solution (Figures 3c and 3d). The

bloom in July coincides with a deepening of the mixed layer and an associated shoaling of the 0.2 μ M nitrate isopleth. Additionally, this SWM bloom is more than 10 m shallower in the GISS solution. In both the SIM-period DCM and the July surface bloom, the P_R contours show a stronger shift toward netplankton in the GOCART solution. This distinction is most pronounced within the July bloom, which exhibits P_R below 0.35 down to 100 m, whereas in the GISS solution P_R is always > 0.35 . An especially striking difference between the two solutions is the 0.2 μ M nitrate isopleth which resides at ~ 60 m during the SWM bloom in the GOCART solution and which extends to the surface in the GISS solution, indicating that the ecosystem in the latter case is severely iron limited. Indeed, in the GOCART solution iron concentrations exceed 60 pM from the latter NEM through the mid-SWM and, except in September, are always greater than 25 pM (Figures 4c and 4d). On the other hand, in the GISS solution iron concentrations over the upper 35 m remain below 25 pM throughout the year and the 60 pM isopleth never shoals above 50 m. Furthermore,

Figure 3. Chla time series over the upper 120 m at (a) BR1 with GOCART deposition; (b) BR1 with GISS deposition; (c) BR2 with GOCART deposition; and (d) BR2 with GISS deposition. The superimposed contours (thin black lines) are $P_R (=P_S/(P_S + P_L))$. The thick white line shows the temporal evolution of the $0.2 \mu\text{M}$ nitrate isopleth. The thick black line shows the temporal evolution of mixed layer depth, which is identical for both solutions and included to provide some context of the local physical environment.

the difference in nitrate concentration (GOCART-GISS) from 65 m to 120 m ranges from -0.3 to $-0.6 \mu\text{M}$ (data not shown).

3.2. Export Flux Patterns in the GOCART and GISS Solutions

[19] A comparison of export flux in both solutions with sediment trap observations at sites AS2–AS4 is shown in Figure 5. The main difference between the two solutions is confined to the latter SWM when export flux is consistently higher in the GOCART solution. The peak rates in August are 8% to 73% higher than those in the GISS solution. Maximal export flux during this time frame in the GOCART solution is in accord with the observed rates at AS2 and AS3, while at AS4 no coincident observations are available. During the less productive periods, export flux in the two solutions is essentially identical. The comparison to the observed fluxes further shows that export of organic matter in both solutions is consistently too low during the NEM, especially at AS2. Some of the high-frequency flux variability at AS3 is associated with passing mesoscale features [Honjo *et al.*, 1999]; thus capturing such variability in a $1/2^\circ \times 1/3^\circ$ model application is not to be expected.

[20] The clear distinction between the two solutions in the phytoplankton speciation shifts that occurs at BR2 in July is expected to manifest as a modification in the subsequent export flux. A time series comparison of export flux at 200 m for the two cases is provided for all four sites in Figure 6. These demonstrate maximal export in August (Figures 6b and 6c), with 60–90% greater particle flux at BR2 and BR3 in the GOCART solution as well as fluxes that are more than 100% greater in July and 20–40% greater in September. Over the rest of the year, there is little difference between the two solutions. On an annual basis, GOCART particle flux is greater by 35% and 52% at BR2 and BR3, respectively. At the other two sites there is little difference in export over the entire annual cycle (Figures 6a and 6d) and the annual mean differences are 3% at BR1 and 14% at BR4.

[21] Clearly, the most pronounced export differences between the two solutions occur during the latter stages of the SWM as organic matter resulting from the summer bloom descends through the water column at BR2 and BR3. The spatial pattern of the difference in export flux between the two solutions (ΔExp , GOCART-GISS) for July–September is shown in Figure 7. Consistent with the time

Figure 4. Dissolved iron time series (nmol Fe m^{-3}) over the upper 120 m at (a) BR1 with GOCART deposition; (b) BR1 with GISS deposition; (c) BR2 with GOCART deposition; and (d) BR2 with GISS deposition. The thin and thick solid lines show the temporal evolution of the $0.2 \mu\text{M}$ and $0.6 \mu\text{M}$ nitrate isopleths, respectively. The thin and thick dashed lines show the 60 pM and 120 pM iron isopleths, respectively. In the model the half-saturation coefficients for nitrate uptake by small and large phytoplankton are $0.4 \mu\text{M}$ and $0.8 \mu\text{M}$, whereas the corresponding half-saturation coefficients for iron uptake are 25 pM and 120 pM.

series in Figure 6, maximal ΔExp occurs in August (Figure 7b and Table 2). In general, the excess GOCART export flux shifts eastward as the SWM progresses. In July there are two maxima, both having peak ΔExp of $0.3\text{--}0.4 \text{ mmol N m}^{-2} \text{ d}^{-1}$

(Figure 7a). One is located just offshore of the western Arabian Peninsula at 14°N , 56°E while the other appears in the central Arabian Sea around 13°N , 61.5°E . By August, the offshore maximum has significantly increased in size

Figure 5. Time series of export flux ($\text{mmol N m}^{-2} \text{ d}^{-1}$) from the GOCART (solid line) and GISS (dashed line) model solutions compared with JGOFS sediment trap data (asterisks) at (a) AS2 (17.2°N , 59.8°E), (b) AS3 (16°N , 62°E), and (c) AS4 (10°N , 65°E). These location designations are as defined originally by Wiggert *et al.* [2006], and additional information can be found therein. Model export fluxes have been extrapolated to the trap depth (873 m, 814 m, and 800 m) using the Martin curve [Martin *et al.*, 1987]. Reprinted, with modification, from Wiggert *et al.* [2006], with permission from Elsevier.

Figure 6. Time series of export flux ($\text{mmol N m}^{-2} \text{d}^{-1}$) at 200 m for the GOCART (solid line) and GISS (dashed line) solutions at (a) BR1, (b) BR2, (c) BR3, and (d) BR4.

and magnitude (up to $0.8 \text{ mmol N m}^{-2} \text{d}^{-1}$) and has moved northeastward, while the coastal maximum has diminished and moved offshore somewhat (Figure 7b). By September the coastal feature has largely dissipated (Figure 7c), peak values of ΔExp in the offshore maximum are lower ($0.4\text{--}0.5 \text{ mmol N m}^{-2} \text{d}^{-1}$), and this feature's core has moved to the southeast (15°N , 65.5°E). Over the other 9 months ΔExp never exceeds $\pm 0.1 \text{ mmol N m}^{-2} \text{d}^{-1}$ south of 22°N (data not shown) while for the Arabian Sea region as a whole ΔExp is $\leq 3.2 \text{ Tg C yr}^{-1}$ (Table 2).

[22] Table 2 demonstrates that the three highest monthly export rates occur in July–September, a characteristic that is well established within the observational literature [e.g., *Rixen et al.*, 2005]. The GOCART solution exhibits export fluxes for the Arabian Sea as a whole that are 18–21% higher in July and August (Table 2). On an annual basis ΔExp for the entire Arabian Sea is 3.5 Tg C with 56% of this occurring over the 3 months (July–September) featured in Figure 7. For the Indian Ocean, the annual difference is 4.6 Tg C with 54% of this additional export of organic matter occurring in the July–September time frame (Table 2). On a monthly basis, ΔExp within the Arabian Sea accounts for 55–100% of basinwide ΔExp and accounts for 83–88% of the basinwide enhancement during July and August.

