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The role of seasonal and diel changes in mixed-layer depth on carbon and chlorophyll distributions in the Arabian Sea

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Abstract

The effects of changes in the mixed-layer depth on the distribution of particulate organic carbon (POC) and chlorophyll *a* were examined in the Arabian Sea during the Northeast Monsoon (January and December), Spring Intermonsoon (February–March), and Southwest Monsoon (July and August) of 1995. POC distributions were derived from profiles of beam attenuation calibrated with POC, and chlorophyll *a* distributions were derived from calibrated fluorescence profiles. Depth of the seasonal mixed layer ($\Delta\sigma$ of 0.125 kg m^{-3} from surface density) increased with distance offshore during both monsoons, especially in the southern Arabian Sea where the range was 10–80 m nearshore to 80–120 m offshore. The deepest seasonal mixed layers occurred during the Northeast Monsoon. During the Spring Intermonsoon the seasonal mixed layer was only 10–40 m. Variations in the depth of the diel mixed layer ($\Delta\sigma$ of 0.03 kg m^{-3} from surface density) were up to 90 m during the Northeast Monsoon, but were seldom over 20 m during the Southwest Monsoon. During the Spring Intermonsoon when mixed layers and diel variations in the mixed layer were small, nutrients became depleted, producing oligotrophic conditions plus a strong deep chlorophyll *a* maximum (>2 mg chl m^{-3}) below the mixed layer. The chlorophyll *a* maximum was centered at ~ 50 m, which is significantly beneath the effective depth of satellite color sensing. When mixing is active throughout the diel cycle, particulate organic carbon (POC) and chlorophyll distributions are quite uniform within the mixed layer. Nighttime increases in mixed layer depths can mix POC and chlorophyll *a* produced during the day downward and can entrain new nutrients to enhance primary production. Although mixing from diel variations may be effective in redistributing components within the mixed layer and may be an important mechanism for removing particles from the mixed layer, regional upwelling of nutrients and diatom blooms

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appear to dominate over diel mixed layer dynamics in the production and export of carbon in the Arabian Sea.

1. Introduction

Changes in the depth of mixing at the ocean surface will influence the distribution of particulate and dissolved components whenever there is a gradient in those components over the maximum depth of mixing. Atmospheric and oceanographic forcing of the mixed layer have been observed and modeled, with primary emphasis on physical transport of heat and energy (Imberger, 1985; Weller et al., 1985; Price et al., 1986; Moum et al., 1989; Brainerd and Gregg, 1993,1995). Seasonal variations in wind patterns can affect the mixed layer depth (MLD) as a result of Ekman dynamics. On a time scale relevant to phytoplankton, daily changes in the MLD result from daytime heating of the ocean surface and cooling at night when water temperature exceeds air temperature. This generally leads to stratification and shallower surface mixed layers during the day, and convective cooling and deeper mixed layers at night. Variations in the wind strength also influence day–night mixing. The effects of diel changes in the MLD on the life cycle of plankton in the ocean has been discussed and modeled (Woods and Onken, 1982; Denman and Marra, 1986; Vaulot et al., 1995; Farmer and McNeil, 1997).

The thickness of the mixed layer is usually defined as the depth to which some property is "uniform" – usually temperature or density. Operationally this is the depth at which some property has changed by a specified amount from the surface-water values. Temperature and salinity are usually used to detect the zones of mixing because they are conservative properties that can be measured continuously in situ using a CTD. Discrete bottle samples of properties cannot give resolution comparable to CTD temperature or salinity to define mixing depths. However, the overall abundance and distribution of POC or particulate matter (PM) can be monitored with high spatial and temporal resolution by measuring inherent optical properties of seawater such as beam attenuation using a transmissometer calibrated with discrete POC, PM or plankton samples (Gardner et al., 1993; Chung et al., 1996; Gundersen et al., 1998). Fluorometers calibrated with discrete samples of chlorophyll *a* can be used to estimate the distribution of chlorophyll *a* throughout the water column as an indicator of phytoplankton abundance (Broenkow et al., 1985). POC and chlorophyll *a* are not conservative, but over short time scales their distribution is useful in connecting the biological and physical forcing in a given region.

The mean large-scale distribution of POC and chlorophyll *a* in the Arabian Sea was discussed by Gundersen et al. (1998). The purpose of this paper is to examine the role of changes in the mixed layer depth on seasonal and diel time scales on variations in POC and chlorophyll *a* abundance in the Arabian Sea during different monsoon seasons to assess the potential impact on particle distributions and export.

The Arabian Sea has a number of characteristics that make it an important area to study. There is a large seasonal reversal of atmospheric and surface water circulation (monsoons) that leads to large seasonal variations in biogeochemical processes in the euphotic zone. Some of these changes were assumed to be related to the large seasonal changes in the depth of the mixed layer that resulted from open-ocean upwelling (Bauer et al., 1991; Banse, 1987,1994). Joseph et al. (1990) observed diel changes in the MLD in a region of the Arabian Sea south of our study area, suggesting the possibility that diel time scales might be important to biogeochemical processes in the area. Sediment trap data have demonstrated large changes in particle fluxes that are correlated with monsoonal seasons (Nair et al., 1989; Haake et al., 1993; Lee et al., 1998).

2. Methods

The US portion of the Joint Global Ocean Flux Study (JGOFS) Arabian Sea Process Study was carried out during October 1994–January 1996 (Smith et al., 1998). A 2000 km cruise track that included approximately 30 standard stations was repeatedly occupied by scientists on the R/V *Thomas Thompson* (Fig. 1). Six of the stations were occupied for two days to allow observations over two diel cycles. Hydrographic data were collected using a SeaBird SBE 911$^+$ CTD (Morrison et al., 1998; Morrison, 1997). Mixed-layer depths were calculated from CTD data on each cast using two density criteria; (1) a 0.03 kg m^{-3} density increase from the surface density, and (2) a 0.125 kg m^{-3} density increase. The latter value is the interval routinely used by Levitus (1982) and usually extends down to the top of the seasonal thermocline (Fig. 2). The former value was what was used in the Equatorial Pacific Ocean to identify the depth of daily active mixing (Gardner et al., 1995). The depth of the first CTD value recorded was usually between 2 and 6 m, and changes in density were calculated from that depth. In addition MLDs were calculated based on two temperature standards: (1) a 0.1° temperature change from the surface temperature, which roughly corresponds to a 0.03 kg m^{-3} density change; and (2) a 0.5° change in temperature, which is approximately equivalent to a 0.125 kg m^{-3} density change, assuming in both cases that salinity is constant.

Distributions of particulate organic carbon (POC) and particulate matter (PM) were determined during each cruise from the measurement of light attenuation due to particles (c_p) using SeaTech transmissometers ($\lambda = 660$ nm). Beam c_p was correlated with POC and PM concentrations determined by filtration of water samples collected from the rosette either by our group (TAMU), participants from the groups of Ducklow (VIMS) and Azam (SIO), or Omani scientists. Chlorophyll *a* concentrations were determined using a SeaTech fluorometer calibrated with chlorophyll *a* samples from Niskin bottles analyzed fluorometrically by participants from the groups of Marra (LDEO), Barber (Duke) Bidigare (Hawaii) or Omani scientists. The transmissometer and fluorometer were interfaced with the CTD to obtain continuous profiles at much higher spatial and temporal resolution than is possible with discrete sampling. Gundersen et al. (1998) provided a detailed account of the methods used and a discussion of the regional and seasonal distributions of POC and chlorophyll *a*.