3.3. Differences in Aeolian Mineral Deposition Distribution and the Resulting Soluble Iron Fields

[23] In their annual mean distributions (data not shown), the maximum difference in aeolian enrichment (ΔAE ,

GOCART-GISS) hugs the east coast of Oman whereas the maximum positive difference in iron concentration (ΔFe , GOCART-GISS) is shifted eastward to the Pakistan/India coast and from there extends southwestward into the Arabian Sea. The dissimilarity between the ΔAE and ΔFe distributions is a result of how the variations in deposition timing and magnitude of the two aeolian boundary conditions convolve with the subsequent advection, biological utilization or loss of soluble iron to scavenging. To characterize the spatiotemporal differences in surface enrichment between the two aeolian deposition patterns, eight equal area boxes have been defined. As noted at the beginning of section 3, these areal demarcations are shown on the GISS solution's July Chla distribution (Figure 1b) and their boundary definitions are listed in Table 1. Comparisons of surface enrichment time series for the two deposition fields are presented as bar graphs in Figure 8. Two fundamental patterns emerge in this comparison when the flux boxes are grouped by distance from the Arabian coast. In the offshore group (FB1–FB4), the increase in surface layer iron concentration from GISS deposition is only higher during the SWM (July, August and September), whereas in the shoreward group (FB5–FB8) GOCART deposition always provides greater iron enrichment. Annual surface enrichment via GOCART deposition everywhere exceeds that associated with GISS deposition, though within FB1, FB3 and FB4 the difference is less than $7 \text{ nmol Fe m}^{-3} \text{d}^{-1}$ (Table 1).

[24] The time series show that maximal surface enrichment under the GOCART boundary condition occurs within FB5 in June, whereas for GISS deposition this occurs within

Figure 7. Distributions of the difference in export flux ($\text{mmol N m}^{-2} \text{d}^{-1}$) at 200 m between the two solutions (GOCART-GISS) over the entire Arabian Sea for (a) July, (b) August, and (c) September.

FB1 in July (Figures 8a and 8e). Though it is still positive, this latter site has the lowest annual net difference ($0.4 \text{ nmol Fe m}^{-3} \text{d}^{-1}$) due to the predominant GISS-derived surface enrichment during July and August (Figure 8a and Table 1). The timing difference between the GOCART and GISS enrichment maxima apparent in boxes FB1 and FB5 generally persists over the entire region. This is summarized in Table 1 where the maximum monthly enrichment rates for the two cases are listed with the month of their occurrence in brackets. This demonstrates that maximal GOCART enrichment occurs in May or June with one exception (the FB8 peak occurs in July), while maximal GISS enrichment typically occurs in July. Furthermore, the maximum (absolute) ΔAE also typically occurs in May or June and, except at FB1, is always positive (Table 1).

[25] With regard to the more prominent central Arabian Sea phytoplankton bloom in the GOCART solution (Figures 1a and 1b), the time series indicate that aeolian enrichment over these two areas (FB2 and FB3) is quite low in July (Figures 8b and 8c). Indeed, the maximum July enrichment rate over these two boxes is $3.9 \text{ nmol Fe m}^{-3} \text{d}^{-1}$ (Table 1), with GISS enrichment being slightly higher at both sites (Figures 8b and 8c). However, relative to other locations this maximal local enrichment rate is quite modest, and is more

than 85% lower than the maxima at FB1 (GOCART) or FB5 (GISS). While its magnitude and spatial distribution in the model solution depends on which of the two aeolian mineral flux conditions is applied, this offshore phytoplankton

Table 2. Monthly Export Production at 200 m^a

Month	Export, GISS: Arabian Sea	Arabian Sea	ΔExp	Indian Ocean
JAN	23.4	1.8		2.9
FEB	20.0	1.3		2.5
MAR	14.9	1.3		1.9
APR	13.3	1.8		1.9
MAY	13.5	2.0		2.0
JUN	16.2	2.1		2.4
JUL	35.3	7.5		8.6
AUG	61.5	11.0		13.3
SEP	47.7	5.2		8.2
OCT	32.9	3.2		5.6
NOV	27.4	2.9		3.8
DEC	23.8	2.4		2.9

^aThe transformation to carbon currency is based on a constant C:N ratio. The second column shows monthly flux rates in the Arabian Sea (48°E – 78°E north of 4°N) for the GISS solution. Columns 3 and 4 show the ΔExp for the Arabian Sea and the Indian Ocean (35°E – 125°E north of 20°S), respectively. Units are Tg C yr^{-1} .

Surface Iron Enrichment by Aeolian Deposition

Figure 8. Time series of surface layer iron enrichment rate ($\text{nmol Fe m}^{-3} \text{d}^{-1}$) for the two deposition fields over the eight aeolian flux boxes (FB1–FB8) depicted in Figure 1b. These values represent the increase in soluble iron concentration afforded by aeolian deposition of dust into the surface layer of the model. For these simulations, 3% of the deposited iron entered the soluble pool and became available for biological uptake.

bloom is clearly not the product of local aeolian deposition by either aeolian flux condition.

[26] In order to more fully explore how the differences in the two deposition fields impact the four offshore locations, time series of surface enrichment by aeolian deposition, entrainment and mixing, and horizontal advection have been extracted from the model solutions (Figure 9). These underscore the 1–2 month lag in peak surface enrichment associated with GISS deposition (Figures 9a and 9d), which occurs at BR1–BR4. These time series also indicate that the mixed layer deepening that accompanies the onset of the SWM acts to dilute surface iron concentrations at BR1, and in June at BR2, when GOCART deposition is applied (Figure 9b). At BR1 the mixed layer does not deepen below 50 m during the SWM (Figure 3a). Thus the local ferrocline is not reached and the surface iron maximum must fuel the surface phytoplankton bloom in August and act as a source to the waters below as the mixed layer slowly deepens (Figures 3a and 4a). At BR2, the loss via mixing of iron from the surface layer indicated in the GOCART solution (Figure 9b) is also associated with a surface maximum (Figure 4b). Here the surface maximum is much less prominent and becomes quickly dissipated by the mixed layer's more rapid descent (Figure 3c). Moreover, as

evidenced by the shoaling iron isopleths, the mixed layer extends into the ferrocline in July at BR2, which results in the sign reversal that occurs in the entrainment/mixing contribution to the surface iron budget (Figures 4c and 9b). At BR3 and BR4, the oceanic response to monsoon forcing results in surface enrichment. Alternatively, response to the Findlater Jet drives entrainment of iron into surface waters at all four sites when GISS deposition is applied (Figure 9e), a direct result of the iron distribution exhibiting a more typical nutrient profile (Figures 4b and 4d). In July, peak rates of surface layer iron enrichment via mixed layer deepening attained at BR2 and BR3 are similar in magnitude ($140\text{--}170 \text{ nmol Fe m}^{-3} \text{d}^{-1}$) in both solutions (Figures 9b and 9e).

[27] The horizontal advection of bioavailable iron at the four sites shows profound differences between the two solutions (Figures 9c and 9f). At BR1, advection enhances the iron pool during the SWM and reduces it during the NEM when GOCART deposition is applied, while the opposite trends are true under GISS deposition. In August at BR4, both cases show increases in iron concentration resulting from horizontal advection, with the GOCART solution also exhibiting such enrichment in June. At BR2 and BR3, the GISS solution indicates that horizontal ad-

Figure 9. Surface layer iron budget time series at the four extraction sites (BR1–BR4) for the two deposition boundary conditions. The three components of the budget are (a) aeolian enrichment, (b) entrainment and mixing, and (c) horizontal advection for the GOCART solution. (d–f) Corresponding budget terms for the GISS solution.

vection does not impact the local iron pool, whereas the GOCART solution indicates that significant additions of soluble iron ($> 80 \text{ nmol Fe m}^{-3} \text{ d}^{-1}$) occur during June and July (Figures 9c and 9f). This distinction is reversed for nitrate transport; no addition of nitrate via horizontal advection at BR2 manifests in the GOCART solution whereas in the GISS solution this occurs at a rate of $0.5\text{--}0.75 \text{ mmol NO}_3 \text{ m}^{-3} \text{ d}^{-1}$ in July and August (data not shown). This is consistent with the shoaling of the $0.2 \mu\text{M}$ nitrate isopleth that appears in the GOCART solution at this time (Figure 3d).

[28] To attain a broader view of the differences in aeolian enrichment of surface iron and the accompanying advection pathways, flowlines derived from the monthly surface current fields are superimposed on the ΔAE distributions

for the 4 months that culminate in the July appearance of the central Arabian Sea bloom (Figure 10). This time frame encompasses the late SIM through the mid-SWM, when maximal aeolian enrichment occurs everywhere (Figures 8, 9a, and 9d). The spatial transition between GOCART- and GISS-dominated aeolian enrichment is demarked by the zero isoline in the four panels of Figure 10. To provide further context for the differences between the two model experiments for the early SWM time frame, the iron flux boundary conditions derived from the two atmospheric transport models are shown in Figure 11. It should be kept in mind when comparing Figures 10 and 11 that the mapping between surface enrichment and surface iron flux will not be exact since mixed layer depth factors into the former.

Figure 10. Monthly difference in aeolian iron enrichment ($\text{nmol Fe m}^{-3} \text{d}^{-1}$) between the two deposition boundary conditions (ΔAE , GOCART-GISS) for (a) April, (b) May, (c) June, and (d) July. The zero ΔAE isoline is shown for emphasis (thick solid lines). The superimposed flowlines are pathline integrations of each month's surface current field that are representative of the horizontal advection pathways. Arrows qualitatively indicate spatial variation in current speed but are not comparable between months.

[29] One trend revealed by these ΔAE distributions is that the area over which GISS enrichment exceeds GOCART enrichment expands continually over this period (Figure 10). By July, this encompasses the eastern Arabian Sea and stretches southwestward across the central Arabian Sea. This is consistent with the iron flux distribution, especially for waters off the west coast of India around the Gulf of Khambhat that receive the most notable additional iron input from GISS deposition (Figures 10d, 11c, and 11d). The other notable characteristic relates to the Arabian Peninsula's coastal regions, which over the period shown here receive progressively greater iron enrichment from GOCART deposition (Figure 10). The iron flux distributions also demonstrate this trend and it can in particular be seen that aeolian deposition over the Gulf of Oman and the Gulf of Aden is notably higher in the GOCART case (Figure 11). The two gulfs exhibit the greatest difference in surface enrichment, with values exceeding $80 \text{ nmol Fe m}^{-3} \text{d}^{-1}$ in both June and July (Figures 10c and 10d), which is more than 3x greater than the largest ΔAE for FB1–FB8 (Figure 8 and Table 1). For the Gulf of Oman, an enrichment difference of this magnitude also occurs in May (Figure 10b).

[30] The ΔAE distributions suggest that the supplemental source of soluble iron to the central Arabian Sea originates in either the Gulf of Oman or the western Arabian Peninsula/Gulf of Aden (Figures 10b–10d). The superimposed flowlines indicate a direct link from elevated deposition over the western Arabian Sea, whereas transport from the Gulf of Oman seems less likely. In April there is a weak cyclonic circulation present that could provide a pathway between the Gulf of Oman and the offshore bloom region south of 15°N. However, this circulation pattern is absent in May and surface waters from the Gulf of Oman look to be transported eastward toward the coast of Pakistan. Thus transportation of soluble iron directly from the major deposition region east of Oman to the offshore bloom region appears unlikely.

[31] To determine whether soluble iron is indeed horizontally advected from the west, the monthly mean current speed was determined within two boxes that extend eastward from BR2 to the Gulf of Aden (13.3–13.7°N, 52–62°E) and the western Arabian Peninsula (15.3–15.7°N, 54–62°E). Since in April these areas exhibit no spatial gradient in ΔAE as well as a cyclonic recirculation (Figure 10a), only the mean current speeds for May and June were

Figure 11. Distributions of iron flux boundary condition ($\mu\text{mol m}^{-2} \text{d}^{-1}$) for the two deposition fields applied as model forcing in June (a) GOCART, (b) GISS, and July (c) GOCART, (d) GISS.

considered in ascertaining the likelihood of these western waters acting as a source of soluble iron to the central Arabian Sea. From the mouth of the Gulf of Aden the mean speeds were 7.3 cm/s in May and 14.8 cm/s in June, while from the western Arabian Peninsula they were 5.2 cm/s and 11.3 cm/s. On the basis of these velocities the distance traveled over a 6-week period in either case would only be $\sim 35\%$ of that needed to reach BR2, which exhibits peak advective iron input in June. So while the flowlines indicate that this path could be a direct conduit to the central Arabian Sea, the surface current speeds in the model are clearly insufficient to accomplish this.

[32] The time series and the sequence of ΔAE distributions unequivocally demonstrate that, relative to GISS deposition, the Gulf of Oman receives the greatest additional aeolian enrichment from GOCART deposition (Figures 8e and 10). Moreover, the annual mean distribution of ΔFe between the two solutions reveals that the greatest excess soluble iron in the GOCART solution ($\Delta\text{Fe} > 100 \text{ pM}$) covers the entire region north of 18°N between 61°E and 67°E (data not shown). The magnitude of this excess pool suggests the Omani deposition region as the most likely supplemental source of soluble iron for fueling the offshore SWM-period phytoplankton bloom. However, as already noted, an obvious

advective transport linkage between the May/June maximum in deposition and the bloom's offshore appearance in July is not evident. The flowlines suggest that during the SIM to SWM transition, when ΔAE and the accumulation of this excess in the Gulf of Oman is most acute (Figures 8e and 10), any of this additional soluble iron that is not locally consumed through photosynthesis will be transported southeastward.

[33] To clarify how the differing surface distributions of soluble iron for the two solutions are transported around the Arabian Sea, maps of ΔFe with superimposed flowlines were created (Figure 12). These show that, in addition to a primary ΔFe feature which coincides with the GOCART aeolian enrichment maximum, the April and May distributions of ΔFe show a second prominent feature that is centered around 16°N , 60°E (Figures 12a and 12b). This secondary ΔFe maximum spatially coincides with the northern limb of the April cyclonic feature described earlier. The bimodal nature of the April ΔFe distribution suggests that the southern maximum does not result from the concurrent ΔAE pattern as this decreases monotonically with offshore distance (Figure 10a). Further, the southern ΔFe maximum appears to arise solely from horizontal advection as no contribution by entrainment/mixing is indicated in either model solution (data not shown).