Data were analyzed from five cruises (designated TN0_ _) during different segments of the monsoonal seasons. Seasons were delineated based on an analysis of surface

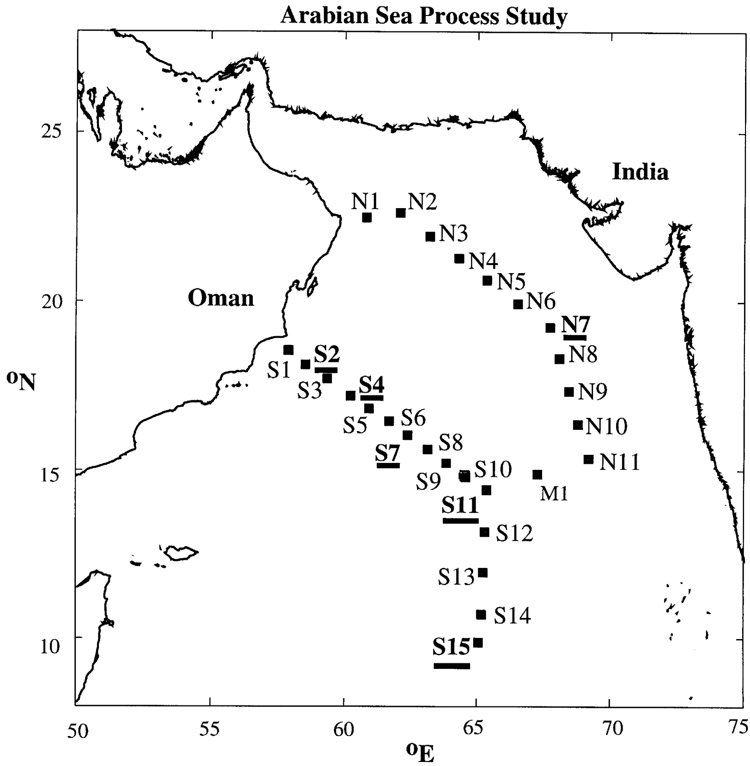


Fig. 1. Stations occupied during the Arabian Sea Expedition. Although stations are labeled for the Northern (N) and Southern (S) transects, the order in which stations were occupied always started along the northern transect moving southeastward and then proceeded northwest along the southern transect in reverse numerical order. Stations underlined were occupied for 2 d during each cruise.

winds measured at the meteorological buoy at 17°N, 60°E (Weller et al., 1998) and the cruise nomenclature used for the project is:
TN043 – Late Northeast Monsoon (LNEM)
TN045 – Spring Intermonsoon (SI)
TN049 – Mid Southwest Monsoon (MSWM)
TN050 – Late Southwest Monsoon (LSWM)
TN054 – Early Northeast Monsoon (ENEM)
Wind and PAR data also were collected via the data acquisition system (DAS) on the R/V *Thompson*. Data reported here reside in the JGOFS data base, which can be accessed via the World Wide Web URL http://www1.whoi.edu/JGOFS.html.

3. Results

The hydrography of the Indian Ocean, including the Arabian Sea, is thoroughly synthesized in the oceanographic atlas of the Indian Ocean Expedition (Wyrtki, 1971).

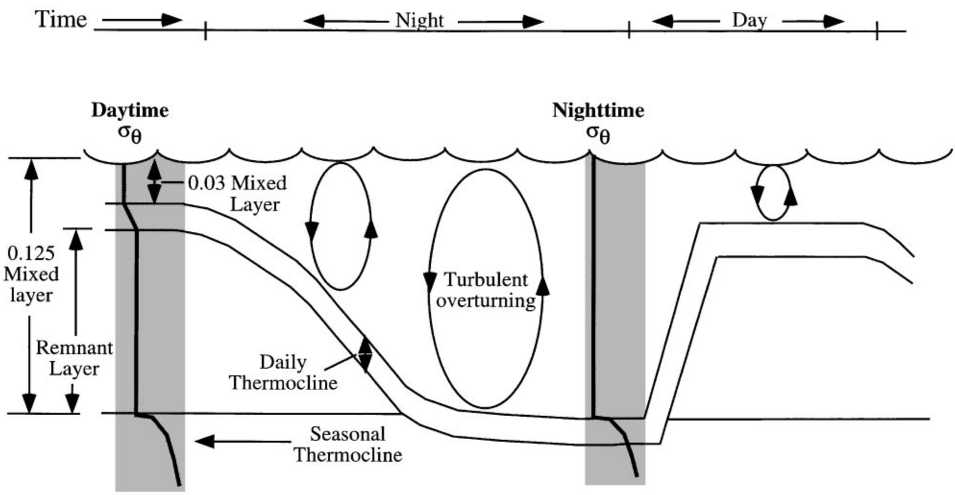


Fig. 2. Daytime heating can stratify and thin the mixed layer, which is identified by a small $\Delta\sigma$ from the surface value – 0.03 kg m^{-3} in this paper. Nighttime cooling and wind mixing can deepen the mixed layer down to the top of the seasonal thermocline, identified by a $\Delta\sigma$ of 0.125 kg m^{-3} (modified from Brainerd and Gregg, 1995).

The hydrographic measurements made during this JGOFS program matched these historical data well (Morrison et al., 1998). Lower temperatures nearshore in both the northern and southern transects during the Southwest Monsoon suggested strong coastal upwelling during that time (Morrison et al., 1998; Gundersen et al., 1998). Offshore of the upwelling areas the water was generally coolest in January (LNEM) along both transects and warmest during the SWM along the Northern transect and during the SI along the Southern Transect.

The time boundaries of the monsoon periods were defined based on atmospheric conditions (Weller et al., 1998; Smith et al., 1998) at the central mooring site (near Sta. S7; Fig. 1). Winds modeled by Kindle and Rochford (NRL) at station S7 (Fig. 3) show the monsoonal directional shifts along with the periods of the monsoons and the cruise intervals. Obviously winds varied somewhat spatially around the basin, so the winds were not identical everywhere to conditions at the central mooring; however, the general monsoon pattern was very similar throughout the basin. Later we will discuss evidence of a lag between the atmospheric conditions and the oceanic physical and biogeochemical response, but no quantitative analysis was made.

Sometimes MLDs must be calculated based on changes in temperature rather than density if salinity measurements are not available, as sometimes occurs with moored instruments (Rudnick et al., 1997). Temperature-based MLDs from CTD data were compared to density-based MLDs and were found to agree to within 2 m 75% of the time. However, differences of up to 10 s of meters occurred at other times when salinity was not constant with depth. A subsurface salinity maximum was sometimes associated with a slight increase in temperature, which almost always resulted in deeper MLDs based on the temperature criteria rather than density. In some parts of

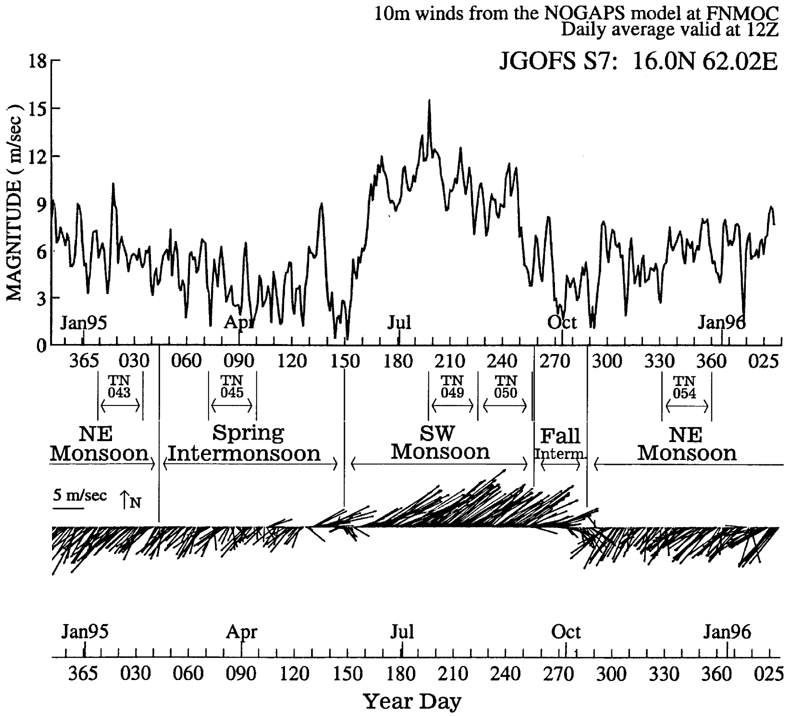


Fig. 3. Magnitude and direction of 10 m winds at station S7 from the NOGAPS model at FNMOC (modified from John Kindle and Peter Rochford, NRL). Time spans of each cruise discussed in this paper and the monsoon periods are indicated.

the Arabian Sea there is a subsurface salinity maximum that is a large-scale advected feature (Morrison et al., 1998). In addition, low air temperatures and low relative humidity during the winter cause extensive evaporation (Wiggert et al., 1999), which creates cool, salty surface water that significantly affects the MLD when calculated based on temperature rather than density. Since we are looking for the effects of mixing on plankton, etc., if temperature and salinity are not uniform, we can not expect plankton or nutrients to be well-mixed. Although none of the temperature or density criteria accurately identify the depth of active mixing all the time, the MLDs based on a density change of $0.03\ \mathrm{kg\,m^{-3}}$ appear most frequently to be the most accurate measurement based on visual identification of changes in slope in density and temperature profiles.