Figure 12

[34] As positive ΔAE along the Arabian Peninsula intensifies from April into June (Figures 8f, 8g, and 10a–10c), the extent and magnitude of positive ΔFe between 12°N and 20°N generally increases (Figures 12a–12c). The one area subjected to the greater aeolian deposition provided by the GOCART boundary condition (Figure 8h) for which ΔFe exhibits only a slight accompanying increase is the highly dispersive coastal upwelling zone west of 58°E (Figures 12b–12d), which indicates that there is little distinction between the subsurface iron fields in the two solutions. The June and July distributions show a tongue of positive ΔFe that extends offshore from the northern coast of Somalia (Figures 12c and 12d), which derives from the elevated aeolian deposition over the Gulf of Aden in the GOCART boundary condition (Figures 10c and 10d). This ΔFe feature does not extend beyond 56°E in July, which is consistent with the earlier estimate of transport distance based on the mean monthly current speed along 13.5°N and provides further evidence that horizontal advection of these western waters does not contribute to supporting the offshore phytoplankton bloom in July (Figure 1a). However, the nearshore feature in the July and August distributions of ΔExp do suggest a link to this ΔFe feature emanating from the Gulf of Aden (Figures 7a, 7b, 12c, and 12d). It can also be seen in the surface Chla distributions that the phytoplankton bloom along the western Arabian Peninsula extends somewhat farther offshore in the GOCART solution (Figure 1). Thus the additional aeolian enrichment over the western Arabian Sea does engender a biogeochemical response, though this is spatially distinct from the prominent offshore phytoplankton bloom.

[35] In the central Arabian Sea where this bloom manifests, significant positive ΔFe is present in June and July (Figures 12c and 12d). The source of this iron remains unclear, as the evidence presented so far indicates that neither eastward transport from waters north of Somalia nor direct southwestward transport from the Gulf of Oman during the SIM are viable pathways. To address this dilemma, the complete annual cycle of ΔFe with corresponding flowlines was examined and every other month, starting with September, has been included here (Figures 12e–12h). These show that by the end of the SWM the elevated ΔFe feature which originally manifests off the coast of Oman has progressed across the Gulf and propagated southward along the coasts of Pakistan and India (Figure 12e). At this time, ΔFe in the core of this feature is still $> 200 \text{ pM}$. By the FIM/NEM transition the surface current field along the Indian coast is directed offshore south of 20°N and the ΔFe feature, though reduced in both spatial extent and magnitude, has begun to advect southwestward (Figure 12f).

[36] As the NEM peaks in January, the ΔFe feature is centered at 18°N , 63°E and is continuing to move offshore (Figure 12g). At this time its magnitude is reduced to ca. 120 pM . By March this ΔFe signature has moved slightly farther to the southwest (17°N , 61°E) and is now located

within FB6. Additionally, its magnitude is unchanged while its spatial extent has increased slightly as local aeolian enrichment is now greater than $5 \text{ nmol Fe m}^{-3} \text{ d}^{-1}$ and steadily increasing (Figures 8f and 12h). During the NEM and SIM, this ΔFe signature is the most prominent over the entire Arabian Sea (Figures 12g and 12h). Thus heading into the SIM-SWM transition the offshore ΔFe feature is already in place and undergoing modest intensification (Figures 12a, 12b, and 12h) as GOCART-derived aeolian enrichment begins a seasonal increase that peaks in May/June (Table 1 and Figure 8). In particular, over FB6 where the ΔFe feature is located, the difference between GOCART and GISS iron enrichment increases continually from January through May (Figure 8f). Therefore the appearance of the prominent SWM-period phytoplankton bloom in the GOCART solution is associated with this offshore accumulation of iron in the model solution. The manifestation of this nutrient pool is the culmination of a yearlong propagation that originates off the east coast of Oman combined with moderate local intensification during the SIM.

4. Discussion

[37] In their status report on the planning of the US JGOFS Arabian Sea program, *Codispoti and Smith* [1994] posed the following questions regarding atmospheric dust deposition: (1) “How important are aeolian fluxes in supporting the productivity of the Arabian Sea?” and (2) “Do these fluxes influence phytoplankton species composition?” These questions have been addressed here by applying a coupled biophysical model to conduct iron fertilization experiments in silico. Two distinct aeolian deposition fields, derived from the GOCART and GISS atmospheric transport models, were applied as surface boundary conditions. Results from these experiments indicate that biogeochemical variability in the Arabian Sea ecosystem is fundamentally impacted by aeolian mineral deposition and is highly sensitive to its spatiotemporal distribution. Over the past decade, a number of in situ iron enrichment experiments have been performed [Boyd, 2002; Coale *et al.*, 1996; de Baar *et al.*, 2005]. These provide useful context for the differences between the two deposition boundary conditions applied in the numerical experiments reported herein. For example, during EisenEx an increase in soluble iron of up to $1 \text{ } \mu\text{mol Fe m}^{-3}$ was observed [cf. de Baar *et al.*, 2005, Figure 20], which is similar to the maximum value of ΔFe ($0.3 \text{ } \mu\text{mol Fe m}^{-3}$) in the model (Figure 12). In addition, the in situ enrichment studies consistently observed a shift in the phytoplankton population toward larger species, a trend that the model also exhibits.

[38] All of the biogeochemical differences that arise between the two model experiments relate directly to the higher soluble iron concentrations that develop when the GOCART boundary condition is applied and the promotion

Figure 12. Monthly difference in surface iron concentration (nmol Fe m^{-3}) between the two deposition boundary conditions (ΔFe , GOCART-GISS) for (a) April, (b) May, (c) June, (d) July, (e) September, (f) November, (g) January, and (h) March. The superimposed contours (dark red lines) are included at $60 \text{ nmol Fe m}^{-3}$ intervals. The superimposed flowlines are as described in Figure 10.

of the larger phytoplankton size class that this engenders. Interestingly, waters upwelled and advected around the anticyclonic Great Whirl (8°N, 53°E) during the SWM have prominent and essentially identical Chla signatures in both solutions as well as no coinciding feature in the ΔExp distributions (Figures 1a, 1b, and 7). Given that these upwelled waters inherently tend toward iron limitation as nutrients are utilized [see Wiggert *et al.*, 2006, Figure 15c] and surface iron enrichment from GOCART deposition is appreciably higher (Figure 8h), it can be concluded that primary production off the Somali coast is not regulated by nutrient availability during the SWM.

[39] The most prominent GOCART-enhanced biological feature is the phytoplankton bloom in the central Arabian Sea in July. There is a counterpart to this bloom in the GISS solution, however it is significantly weaker and shifted to the north. In terms of its location, spatial extent and the contrast in concentration of chlorophyll *a* (Chla) with surrounding waters, the phytoplankton bloom in the GOCART solution exhibits much better agreement with the SeaWiFS observations. Moreover, the monthly time series of surface Chla from the offshore bloom region demonstrate that the GOCART solution more accurately captures both the magnitude and timing of this bloom. At BR2, the modeled bloom has clear differences throughout the euphotic zone as in the GOCART solution it is 10–15 m deeper, exhibits a stronger speciation shift and has export fluxes in July and August that are twice as high. These differences are not the result of local aeolian deposition, which is actually higher in the GISS boundary condition. Rather, they develop in response to a significant horizontally advected injection of iron.