3.1. Maximum mixed layer depth

The maximum MLDs generally increased with distance offshore (except during the SI), but beyond 500 km offshore, there were less consistent trends, except that the MLD continued to increase in depth to the south during the SWM (Fig. 4). The

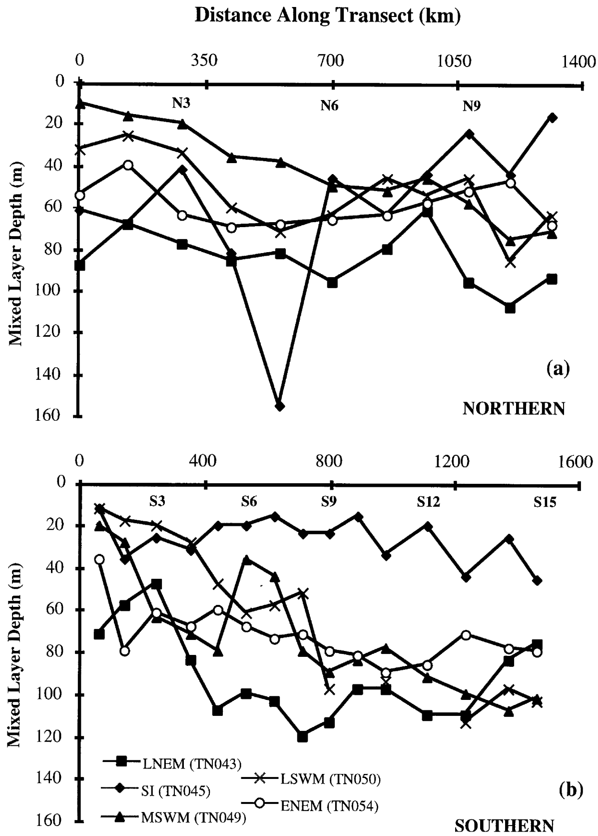


Fig. 4. Maximum MLD at each station based on a $\Delta\sigma$ of 0.125 kg m^{-3} for the (a) Northern transect and (b) Southern transect during each process cruise.

deepest MLDs usually occurred during the LNEM, and the shallowest MLDs usually occurred during the SI. The differences were most marked along the southern transect where > 400 km offshore MLDs were 15–40 m during the SI and 90–120 during LNEM (Fig. 4). The cool surface temperatures offshore during the LNEM were consistent with net heat loss in January (Weller et al., 1998) and high winds that caused deep mixing. The Spring Intermonsoon season was characterized by low winds (Weller et al., 1998; Fig. 3) and moderate solar radiation, leading to thermal stratification and the warmest surface temperatures of the year along the southern transect. The cooler waters and deeper mixed layers along the northern transect during the SI cruise suggest a lag between the SI atmospheric conditions and the hydrographic conditions, which were closer to NEM conditions.

The deepest MLDs generally occurred along the southern transect during the Northeast and Southwest Monsoons, when winds had been blowing strongly in one direction for one to two months (Fig. 3). Deep MLDs southeast of the Findlater Jet result from Ekman pumping and vertical mixing by the strong winds (Bauer et al.,

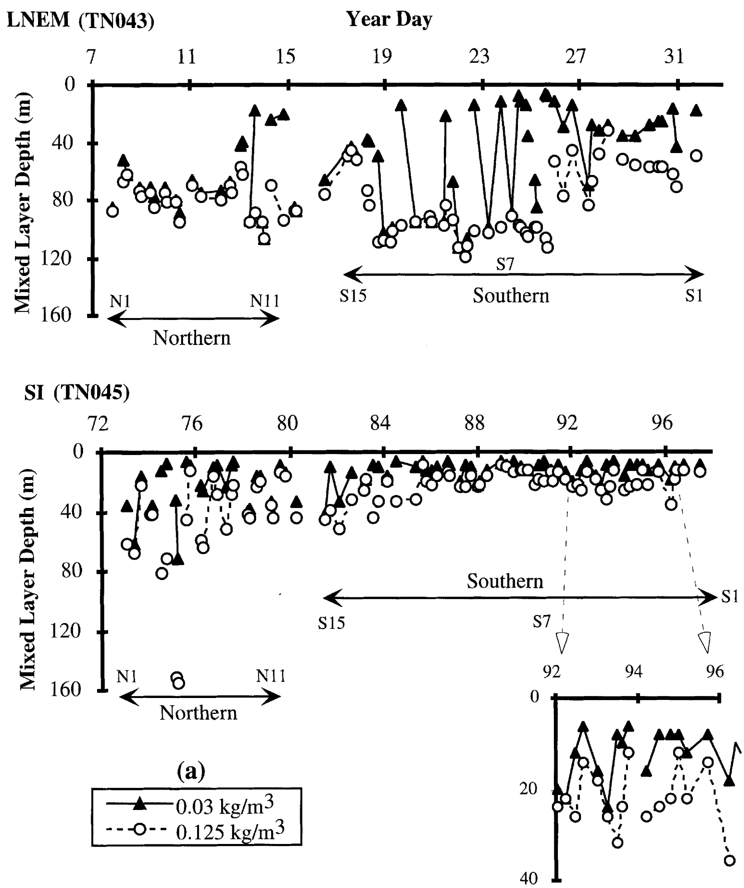


Fig. 5. MLD as determined by a $\Delta\sigma$ of 0.03 and 0.125 kg m^{-3} from surface density for cruises (a) LNEM and SI and (b) MSWM, LSWM and ENEM. Solid and dotted lines connect the data from all casts at a single station for each density interval. The inset on SI (TN045) is an expansion of a portion of the southern transect to show detail. See Fig. 1 for station locations.

1991), while deep MLDs during the LNEM result more from convective winter cooling as well as strong NE winds. The MLDs during this period were deeper than predicted by Bauer et al. (1991). Mixed-layer depths at any one station were quite variable throughout the year, with changes ranging from 20 m at station N8 to >100 m at N5 and S8 (Fig. 4). Overall the SWM displayed the greatest vertical variation with MLDs ranging from 10 to 110 m.

Data showing an exceptionally deep MLD (160 m on multiple casts) were examined and confirmed at station N5 during the SI. Beam attenuation remained high to a much greater depth than normal, suggesting that the anomalous 160 m layer resulted from local downwelling, probably because of a mesoscale feature. However, there was a slight increase in density at about 24 m associated with a slight decrease in

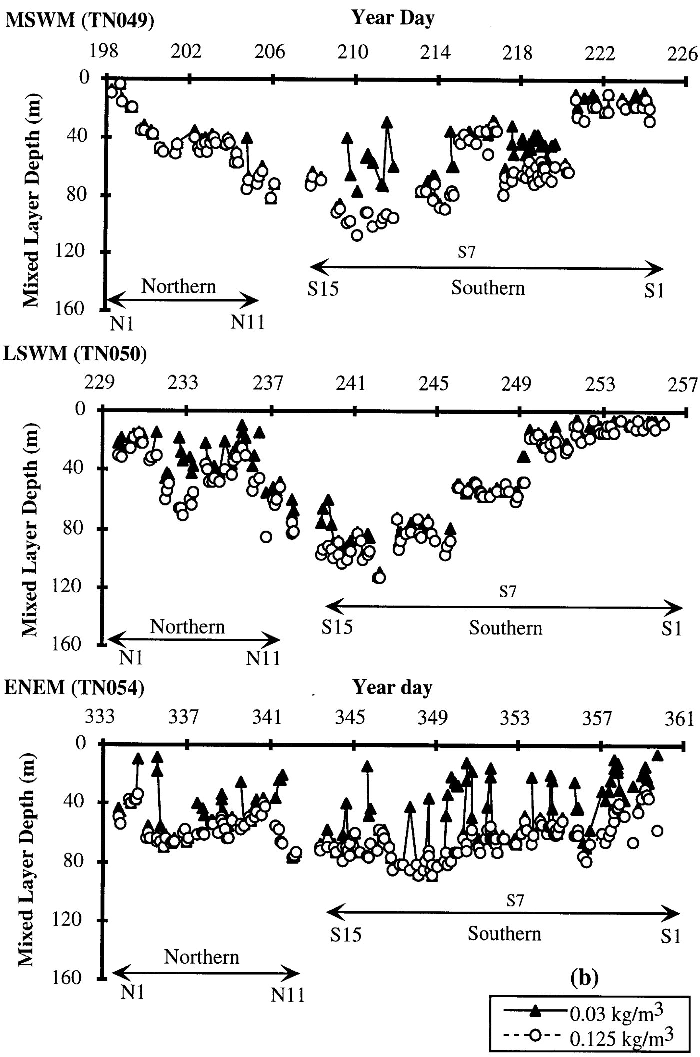


Fig. 5. Continued.

beam attenuation at the same depth, which we interpret as the most recent mixing depth.

3.2. Diel variations in MLD

In general the nighttime MLDs based on a density change of 0.125 kg m^{-3} were not much deeper than the MLDs based on a density change of 0.03 kg m^{-3}, suggesting that nighttime mixing usually extended down to the thermocline (Fig. 5). The largest differences occurred during the day when the more stringent 0.03 kg m^{-3} density change often indicated a much shallower MLD during the day. Diel changes in the 0.125 kg m^{-3} MLDs were observed occasionally, but the variations were much smaller than the 0.03 kg m^{-3} MLDs. Diel variations during the Northeast Monsoon (LNEM and ENEM) were much larger and more common than during other seasons. Diel fluctuations in the MLD using the 0.03 kg m^{-3} criterion were as great as 90 m (Day 24, LNEM, Fig. 5), but averaged 50 m. Thus, at a single station the variation in MLD on a time scale of hours equaled the spatial and seasonal variation in MLD for nearly the entire Arabian Sea. Such extremes were not the norm.

During the SI the diel MLD never changed by more than 50 m on the northern transect, but as noted above, the ocean during the first part of that cruise may still have had characteristics of NEM forcing. Other than at station S15 on the southern transect, diel variations during the SI were usually only 10–20 m (Fig. 5), but the percentage change in the MLD was sometimes 200–400% (See insert in Fig. 5a). Both the magnitude and percentage changes in the MLD during the Southwest Monsoon were usually small (average diel changes of 15 m for MLDs of 20–100 m). Near the central current and meteorological mooring (station S7) both the maximum MLD and diel MLD changes varied substantially through the seasons (Fig. 6). A comparison with the wind for that period of time (Fig. 3) indicates that MLD is not simply a function of wind speed; Ekman dynamics, regional variations in wind stress curl and

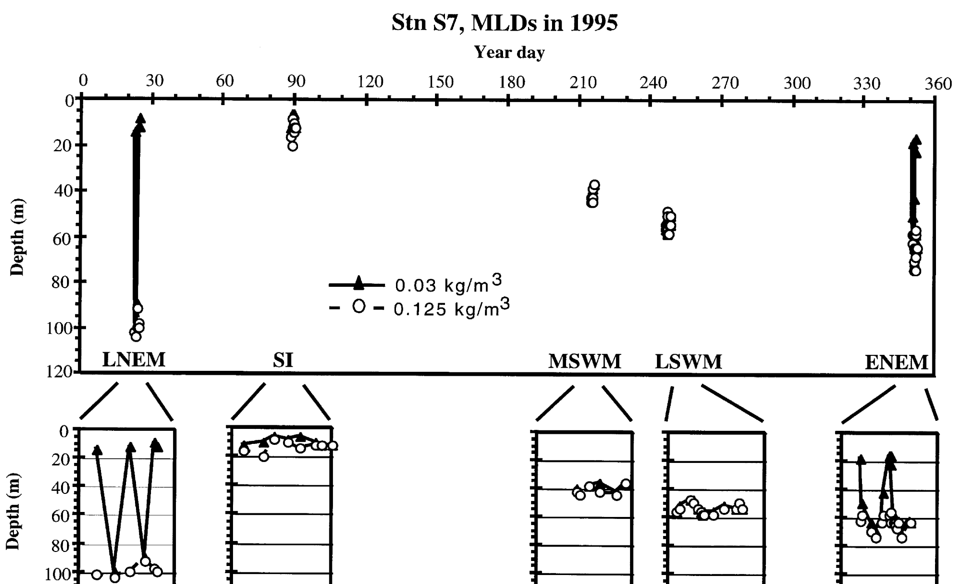


Fig. 6. Mixed-layer depths at station S7 during five cruises of 1995. The data from each cruise are expanded in the bottom panels. Diel differences between MLDs based on a $\Delta\sigma$ of 0.03 and 0.125 kg m^{-3} were large during the Northeast Monsoon cruises (LNEM and ENEM), moderate during the Spring Intermonsoon (SI) and generally small during the Southwest Monsoon (MSWM and LSWM).

convective cooling must also be factored in (Bauer et al., 1991; Weller et al., 1998; Wiggert et al., 1999).

3.3. POC and chlorophyll a

The conversion of beam c_p to POC values and chlorophyll *a* fluorescence to chlorophyll *a* was described in Gundersen et al. (1998). The slope $\pm$ intercept (and r^2) of the Model II regressions between POC ($\mu g\,l^{-1}$) and beam c_p were: TN043 – $387 + 2.5$ (0.87); TN045 – $492 - 12$ (0.89); TN049 – $484 + 5$ (0.92); and TN054 – $508 - 10$ (0.78). Samples for POC were not obtained during TN050, but PM samples were taken. Gundersen et al. (1998) compared POC : PM ratios on the other cruises and estimated a POC to beam c_p slope of 468. The equations for chlorophyll *a* vs fluorescence were based on a regression of all data from cruises TN049, 50 and 54. On cruises TN043 and TN045 the chlorophyll samples were taken from a different rosette than the one containing a fluorometer so no reasonable correlation could be made. The regressions from the three cruises were very similar, so the same regression was used for all cruises. There was, however, a difference in regressions between data shallower than the fluorescence maximum and data within and deeper than the fluorescence maximum. The equation for data down to the base of the fluorescence maximum was $\mathrm{Chl} = 0.357\mathrm{Fl} + 0.08$ ($r^2 = 0.87$). Below the fluorescence maximum, $\mathrm{Chl} = 0.389\mathrm{Fl} + 0.05$ ($r^2 = 0.93$). Detailed explanations and plots of the correlation data are contained in Gundersen et al. (1998).

Sections of POC and chlorophyll *a* in Gundersen et al. (1998) were made by averaging data from all casts at each station to minimize diel variations in the signal. Figures in this paper contain all individual casts at a station in order to illuminate diel variations and the effect of changing MLDs. We note that the influence of advection can not be excluded from our measurements, but the acoustic doppler current profiler (ADCP) data from the cruises were insufficient to assess or correct for advection because the ADCP yielded no data on the POC or chlorophyll *a* concentrations of water prior to it reaching the ship's location.

Diel variations occurred in POC (based on beam c_p) as well as in MLD and the diel POC changes could occur during any season and at most stations (Fig. 7), but diel variability in POC was greatest during the LNEM, especially nearshore along the southern transect and decreased offshore. The greatest diel variability around the basin occurred during the LSWM (Fig. 8a–c).