[40] In conjunction with these offshore phytoplankton blooms, export flux maxima appear at the three westward sites (BR2–BR4) in August in both solutions. This timing, as well as the magnitude, is consistent with the observed fluxes of biogenic material in sediment trap time series (Figure 5). It has been suggested that the delay between the onset of monsoon-driven upwelling in May and the observed export maxima relates to a release of grazing pressure. This hypothesized release is attributed to the attendant copepod population (primarily *Calanoides carinatus*) completing its acquisition of the lipid reserves necessary for successful diapause [Smith, 2001]. Although the peak rates of export are clearly amplified in the GOCART case because of the more pronounced phytoplankton speciation shift (Figures 3c and 3d), these export maxima consistently appear in August despite no accounting for mesozooplankton diapause within the ecosystem model's architecture. Moreover, grazing control, exerted by mature *C. carinatus* that emerge directly from diapause within waters upwelled during the SWM [see Idrisi *et al.*, 2004], has been proposed as a likely explanation for why a more prolific phytoplankton bloom does not manifest given the favorable conditions that are in place [Barber *et al.*, 2001]. Our experiments suggest an alternative, or potentially complementary, bloom regulation mechanism. Both solutions demonstrate that nutrient supply to surface waters is accomplished through a combination of entrainment through mixed layer deepening and lateral advection (Figures 3, 4, and 9). Comparison of the two solutions demonstrates that the latter process can provide a supplemental source of iron, transported from high-

deposition regions, that allows a more prominent bloom to develop than would otherwise occur if inherently iron limited upwelled/entrained waters were the sole nutrient source.

[41] On the basis of transport estimates, the seasonal evolution of surface flow lines, and the yearlong propagation of the primary ΔFe feature apparent in Figure 12, heavy aeolian deposition over the Gulf of Oman provides the iron source that supports the offshore bloom's appearance. The supplemental pool of iron that enhances the magnitude of the offshore phytoplankton blooms in the GOCART solution reaches its May location (16°N, 61°E) following a clockwise progression around the northern Arabian Sea that first entails traveling southeast along the coasts of Pakistan and India before heading offshore and moving southwestward in conjunction with NEM onset. In situ observations of current patterns in the Gulf of Oman are scarce; however a section of current vectors across the Gulf's mouth from 13–16 October 1999 [Pous *et al.*, 2004] is consistent with the characteristics exhibited by the model flow lines as the iron feature is driven offshore (Figures 12e and 12f).

[42] The offshore blooms' link to iron-rich waters advected from the Gulf of Oman that is indicated by our results would appear to contradict the conclusions of two recent studies. Both have argued that the nutrients supporting the offshore phytoplankton bloom were originally upwelled along the northern coast of Somalia and subsequently advected offshore by the Somali Current [Kawamiya, 2001; Kumar *et al.*, 2001]. However, neither study seeks to explain the broad-scale phytoplankton distribution in July; both primarily address potential nutrient sources for waters west of this bloom. Kumar *et al.* [2001] focused on eastward advected nutrients (in their analysis nitrate) reaching the central Arabian Sea in August. The solutions presented herein both exhibit peak advection of iron and nitrate (data not shown) at BR4 that occurs in August, so this aspect of the model is consistent with their conclusions and is independent of deposition field. It also indicates that nutrient transport pathways to a given offshore location in the Arabian Sea, especially during the SWM, could consist of a temporally evolving succession whereby inputs from the east could be superseded by western contributions.

[43] In the modeling study noted in the preceding paragraph, an eddy-permitting model was used to identify and contrast the relative contributions of three source regions to 4-month averaged offshore nitrate distributions [Kawamiya, 2001]. This study indicated that waters originating in the Omani upwelling zone could act as a source of nutrients to the more eastward region where the offshore bloom appears in our GOCART-driven solution. Mesoscale processes would provide quicker, more direct transport to the offshore waters where the July bloom manifests but it was concluded that the Omani upwelling zone was a relatively minor source of nutrients [Kawamiya, 2001]. However, because of the extreme regional heterogeneity noted above, the 4-month averages used in this analysis could introduce significant temporal aliasing that would skew these conclusions [Wiggert *et al.*, 1994]. The iron-enriched feature that appears offshore of the Arabian Peninsula just prior to SWM onset in our model results is ideally located to benefit from such mesoscale-mediated transport once the Omani upwelling zone becomes active. Assessing the possibility

that mesoscale processes would more effectively transport iron-enriched waters from the Gulf of Oman into the central Arabian Sea would require further experiments performed at finer spatial resolution. Indeed, it bears reiteration that the application of coupled biophysical models with grid resolution sufficient to explicitly include mesoscale processes is likely the most pressing need for achieving improved emulation of biogeochemical variability in the northern Arabian Sea [Hood *et al.*, 2003; Wiggert *et al.*, 2005].

[44] Over the years, identifying the processes associated with the appearance and maintenance of the offshore SWM phytoplankton bloom has inspired one of the classical debates in the Arabian Sea literature that has recently been reviewed by Wiggert *et al.* [2005]. The clear sensitivity to choice of aeolian boundary condition exhibited in our model results demonstrates that the spatiotemporal distribution of mineral dust deposition helps define the Arabian Sea ecosystem. These new results introduce an additional aspect to the appearance and maintenance of the Arabian Sea's offshore summertime phytoplankton bloom that must also be included when considering the mechanisms that contribute to this feature's appearance.

[45] The advected iron that leads to the more prominent offshore bloom in the GOCART solution also results in export fluxes at BR2 and BR3 that are twice as high. Distributions of ΔExp reveal that this particle flux enhancement is widespread and has two distinct maxima in July and August. The ΔExp maxima offshore of the western Arabian Peninsula is associated with the elevated deposition over the Gulf of Aden, whereas the ΔExp maxima in the central Arabian Sea is associated with iron advected from the Gulf of Oman as described above. Over the entire Arabian Sea, the model results suggest that the additional, GOCART-derived iron results in organic matter export during the SWM that is up to 20% higher and $\sim 12\%$ higher annually. The model results further indicate that the annual enhancement to Arabian Sea export stimulated by the additional iron amounts to 3.5 TgC and that 56% of this export occurs during the July–September time frame. Over the Indian Ocean basin, an additional 4.6 TgC is exported with 54% of this occurring from July through September. Thus enhanced primary production arising from the deposition hot spots over the Gulf of Aden and the Gulf of Oman contributes fully 80% to the additional basinwide, SWM-period export of 2.5 TgC that occurs when the GOCART boundary condition is applied.

[46] Recent application of a bio-optical model to remote sensing data has resulted in the identification of a *Trichodesmium* bloom in February 1998 [see Westberry *et al.*, 2005, Figure 8] that is spatially coincident with the offshore ΔFe maxima in our results. The alignment of these two features is encouraging, given that the high iron requirement of this diazotroph is well documented [e.g., Kustka *et al.*, 2003]. Past analyses have suggested a significant presence of *Trichodesmium* in the central Arabian Sea [Capone *et al.*, 1998], while a previous isotopic study has indicated that upward of 30% of the nitrate pool in surface waters of the northern Arabian Sea is derived from N_2 -fixation [Brandes *et al.*, 1998]. A recent modeling study also indicates elevated rates of N_2 -fixation in the Arabian Sea [Moore *et al.*, 2004]. They report that only $\sim 5\%$ of the additional primary production that occurs globally when diazotrophy

is included results directly from diazotrophic growth; the remainder is associated with the response of their model's other phytoplankton groups to the additional N supply. The general stimulation of the pelagic system by significant additions of new nitrogen derived through N_2 -fixation is a consistent theme for a number of recent N-cycle analyses [Capone *et al.*, 2005; Coles *et al.*, 2004; Hood *et al.*, 2004]. However, observationally based quantification of trophic transfers associated with nutrients supplied via diazotrophy, the nature of their regeneration within the euphotic zone and their eventual contribution to biogenic export flux has yet to be realized [Mulholland, 2007]. Nevertheless, the results presented herein suggest an intriguing linkage between elevated aeolian iron deposition, monsoon-driven currents and significant diazotrophic production that contributes to the appearance of the SWM phytoplankton bloom in the central Arabian Sea.