3.4. Diel variations at S7

Maximum sampling time at stations in the Arabian Sea was two days, providing only two cycles over which to examine diel variations in any parameter, which is less than ideal for statistical analysis. Nevertheless, diel variations in POC (based on c_p) were observed at these 48-h stations whenever there were diel variations in the MLD (Fig. 9). A plot of diel variations of POC, MLD, chlorophyll *a* and temperature at station S7 during 1995, reveals, however, that diel patterns for all parameters changed dramatically between seasons (Fig. 9). Temperature was contoured only to the depth

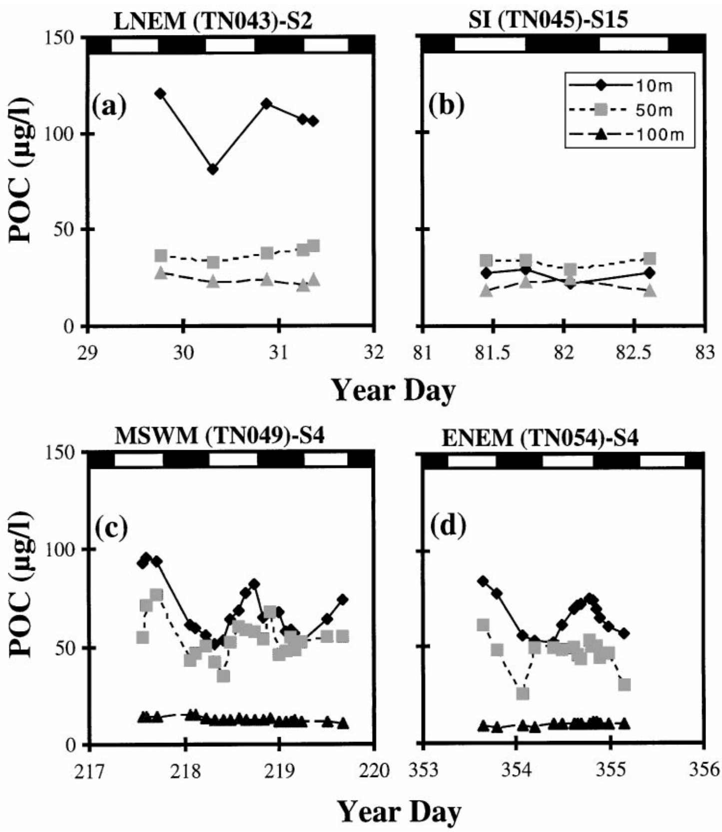


Fig. 7. Examples of diel variations of POC (based on beam c_p) at three depths during the occupation of two-day stations during four seasons. Each data point is a separate cast.

to which temperature decreased by 1°C from the surface value. A 0.5°C temperature change (corresponding to a 0.125 g m^{-3} density change) roughly corresponded with the maximum MLD. Diel changes in surface temperatures were also evident in the upper 10 m when there were large changes in MLD (Fig. 9).

Shallow mixed layers during the NEM (TN043 and TN054) had higher POC concentrations in surface waters that were mixed downward during the nighttime thickening of the MLD (Fig. 9, especially TN054). During the SI the MLD remained shallow and there was a subsurface POC and chlorophyll *a* maximum that was not interrupted by surface mixing. Continuously deep mixing during the LSWM maintained uniform POC concentrations within the mixed layer. Despite the deep mixed layer during the LSWM, however, there appeared to be layers of chlorophyll *a*. We suggest that some of these apparent layers were caused by random spikes in the profile whose source was chlorophyll *a* in aggregates, which Knauer et al. (1982) have shown to contain sufficient chlorophyll *a* to be important sites of primary production. In order to reduce spikiness in the fluorometer data, SeaTech fluorometers internally

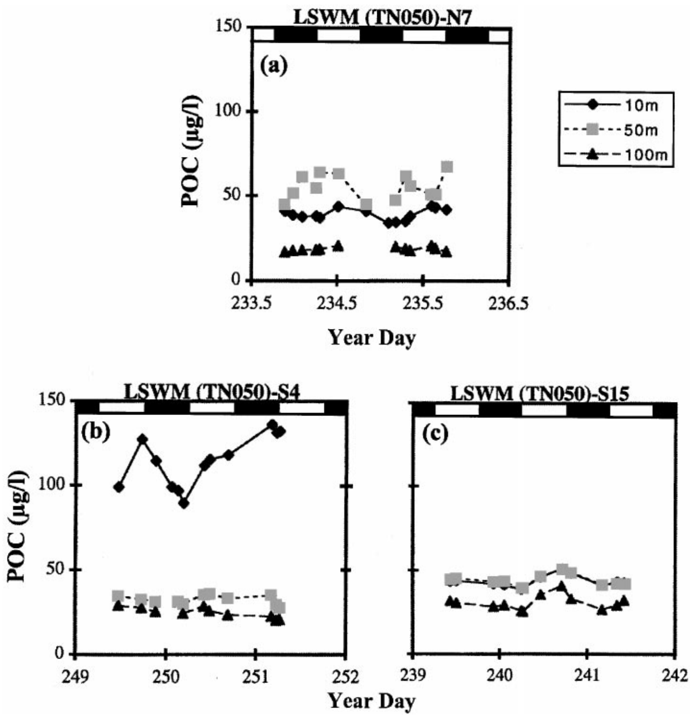
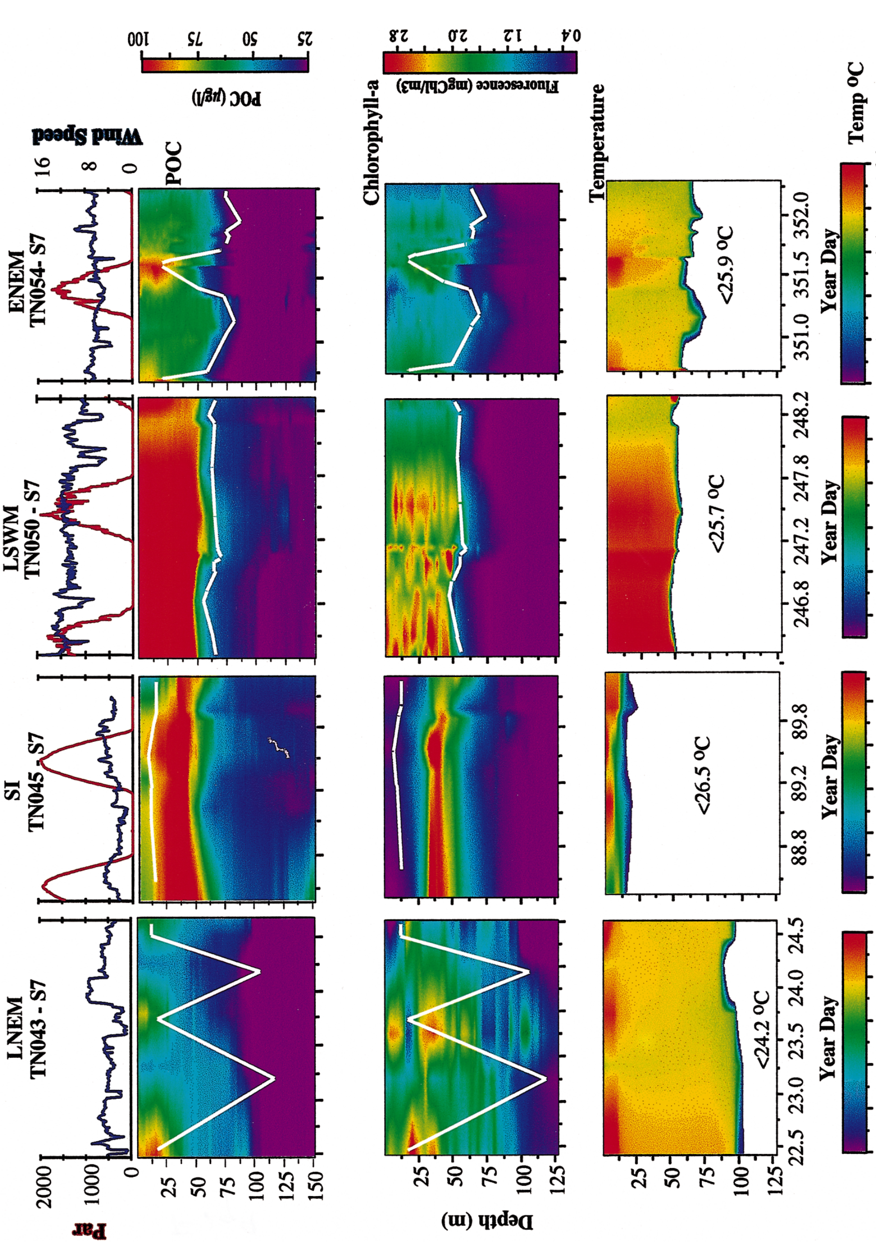


Fig. 8. Examples of diel variations of POC (based on beam c_p) at three depths during the occupation of two-day stations during the late Southwest Monsoon (LSWM). Spatial variability of the diel variations in POC and MLD was much greater during this period than during other cruises.

filter the output with a 3-s exponential decay constant. Thus a spike appears in the data as an instantaneous increase with an exponential decay with time (depth), which, with a lowering rate of 1 m s^{-1}, spread the signal over many meters (Fig. 10). The data were then bin-averaged to 2 m, smoothing the signal and making spikes appear as meters-thick layers when contoured. Note the fairly uniform fluorescence value if the spikes were removed. Aggregates were abundant during the SWM, but we did not obtain good aggregate data in the surface waters at this particular site to confirm our hypothesis. Note the lack of any layering during the LSWM in the POC section (Fig. 9). The chlorophyll *a* profiles during the SI seldom had spikes (Fig. 10).