5. Conclusion

[47] The Arabian Sea ecosystem has conically been regarded as invariably iron replete because of the known magnitude of dust deposition and the region has recently been described as “Mother Nature's iron experiment” [Smith, 2001]. Yet our recently published model results indicate that, at least locally and transiently, iron limitation does occur [Wiggert *et al.*, 2006]. These results are supported by recent observations obtained during a pair of research cruises in the latter part of the 2004 SWM that indicate the occurrence of iron limitation in upwelled waters offshore of Ras al Madrakah along the Omani coast (S. W. A. Naqvi *et al.*, Trace metal deficiency and suboxia limit productivity of upwelled water in the Arabian Sea, manuscript in preparation, 2007). Thus the canonical thinking on soluble iron availability in the Arabian Sea is now being challenged by both our previously published model results and the recent in situ measurements. The new model results presented here suggest a transport pathway from the Gulf of Oman, where excessive aeolian deposition occurs, to the central Arabian Sea that contributes to the SWM-period phytoplankton blooms that develop there. Given the evidence provided within other recent studies it is highly likely that the accumulated iron in our results would lead to an injection of new nitrogen via diazotrophic production. However, since our model does not currently allow for N_2 -fixation, the biogeochemical impacts associated with such a nutrient injection could not be assessed. Further observational surveys with the capacity to appropriately characterize micronutrient fields (esp. bioavailable iron), as well as quantify *Trichodesmium* populations and their associated fixation rates, will be crucial for assessing these implications.

[48] Our results also indicate that dissolved iron concentrations in the Arabian Sea are strongly modulated by the specific spatiotemporal characteristics of aeolian mineral deposition. Thus it is probable that interannual and interdecadal variability in the Arabian Sea ecosystem is fundamentally influenced by climate and land use impacts on the surrounding dust source regions [Mahowald *et al.*, 2006]. Indeed, as indicated by a recent global modeling investigation into the response of air-sea CO_2 exchange to altered atmospheric dust delivery [Dutkiewicz *et al.*, 2006; Parekh

et al., 2006], these factors are likely to make leading-order contributions to defining the magnitude of this region's efflux of CO₂ to the atmosphere [Bates *et al.*, 2006a] through alteration of the Arabian Sea's biological pump and its associated metabolic status (i.e., net autotrophy versus net heterotrophy) that is now believed to switch seasonally [Bates *et al.*, 2006b].

[49] **Acknowledgments.** The NASA Oceanography Program (NAG 58595) supported this research. The authors would like to thank Jim Christian and Raleigh Hood for their invaluable comments and insights. We would also like to thank Paul Ginoux and Ina Tegen for making their aeolian transport products available. The comments of two anonymous reviewers also helped improve this manuscript and are greatly appreciated. The freeware programs Ferret (<http://www.ferret.noaa.gov>) and GMT (<http://gmt.soest.hawaii.edu>) were extensively used as analysis and graphics tools. This is U.S. JGOFS contribution 1060.