4. Discussion

Banse (1994) synthesized data from many cruises in the Arabian Sea and laid out an annual hydrographic and biological scenario very similar to what was observed during the JGOFS Arabian Sea Expedition. However, there was not the high temporal and spatial resolution of hydrographic/optical sampling that JGOFS was able



LNEM TN043 - S7
SI TN045 - S7
LSWM TN050 - S7
ENEM TN054- S7
Par
Wind Speed
POC
POC (µg/l)
Chlorophyll-a
Fluorescence (mgChl/m3)
Temperature
Temp °C
Depth (m)
Year Day
<24.2 °C
<26.5 °C
<25.7 °C
<25.9 °C

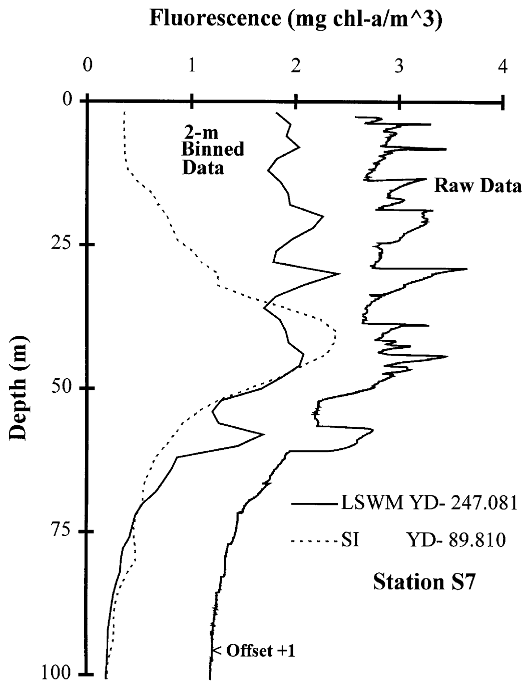


Fig. 10. Profiles of 2 m-binned chlorophyll *a* fluorescence at station S7 for the LSWM when there appeared to be layers within the mixed layer (Fig. 9) when temperature and POC were uniform, and during the SI when the data were very smooth. The raw fluorescence data are displayed and offset for the LSWM profile to demonstrate the "spike and exponential decay" pattern of the data. See text for explanation.

to achieve. Thus, from previous data it has not been possible to examine the diel changes in the region.

Our observations demonstrate that the mixed layer and diel changes in the mixed layer depth significantly affected the distribution of POC and chlorophyll *a*, and presumably influenced primary production by pumping nutrients upward and by controlling the light levels experienced by plankton. Optical measurements such as those of beam attenuation and chlorophyll *a* fluorescence have shown diel variability in many of the worlds oceans (Pacific – Siegel et al., 1989; North Atlantic – Gardner et al., 1993; Equatorial Pacific – Gardner et al., 1995; Walsh et al., 1995). The daytime increases in particles (based on c_p) in surface waters were generally attributed to primary production (less grazing) and the nighttime decreases in

Fig. 9. Time-series at station S7 during four seasons. Wind speed and PAR are from shipboard data; POC is derived from calibrated beam c_p values; chlorophyll *a* is from calibrated fluorescence; and white lines are the MLD based on a density change of 0.03 kg m^{-3}. A profile was made at each point where the MLD line shows a change in slope. Temperature scales are different for each plot, but each plot spans only 1°C; thin layers of 1°C correspond with thin mixed layers (SI) and thick layers of 1°C correspond with thick mixed layers (LNEM).

particles to zooplankton grazing (Siegel et al., 1989; Abbott et al., 1990; Dickey et al., 1991). Grazing plays the major role in particle loss (Landry and Hassett, 1995; Landry et al., 1998), but Gardner et al. (1993,1995) demonstrated that the nighttime decreases in particles also resulted from a thickening of the mixed layer, which entrained deeper, low-particle concentration water into the mixed layer, thus diluting the particle concentration. Aggregation and particle settling play a role in the decrease. Aggregated particles may remain within the mixed layer, but they may not be quantitatively sensed in the beam c_p signal since the volume of water illuminated by a transmissometer is only 40 ml. Aggregate abundance is usually only a few per liter in the open ocean (Asper, 1987; Alldredge and Silver, 1988; Walsh et al., 1998), so the probability of intercepting an aggregate in 0.04 l is small.

Interpreting diel variations from optically determined chlorophyll *a* has added complications of photoadaption and quenching. The fluorescence per unit of chlorophyll *a* in plankton decreases as the cells become "quenched" with light; thus predicted chlorophyll *a* should decrease near midday. This is best observed during the NEM (Fig. 9), but it is difficult to differentiate between effects of quenching vs. mixing. Furthermore, the fluorescence per unit of chlorophyll *a* may vary with phytoplankton species (Marra, 1997). No corrections have been attempted here to quantify this potential variability.

4.1. Diel changes in the mixed layer

Mixed-layer depths varied substantially during the LNEM and ENEM (see also Fig. 5) as heating stratified the surface water during the day and convective cooling mixed the water at night. Nighttime deepening entrained nutrient-rich waters into the mixed layer, and daytime shallowing of the mixed layer confined these nutrient-rich waters to high light levels that would accelerate biological production. Landry et al. (1998) noted that phytoplankton cells grew at rates well below their potential when NO_3 concentrations fell below $0.5\,\mu M\,l^{-1}$. Gundersen et al. (1998) mapped the $0.5\,\mu M\,l^{-1}$ NO_3 isopleth on basin-wide sections of POC and found a reasonable correlation between NO_3 and POC abundance, but nutrients were not sampled sufficiently frequently to determine diel variations. Denman and Marra (1986) demonstrated that large variations in the depth to which phytoplankton are mixed have a significant impact on their subsequent growth because of changes in the light field to which phytoplankton are exposed.

When the mixed layer is shallow, layers of phytoplankton and fluorescence can develop (Fig. 9; LNEM). These layers disappear during deep mixing. However, there appear to be layers of chlorophyll *a* in the deep mixed layer of the LSWM (Fig. 9), but we demonstrated earlier in this paper that the "layers" result primarily from random spikes in the profile that are probably caused by chlorophyll *a* in aggregates (Fig. 10).

During the SI the diel variations of MLD, POC and chlorophyll *a* were minimal at S7 (Figs. 5 and 9). Fig. 5 (inset) shows minimal absolute diel variation in the MLD during the SI (40 m along the northern transect, which was still influenced by NEM conditions, and 20 m along the southern transect); however, the percentage change in the MLD was significant (200–400%). The small MLD variations limited the mixing of nutrients into the surface layer ($NO_3 < 0.5$ micromoles l^{-1}) during this period of low winds and subsequent upwelling (Morrison et al., 1998), thus maintaining oligotrophic conditions (usually characterized by low surface nutrients and primary production, with a subsurface chlorophyll maxima; Gould, 1987). A chlorophyll *a* and POC maximum beneath the mixed layer persisted in most of the Arabian Sea and was matched by a sub-MLD peak in primary productivity (R. Barber, pers. comm.). The shallow mixing was a function of low wind speeds, solar insulation and little night time convective cooling during the SI. Wiggert et al. (1999) suggested that the MLD was controlled by the depth of the seasonal thermocline, but both the MLD and thermocline likely result from the regional and local climatic forcing (Bauer et al., 1991).

During the SWM wind mixing was vigorous and daytime heating was not sufficient to stratify the water, so the MLDs remained consistently deep, with fairly uniform POC in the mixed layer (< 10–20% diel variation) and little diel variation in MLD (Figs. 5 and 9). Unlike the SI, nutrients were plentiful in the surface waters in the SWM due to wind mixing and Ekman-induced upwelling. The 1% light level was about 50–65 m in this region during the SMW (R. Barber, pers. comm.), which encompasses the entire mixed layer. It is not surprising, therefore, that phytoplankton growth exceeded grazing (Landry et al., 1998) and allowed the buildup of POC and chlorophyll *a* (Fig. 9).