References

- Antonov, J., S. Levitus, T. P. Boyer, M. Conkright, T. O'Brien, C. Stephens, and B. Trotsenko (1998), *World Ocean Atlas 1998*, vol. 3, *Temperature of the Indian Ocean*, 166 pp, U. S. Gov. Print. Off., Washington, D. C.
- Barber, R. T., J. Marra, R. C. Bidigare, L. A. Codispoti, D. Halpern, Z. Johnson, M. Latasa, R. Goericke, and S. L. Smith (2001), Primary productivity and its regulation in the Arabian Sea during 1995, *Deep Sea Res., Part II*, 48, 1127–1172.
- Bates, N. R., A. C. Pequignot, and C. L. Sabine (2006a), Ocean carbon cycling in the Indian Ocean: 1. Spatiotemporal variability of inorganic carbon and air-sea CO₂ gas exchange, *Global Biogeochem. Cycles*, 20, GB3020, doi:10.1029/2005GB002491.
- Bates, N. R., A. C. Pequignot, and C. L. Sabine (2006b), Ocean carbon cycling in the Indian Ocean: 2. Estimates of net community production, *Global Biogeochem. Cycles*, 20, GB3021, doi:10.1029/2005GB002492.
- Bauer, S., G. L. Hitchcock, and D. B. Olson (1991), Influence of monsoonally-forced Ekman dynamics upon surface layer depth and plankton biomass distribution in the Arabian Sea, *Deep Sea Res., Part I*, 38, 531–553.
- Bishop, J. K. B., R. E. Davis, and J. T. Sherman (2002), Robotic observations of dust storm enhancement of carbon biomass in the North Pacific, *Science*, 298, 817–821.
- Boyd, P. W. (2002), The role of iron in the biogeochemistry of the Southern Ocean and equatorial Pacific: A comparison of in situ iron enrichments, *Deep Sea Res., Part II*, 49, 1803–1821.
- Boyer, T. P., S. Levitus, J. Antonov, M. Conkright, T. O'Brien, C. Stephens, and B. Trotsenko (1998), *World Ocean Atlas 1998*, vol. 6, *Salinity of the Indian Ocean*, 166 pp., U. S. Gov. Print. Off., Washington, D. C.
- Brandes, J. A., A. H. Devol, T. Yoshinari, D. A. Jayakumar, and S. W. A. Naqvi (1998), Isotopic composition of nitrate in the central Arabian Sea and eastern tropical North Pacific: A tracer for mixing and nitrogen cycles, *Limnol. Oceanogr.*, 43, 1680–1689.
- Brock, J. C., and C. R. McClain (1992), Interannual variability in phytoplankton blooms observed in the northwestern Arabian Sea during the southwest monsoon, *J. Geophys. Res.*, 97, 733–750.
- Brock, J. C., C. R. McClain, M. E. Luther, and W. W. Hay (1991), The phytoplankton bloom in the northwest Arabian Sea during the southwest monsoon of 1979, *J. Geophys. Res.*, 96, 20,623–20,642.
- Capone, D. G., A. Subramaniam, J. P. Montoya, M. Voss, C. Humborg, A. M. Johansen, R. L. Siebert, and E. J. Carpenter (1998), An extensive bloom of the N-2-fixing cyanobacterium *Trichodesmium erythraeum* in the central Arabian Sea, *Mar. Ecol. Prog. Ser.*, 172, 281–292.
- Capone, D. G., J. A. Burns, J. P. Montoya, A. Subramaniam, C. Mahaffey, T. Gunderson, A. F. Michaels, and E. J. Carpenter (2005), Nitrogen fixation by *Trichodesmium* spp.: An important source of new nitrogen to the tropical and subtropical North Atlantic Ocean, *Global Biogeochem. Cycles*, 19, GB2024, doi:10.1029/2004GB002331.
- Chen, D., A. Busalacchi, and L. Rothstein (1994a), The roles of vertical mixing, solar radiation, and wind stress in a model simulation of the sea surface temperature seasonal cycle in the tropical Pacific Ocean, *J. Geophys. Res.*, 99, 20,345–20,359.
- Chen, D., L. M. Rothstein, and A. J. Busalacchi (1994b), A hybrid vertical mixing scheme and its application to tropical ocean models, *J. Phys. Oceanogr.*, 24, 2156–2179.
- Chin, M., P. Ginoux, S. Kinne, O. Torres, B. N. Holben, B. N. Duncan, R. V. Martin, J. A. Logan, A. Higurashi, and T. Nakajima (2002), Tropospheric aerosol optical thickness from the GOCART model and comparisons with satellite and Sun photometer measurements, *J. Atmos. Sci.*, 59, 461–483.
- Christian, J. R., M. A. Verschell, R. Murtugudde, A. J. Busalacchi, and C. R. McClain (2002a), Biogeochemical modelling of the tropical Pacific Ocean. 1: Seasonal and interannual variability, *Deep Sea Res., Part II*, 49, 509–543.
- Christian, J. R., M. A. Verschell, R. Murtugudde, A. J. Busalacchi, and C. R. McClain (2002b), Biogeochemical modelling of the tropical Pacific Ocean. 2: Iron biogeochemistry, *Deep Sea Res., Part II*, 49, 545–565.
- Coale, K. H., et al. (1996), A massive phytoplankton bloom induced by an ecosystem-scale iron fertilization experiment in the equatorial Pacific Ocean, *Nature*, 383, 495–501.
- Codispoti, L. A., and S. L. Smith (1994), The U. S. JGOFS Arabian Sea research program, in *Biogeochemical Processes in the Arabian Sea*, edited by J. C. Goldman and H. D. Livingston, pp. 35–46, Mar. Hydrophys. Inst., Natl. Ukrainian Acad. of Sci., Sevastopol.
- Coles, V. J., R. R. Hood, M. Pascual, and D. G. Capone (2004), Modeling the impact of Trichodesmium and nitrogen fixation in the Atlantic Ocean, *J. Geophys. Res.*, 109, C06007, doi:10.1029/2002JC001754.
- Conkright, M., T. O'Brien, S. Levitus, T. P. Boyer, J. Antonov, and C. Stephens (1998), *World Ocean Atlas 1998*, vol. 12, *Nutrients and Chlorophyll of the Indian Ocean*, 217 pp, U. S. Gov. Print. Off., Washington, D. C.
- de Baar, H. J. W., et al. (2005), Synthesis of iron fertilization experiments: From the iron age in the age of enlightenment, *J. Geophys. Res.*, 110, C09S16, doi:10.1029/2004JC002601.
- Dutkiewicz, S., M. J. Follows, P. Heimbach, and J. Marshall (2006), Controls on ocean productivity and air-sea carbon flux: An adjoint model sensitivity study, *Geophys. Res. Lett.*, 33, L02603, doi:10.1029/2005GL024987.
- Geider, R. J., H. L. MacIntyre, and T. M. Kana (1996), A dynamic model of photoadaptation in phytoplankton, *Limnol. Oceanogr.*, 41, 1–15.
- Ginoux, P., M. Chin, I. Tegen, J. M. Prospero, B. Holben, O. Dubovik, and S. J. Lin (2001), Sources and distributions of dust aerosols simulated with the GOCART model, *J. Geophys. Res.*, 106, 20,255–20,273.
- Hitchcock, G. L., E. L. Key, and J. Masters (2000), The fate of upwelled waters in the Great Whirl, August 1995, *Deep Sea Res., Part II*, 47, 1605–1621.
- Honjo, S., J. Dymond, W. Prell, and V. Ittekkot (1999), Monsoon-controlled export fluxes to the interior of the Arabian Sea, *Deep Sea Res., Part II*, 46, 1859–1902.
- Hood, R. R., K. E. Kohler, J. P. McCreary, and S. L. Smith (2003), A four-dimensional validation of a coupled physical-biological model of the Arabian Sea, *Deep Sea Res., Part II*, 50, 2917–2945.
- Hood, R. R., V. J. Coles, and D. G. Capone (2004), Modeling the distribution of Trichodesmium and nitrogen fixation in the Atlantic Ocean, *J. Geophys. Res.*, 109, C06006, doi:10.1029/2002JC001753.
- Hutchins, D. A., G. R. DiTullio, Y. Zhang, and K. W. Bruland (1998), An iron limitation mosaic in the California upwelling regime, *Limnol. Oceanogr.*, 43, 1037–1054.
- Hutchins, D. A., et al. (2002), Phytoplankton iron limitation in the Humboldt Current and Peru Upwelling, *Limnol. Oceanogr.*, 47, 997–1011.
- Idrisi, N., M. J. Olascoaga, Z. Garraffo, D. B. Olson, and S. L. Smith (2004), Mechanisms for emergence from diapause of *Calanoides carinatus* in the Somali Current, *Limnol. Oceanogr.*, 49, 1262–1268.
- Jickells, T. D., et al. (2005), Global iron connections between desert dust, ocean biogeochemistry, and climate, *Science*, 308, 67–71.
- Johnson, K. S., et al. (2003), Surface ocean-lower atmosphere interactions in the northeast Pacific Ocean gyre: Aerosols, iron, and the ecosystem response, *Global Biogeochem. Cycles*, 17(2), 1063, doi:10.1029/2002GB002004.
- Kalnay, E., et al. (1996), The NCEP/NCAR 40-year reanalysis project, *Bull. Amer. Met. Soc.*, 77, 437–471.
- Kawamiya, M. (2001), Mechanism of offshore nutrient supply in the western Arabian Sea, *J. Mar. Res.*, 59, 675–696.
- Keen, T. R., J. C. Kindle, and D. K. Young (1997), The interaction of southwest monsoon upwelling, advection and primary production in the northwest Arabian Sea, *J. Mar. Syst.*, 13, 61–82.
- Kraus, E. B., and J. S. Turner (1967), A one-dimensional model of the seasonal thermocline: 2, The general theory and its consequences, *Tellus*, 19, 98–105.
- Kumar, S. P., M. Madhupratap, M. D. Kumar, P. M. Muraleedharan, S. N. de Souza, M. Gauns, and V. Sarma (2001), High biological productivity in the central Arabian Sea during the summer monsoon driven by Ekman pumping and lateral advection, *Curr. Sci.*, 81, 1633–1638.
- Kustka, A. B., S. A. Sañudo-Wilhelmy, E. J. Carpenter, D. Capone, J. Burns, and W. G. Sunda (2003), Iron requirements for dinitrogen- and ammonium-supported growth in cultures of *Trichodesmium* (IMS 101): Comparison with nitrogen fixation rates and iron: Carbon ratios of field populations, *Limnol. Oceanogr.*, 48, 1869–1884.
- Latasa, M., and R. R. Bidigare (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 Res., Part II*, 45, 2133–2170.
- Léon, J. F., and M. Legrand (2003), Mineral dust sources in the surroundings of the north Indian Ocean, *Geophys. Res. Lett.*, 30(6), 1309, doi:10.1029/2002GL016690.
- Lierheimer, L. J., and K. Banse (2002), Seasonal and interannual variability of phytoplankton pigment in the Laccadive (Lakshadweep) Sea as observed by the Coastal Zone Color Scanner, *Proc. Indian Acad. Sci. Earth Planet. Sci.*, 111, 163–185.
- Mahowald, N. M., D. Muhs, S. Levis, P. J. Rasch, M. Yoshioka, C. S. Zender, and C. Luo (2006), Change in atmospheric mineral aerosols in response to climate: Last glacial period, pre-industrial, modern and doubled-carbon dioxide climates, *J. Geophys. Res.*, 111, D10202, doi:10.1029/2005JD006653.
- Martin, J. H., G. A. Knauer, D. M. Karl, and W. W. Broenkow (1987), VERTEX: Carbon cycling in the northeast Pacific, *Deep Sea Res., Part I*, 34, 267–285.
- Martin, J. H., R. M. Gordon, and S. F. Fitzwater (1991), The case for iron, *Limnol. Oceanogr.*, 36, 1793–1803.
- McCreary, J. P., K. E. Kohler, R. R. Hood, and D. B. Olson (1996), A four-component model of biological activity in the Arabian Sea, *Prog. Oceanogr.*, 37, 193–240.
- Measures, C. I., and S. Vink (1999), Seasonal variations in the distribution of Fe and Al in the surface waters of the Arabian Sea, *Deep Sea Res., Part II*, 46, 1597–1622.
- Moore, J. K., S. C. Doney, and K. Lindsay (2004), Upper ocean ecosystem dynamics and iron cycling in a global three-dimensional model, *Global Biogeochem. Cycles*, 18, GB4028, doi:10.1029/2004GB002220.
- Mulholland, M. R. (2007), The fate of nitrogen fixed by diazotrophs in the ocean, *Biogeosci.*, 4, 37–51.
- Murtugudde, R., and A. J. Busalacchi (1999), Interannual variability of the dynamics and thermodynamics of the tropical Indian Ocean, *J. Clim.*, 12, 2300–2326.
- Murtugudde, R., R. Seager, and A. Busalacchi (1996), Simulation of the tropical oceans with an ocean GCM coupled to an atmospheric mixed-layer model, *J. Clim.*, 9, 1795–1815.
- Murtugudde, R., J. P. McCreary, and A. J. Busalacchi (2000), Oceanic processes associated with anomalous events in the Indian Ocean with relevance to 1997–1998, *J. Geophys. Res.*, 105, 3295–3306.
- Neuer, S., M. E. Torres-Padrón, M. D. Gelado-Caballero, M. J. Rueda, J. Hernández-Brito, R. Davenport, and G. Wefer (2004), Dust deposition pulses to the eastern subtropical North Atlantic gyre: Does ocean's biogeochemistry respond?, *Global Biogeochem. Cycles*, 18, GB4020, doi:10.1029/2004GB002228.
- Parekh, P., M. J. Follows, and E. Boyle (2004), Modeling the global ocean iron cycle, *Global Biogeochem. Cycles*, 18, GB1002, doi:10.1029/2003GB002061.
- Parekh, P., S. Dutkiewicz, M. J. Follows, and T. Ito (2006), Atmospheric carbon dioxide in a less dusty world, *Geophys. Res. Lett.*, 33, L03610, doi:10.1029/2005GL025098.
- Pease, P. P., V. P. Tchakerian, and N. W. Tindale (1998), Aerosols over the Arabian Sea: Geochemistry and source areas for aeolian desert dust, *J. Arid Environ.*, 39, 477–496.
- Pous, S. P., X. Carton, and P. Lazure (2004), Hydrology and circulation in the Strait of Hormuz and the Gulf of Oman: Results from the GOGP99 experiment: 2. Gulf of Oman, *J. Geophys. Res.*, 109, C12038, doi:10.1029/2003JC002146.
- Price, J. F., R. Weller, and R. Pinkel (1986), Diurnal cycling: Observations and models of the upper ocean response to diurnal heating, cooling and wind mixing, *J. Geophys. Res.*, 91, 8411–8427.
- Prospero, J. M., P. Ginoux, O. Torres, S. E. Nicholson, and T. E. Gill (2002), Environmental characterization of global sources of atmospheric soil dust identified with the NIMBUS 7 Total Ozone Mapping Spectrometer (TOMS) absorbing aerosol product, *Rev. Geophys.*, 40(1), 1002, doi:10.1029/2000RG000095.
- Rixen, T., M. V. S. Gupta, and V. Ittekkot (2005), Deep ocean fluxes and their link to surface ocean processes and the biological pump, *Prog. Oceanogr.*, 65, 240–259.
- Saager, P. M., H. J. W. Debaar, and P. H. Burkill (1989), Manganese and iron in Indian Ocean waters, *Geochim. Cosmochim. Acta*, 53, 2259–2267.
- Seager, R., M. Blumenthal, and Y. Kushnir (1995), An advective atmospheric mixed layer model for ocean modeling purposes: Global simulation of surface heat fluxes, *J. Clim.*, 8, 1951–1964.
- Smith, S. L. (2001), Understanding the Arabian Sea: Reflections on the 1994–1996 Arabian Sea expedition, *Deep Sea Res., Part II*, 48, 1385–1402.
- Sunda, W. G., and S. A. Huntsman (1997), Interrelated influence of iron, light and cell size on marine phytoplankton growth, *Nature*, 390, 389–392.
- Takeda, S., A. Kamatani, and K. Kawanobe (1995), Effects of nitrogen and iron enrichments on phytoplankton communities in the northwestern Indian Ocean, *Mar. Chem.*, 50, 229–241.
- Tegen, I. (2003), Modeling the mineral dust aerosol cycle in the climate system, *Quat. Sci. Rev.*, 22, 1821–1834.
- Tegen, I., and I. Fung (1994), Modeling of mineral dust in the atmosphere: Sources, transport, and optical thickness, *J. Geophys. Res.*, 99, 22,897–22,914.
- Tindale, N. W., and P. P. Pease (1999), Aerosols over the Arabian Sea: Atmospheric transport pathways and concentrations of dust and sea salt, *Deep Sea Res., Part II*, 46, 1577–1595.
- Veldhuis, M. J. W., G. W. Kraay, J. D. L. VanBleijswijk, and M. A. Baars (1997), Seasonal and spatial variability in phytoplankton biomass, productivity and growth in the northwestern Indian Ocean: The southwest and northeast monsoon, 1992–1993, *Deep Sea Res., Part I*, 44, 425–449.
- Weber, L., C. Volker, M. Schartau, and D. A. Wolf-Gladrow (2005), Modeling the speciation and biogeochemistry of iron at the Bermuda Atlantic Time-series Study site, *Global Biogeochem. Cycles*, 19, GB1019, doi:10.1029/2004GB002340.
- Wells, M. L. (2003), The level of iron enrichment required to initiate diatom blooms in HNLC waters, *Mar. Chem.*, 82, 101–114.
- Westberry, T. K., D. A. Siegel, and A. Subramaniam (2005), An improved bio-optical model for the remote sensing of *Trichodesmium* spp. blooms, *J. Geophys. Res.*, 110, C06012, doi:10.1029/2004JC002517.
- Wiggert, J., T. Dickey, and T. Granata (1994), The effect of temporal under-sampling on primary production estimates, *J. Geophys. Res.*, 99, 3361–3371.
- Wiggert, J. D., R. R. Hood, K. Banse, and J. C. Kindle (2005), Monsoon-driven biogeochemical processes in the Arabian Sea, *Prog. Oceanogr.*, 65, 176–213.
- Wiggert, J. D., R. G. Murtugudde, and J. R. Christian (2006), Annual ecosystem variability in the tropical Indian Ocean: Results of a coupled bio-physical ocean general circulation model, *Deep Sea Res., Part II*, 53, 644–676.
- Wu, J. F., and E. Boyle (2002), Iron in the Sargasso Sea: Implications for the processes controlling dissolved Fe distribution in the ocean, *Global Biogeochem. Cycles*, 16(4), 1086, doi:10.1029/2001GB001453.
- Young, D. K., and J. C. Kindle (1994), Physical processes affecting availability of dissolved silicate for diatom production in the Arabian Sea, *J. Geophys. Res.*, 99, 22,619–22,632.

R. G. Murtugudde, Earth System Science Interdisciplinary Center, University of Maryland, College Park, MD 20742-2465, USA.
J. D. Wiggert, Center for Coastal Physical Oceanography, Old Dominion University, Norfolk, VA 23508-2026, USA. (jwiggert@ccpo.odu.edu)