4.2. Particle loss from the mixed layer

We also seek to assess the role of changing MLD in the export of particles from surface waters. Concentrations of POC and chlorophyll *a* in the ocean are generally high in the surface 50–100 m and then decrease rapidly below that depth. The main reason for the high values is biogenic production in the euphotic zone, but the dynamics of the mixed layer also play a role.

To understand better the dynamics of the mixed layer, D'Asaro et al. (1996) developed a neutrally buoyant float that could be deployed in the mixed layer to trace the movement of a parcel of water during active mixing. The results of a deployment over several days in the northeast Pacific Ocean were compiled as a single 24-h plot and clearly delineated the differences in the depth of mixing between day and night (Fig. 11). The trace of the float matches the MLD model of Brainerd and Gregg (1995) (Fig. 2), and would be very similar to the path of a random non-motile plankter in the mixed layer as modeled by Woods and Onken (1982). The 2–4 h cycles of the float are direct evidence that water is actively overturning in surface waters. This mixing will have a significant impact on the distribution of both dissolved (nutrients, salinity, gases) and particulate components (plankton, detritus, aggregates). However, it does not explain how particles are removed from surface layers.

Carbon fixed by plankton into particulate form in surface waters can settle due to gravitational forces, aggregate by physical or biological forces – which generally causes them to settle more rapidly (McCave, 1984; Alldredge and Silver, 1988), be consumed and transported by migrating organisms (Dam et al., 1995), or remineralized to DOC and DIC and advected laterally or diffused and mixed vertically (Ducklow et al., 1995; Hansell et al., 1997). It is generally estimated that about 90% of

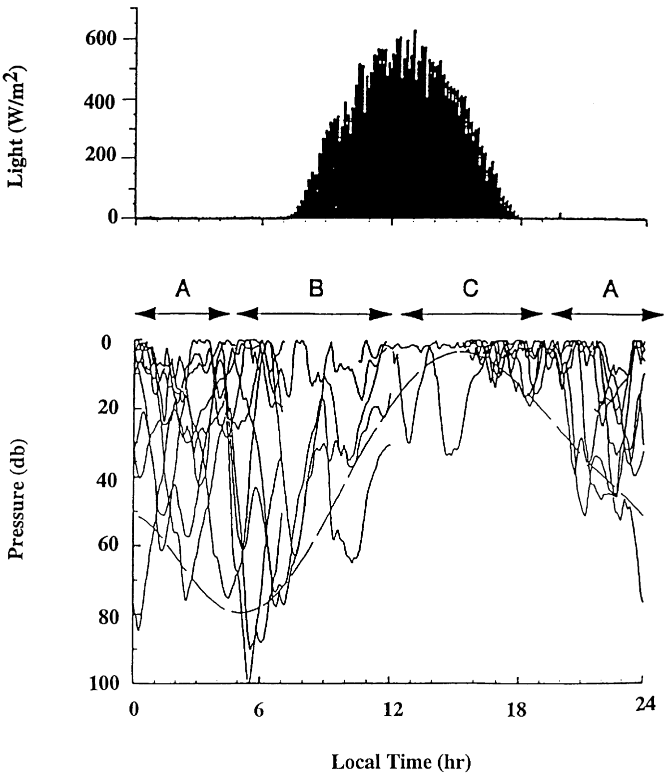


Fig. 11. Time–depth trajectory of a neutrally buoyant float designed by D'Asaro et al. (1996) over several daily cycles superimposed on one 24-h plot showing (A) deepening convection during the night, (B) inhibition of the mixing depth and subsequent restratification, and (C) isolation of a thermally stratified surface layer by solar radiation allowing only shallow mixing during the day (after McNeil and Farmer, 1995). Mean solar input is shown in the upper panel. Excursions through the mixed layer took 1–4 h under the conditions at that time.

the carbon fixed during primary production is recycled in surface waters and only about 10% is exported. Flux estimates from measurements of the ^{234}Th deficiency in surface waters (Buesseler, 1998) support this estimate, except during blooms, when substantially more material may be exported. Data from Joint Global Ocean Flux Studies (JGOFS) suggest that 50–80% of the carbon exported is in particulate form (Ducklow et al., 1995; Hansell et al., 1997). To understand mechanisms by which slow-settling particles leave the mixed layer, Lande and Wood (1987) made a theoretical analysis that suggested that in the presence of a surface mixed layer with high turbulent diffusion, particles with slow sinking rates make many rapid excursions up and down between the interior of the mixed layer and the top few meters of the underlying thermocline before they permanently leave the surface layer. To understand better distributions of dissolved and particulate components in the mixed layer and the mechanisms by which particulate matter exists the surface layer, detailed field measurements were needed.

In the Equatorial Pacific Ocean, the mixed layer sometimes deepened from 10 to 70 m in just a few hours, moving small particles downward at a much faster rate than their sinking velocity. Restratification the following day left many of these particles in deeper water where they could settle uninfluenced by mixing in surface layers. Rapidly settling aggregates could easily settle to a depth deeper than the deepest mixing of the following day, so they could not be re-entrained. This "mixed-layer pump" was shown to be a potentially important mechanism for removing particles from the surface layer Gardner et al. (1995). It could also be important in transferring upward or downward any component for which a gradient had been developed over the depth of the mixed-layer changes. For example, nutrients could be "pumped" upward on a daily basis, enhancing primary production. This condition has also been noted for longer-period changes in mixed-layer depth (Eppley et al., 1988; Marra and Ho, 1993).

Kerr and Kuiper (1997) performed laboratory experiments and theoretical analysis on the effect of diffusional mixing on the loss of particles from two-layer and multilayer systems. They found that the number of particles in a convecting fluid decreased exponentially with time, with a decay rate equal to the settling velocity of the particles divided by the depth of the layer. $N = N_0 e^{-(w_s t/h)}$ where N_0 is the original number of particles in the layer, N is the number at time t, w_s is the particle settling velocity and h is the layer thickness. Thus, after the length of time required for a particle with a given settling rate to sink through the entire depth of the layer ($t = h/w_s$), $> 1/3$ of the particles initially mixed in the layer still remained. The settling velocity of a given particle was not altered, but the convective mixing increased the time that particles circulated in the layer (i.e., their residence time). Kerr and Kuiper (1997) noted that this result was predicted by Smith (1982) and was confirmed by Martin and Nokes (1988,1989) and holds as long as (1) the settling velocity of the particles is less than half the vertical convecting velocity, so that the particles are uniformly mixed in the flow, (2) the particle concentration is sufficiently small that the particles do not influence each other or the flow, and (3) the particles are not re-entrained from the boundary. This implies that rapidly settling particles are affected less than slowly settling particles. However, the particles used in their experiments had settling velocities of about $0.15\ \mathrm{cm\ s^{-1}}$ ($130\ \mathrm{m\ day^{-1}}$), which suggests that most aggregates and fecal pellets are affected by this type of mixing.

Indeed, Alldredge et al. (1987) found evidence of sustained residence times for large fecal pellets in the mixed layer, and presented arguments similar to the above to explain how they can be maintained there. They also appealed to turbulence as a means of maintaining particles in the mixed layer. Although it might seem intuitive that the intensity of mixing would affect the length of time spent in the mixed layer, the theory and experiments of Kerr and Kuiper (1997) indicate it is only the settling velocity and mixed layer thickness which control the retention of particles.

Kerr and Kuiper (1997) point out that if particles are released at a constant rate at the top of the water column, there will be an equal flux of particles out of the mixed layers once steady state is achieved, regardless of the presence or absence of convective mixing. In the open ocean, however, regular changes in the MLD could affect the export of particles from the mixed layer. If we assume that steady state would require 1–2 decay periods for the settling rates of the particles in question, this would take 10 to 20 h for a 60-m mixed layer for particles settling 130 m d^{-1}, but based on a settling rate of a 2–4 m d^{-1} for individual plankton, it would take weeks to achieve steady state.

Diel changes in the MLD could regularly isolate particles from the intense mixing in surface waters as mixed layers shallow during the day and allow the particles below the mixed layer to settle at their terminal settling velocity. This would be particularly important for aggregates or fecal pellets because they could easily settle deeper than the MLD of the following evening. In this manner, particles could be "pumped" from surface waters (Gardner et al., 1995). Conversely, Alldredge et al. (1987) argued that nighttime increases in the intensity of mixing (expressed here as an increase in the depth of active mixing) would redistribute throughout the mixed layer any pellets not yet lost from the mixed layer, thus prolonging their residence time. Obviously the net effect will be driven by the variations of the MLD and settling velocity of the particles present.

4.3. Diel changes in MLD and carbon export

To quantify the potential role of diel changes in MLD on the export of carbon in the Arabian Sea, we compared POC profiles from afternoon and evening at station S7 during the ENEM when the MLD increased from 16 to 66 m (Figs. 9 and 12a). Subtracting evening values from afternoon values shows that there was a 640 mg C m^{-2} loss of POC through 48 m and a 279 mg C m^{-2} gain of particles between 48 and 75 m. The net loss down to 75 m was $\sim$ 360 mg C m^{-2} (Fig. 12b). The standing crop of POC (0–75 m) in the afternoon was 2120 mg C m^{-2}, so about 17% of the standing crop was moved downward, where it is more likely to sink or be grazed.

For comparison. measurements of primary productivity at this station were estimated to be $\sim$ 870 mg C m^{-2} during the day (R. Barber, pers. comm.), and the grazing ratio (grazing: growth) was 1.0 (Landry et al., 1998), so grazing consumed $\sim$ 870 mg C m^{-2} d^{-1}. Thus, the 360 mg C m^{-2} moved downward by mixed layer deepening during the afternoon could have been removed by grazing, including microbial activity, and some could have been packaged.

The carbon flux measured in traps at $\sim$ 800 m at the nearest station where traps functioned properly during this time (S4) was $\sim$ 10 mg C m^{-2} d^{-1} (Honjo et al., 1999), so only $\sim$ 3% of the small-particle carbon moved downward during the night needs to reach the traps to produce the observed flux. It is more likely for permanent losses of POC to occur in the form of fecal pellets or rapidly settling aggregates, which are not quantitatively measured in the beam attenuation signal. Most of the beam attenuation signal is caused by particles < 20 μm (Pak et al., 1988; Chung et al., 1996).

While this one example suggests that changes in MLD can be important for removing POC from surface waters, short-term measurements such as these are likely to vary significantly as environmental conditions change. Furthermore, the distance between 75 m (bottom of the mixed layer) and 800 m (depth of the sediment trap

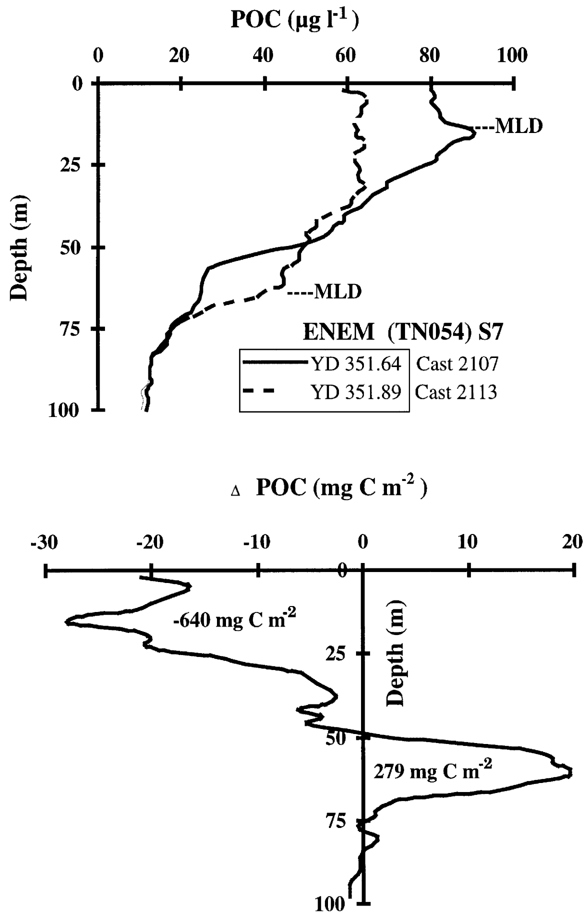


Fig. 12. (a) Profiles of c_p in the afternoon and evening at station S7 when the mixed layer depths were 16 and 66 m respectively; (b) The difference in POC concentrations (as predicted by the c_p : POC ratio) between the two profiles.

closest to the surface) is large, so considerable grazing and bacterial remineralization would be expected between those depths, as was insightfully discussed by Banse (1994). Mooring and modeled data of Wiggert et al. (1999) showed large diel variations in MLD at S7 during January (LNEM). Unfortunately, we do not have continuous measurements of POC and MLD variations during that cruise. If diel mixing is important for export of particles from the mixed layer, one would expect a positive correlation between frequency of large diel variations in the MLD and the fluxes registered in the moored traps in different seasons.

Diel changes in MLD and POC were greatest during the LNEM and ENEM, so conditions should have been optimal for pumping organic matter out of the surface layer and pumping nutrients into the surface layer if the MLD straddles the nutricline.

However, Chavez and Toggweiler (1995) argue that entrainment of nutrient-rich waters by mixing is not as efficient as upwelling as a source of new nutrients in surface waters. They state that globally, 65% of new production is due directly to upwelling whereas only 35% is due to convective and vertical mixing. During the NEM the diel changes in MLD seldom crossed the nutricline, so primary production was not extremely high, though it was much higher than expected (R. Barber, pers. comm.). During the SI, winds were low and changes in MLD were very small, so oligotrophic conditions prevailed. The wind-driven Ekman dynamics of the Southwest Monsoon causes nutrients to be pumped into the surface waters (Bauer et al., 1991), creating blooms of large diatoms that increase the biomass in surface waters because of high production. Eventually they aggregate or settle rapidly and carry carbon out of surface waters. These large diatoms or aggregates will settle much more rapidly (0.1–100 m/day; Smayda, 1970; Alldredge and Gotschalk, 1989; Diercks and Asper, 1997) than the small picoplankton (< 1 m d^{-1}; Smayda, 1970) that create the chlorophyll *a* maximum below the mixed layer during the SI (Fig. 9). As a result, trap fluxes were higher during the NEM than during the SI when diel MLD changes were smallest, but trap fluxes were 2–3 times larger during the SWM when diel MLD changes were small.

Banse (1994) noted that in past studies the deep chlorophyll *a* maximum during the spring intermonsoon was deeper than the detection depth of a satellite color sensor, which is $1/k$ (k is the vertical attenuation coefficient of light in m^{-1}), and is seldom > 25 m. The same is true in this study where the chlorophyll *a* maximum is around 50 m (Gundersen et al., 1998). Campbell et al. (1999) found subsurface maxima in heterotrophic bacteria, Prochlorococcus, Synechococcus and Picoeucaryotic algae during the SI. The latter three are the most likely source of the high chlorophyll *a* in the deep chlorophyll maximum. Prochlorococcus were probably the largest contributors to the deep chlorophyll maximum because of their high chlorophyll : fluorescence ratio (Veldhuis et al., 1993), which actually increased with depth in the Arabian Sea (Johnson et al., 1999). These picoplankton have slow settling velocities (m d^{-1}), but were still actively growing (Johnson et al., 1999) in the low-light, high nutrient environment. Without large, rapidly settling frustules of carbonate or opal, they probably contribute little to the flux to the seafloor as evidenced by the small trap fluxes during the SI (Honjo et al., 1999).

Thus, although the dynamics of the mixed layer are important in controlling the short-term distribution of dissolved and particulate components, the seasonal productivity and export of POC to a depth of 1 km is controlled more by changes in the large-scale, longer term forcing functions such as upwelling, which can inject new nutrients into the mixed layer, thus changing community structure by stimulating diatom blooms. Diatoms contribute more to the export flux because of their large size and settling rate, and their dominant role in blooms has been argued by many investigators (Banse, 1994; Landry et al., 1997; R. Barber, pers. comm.). Our data suggest that changes in the MLD can move POC downward. However, we have much to learn about the recycling that occurs beneath the photic zone. It is important to consider that at times in the Arabian Sea (and in other oceans) when wind-driven Ekman dynamics are not as extreme, the daily dynamics of the MLD may be of greater importance to the production and export of organic matter and other biogeochemical components.

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