The Oman upwelling zone during 1993, 1994 and 1995

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Abstract

Satellite-derived sea-surface temperature, TOPEX/POSEIDON (T/P) sea-level anomalies (SLAs), model wind data, and hydrographic data are used to characterize the upwelling along the Oman coast during the US Joint Global Ocean Flux Study (US JGOFS) Arabian Sea Process Study (ASPS) in 1995 as well as to look at interannual variability in the upwelling over the period 1993–1995.

Empirical orthogonal function (EOF) analysis of the satellite-derived sea-surface temperature (SST) at the locations of the US JGOFS standard stations shows the first mode, which represents a biannual variability, contributes 67\% of the total variance. In addition, the SST shows the upwelling “front” moving offshore with the development of Southwest (SW) Monsoon in early June 1995, reaching a maximum distance of approximately 120 km by late August 1995. Finally, SST shows the persistence of cold upwelling waters for nearly a month after the end of the SW Monsoon within the bays along the Oman coast.

TOPEX/POSEIDON SLAs indicate that with the onset of the SW Monsoon, a 30-cm drop in steric height is observed along the Oman coast associated the presence of the cool upwelled waters. This drop in steric height sets up a horizontal pressure gradient and results in a compensating along-shore, northeastward-flowing, geostrophic current (East Arabian Current; EAC) during the SW Monsoon. Similarly, the altimeter data show an offshore decrease in steric height during the Northeast (NE) Monsoon, indicating a seasonal reversal in direction of the EAC with flow to the southwest. Subsurface temperature data indicate that the actual uplifting of isotherms associated with the upwelling can be found to a distance of approximately 260 km from the shore and to a depth of 150–200 m. Using along-track altimetry data, we estimate that, for a region 260 km in offshore distance and 600 km alongshore, $2.2 \times 10^6$, $1.4 \times 10^6$ and $0.55 \times 10^6$ m$^3$ s$^{-1}$ were upwelled through the 100 m level with upwelling velocities...
O (2.0 × 10$^{-5}$ m s$^{-1}$), during the SW Monsoons of 1993, 1994 and 1995, respectively. The reduced upwelling in the summer of 1995 is attributed to a reduction in wind-stress curl along the Arabian coast when compared to 1993 and 1994. © 2000 Elsevier Science Ltd. All rights reserved.

1. Introduction

The wind field over the Arabian Sea undergoes a semiannual reversal associated with the monsoon. In summer, heating of the Asian land mass results in low pressure over Asia and high pressure over the Indian Ocean. The geostrophically balanced airflow results in strong, topographically steered southwesterly winds (SW Monsoon) during the summer (e.g., Findlater, 1969; Bruce, 1983; Hastenrath and Lamb, 1979). During winter, the cooling of the northern hemisphere land mass results in a high pressure over land and low pressure over the Indian Ocean that reverses the direction of the monsoon winds from southwesterly to northeasterly (NE Monsoon). The winds are generally weak during the intermonsoonal periods, with the Fall Intermonsoon being significantly shorter than the Spring Intermonsoon (Weller et al., 1998). The canonical description of the wind-driven upwelling regime during the SW Monsoon involves coastal upwelling driven by the wind stress and an offshore upwelling regime driven by the curl of the wind stress (e.g. Smith and Bottero, 1977) that occurs inshore of the high-velocity core of the Findlater Jet (Findlater, 1969). Mesoscale sampling (Brink et al., 1998), in combination with satellite sea-surface temperature (SST) data, shows that during the SW Monsoon the dynamics in the upwelling zone include cool filaments, similar to those found in other upwelling regions (Abbott and Barksdale, 1995).

The most intense upwelling, with SSTs 5°C or more below the ambient offshore temperature, is observed near the headlands of the capes along the Arabian Peninsula (especially along the coast of Oman), with filaments of cold water extending into the Arabian Sea (Elliot and Savidge, 1990; Manghnani et al., 1998). Smith and Bottero (1977), using the wind and hydrographic data, estimate that in a region that extends 1000 km along the coast of the Arabian Peninsula and 400 km offshore, 8 × 10$^6$ m$^3$ s$^{-1}$ water upwells through the 50-m level with maximum vertical velocities of 1–2 × 10$^{-5}$ m s$^{-1}$ during the Southwest Monsoon.

Due to the geostrophic adjustment associated with the decrease in sea level along the coast during the upwelling season, a northeastward alongshore current (East Arabian Current) is observed along the coast (Shi et al., 1999). This current leaves the coast as a jet at Ras al Hadd (Böhm et al., 1999), effectively separating the Gulf of Oman and northern extremes of the Arabian Sea from the region studied by ASPS. The alongshore flux in the top 300 m of a region extending 100 km from the Oman coast is estimated to be 10 × 10$^6$ m$^3$ (Elliot and Savidge, 1990).


2. Data

2.1. TOPEX/POSEIDON altimeter data

The TOPEX/POSEIDON (T/P) altimeter uses a dual-frequency radar system to determine the height of the ocean surface to an accuracy of about 4 cm pointwise (Fu et al., 1994). The data used in this study are Sea Level Anomalies (SLAs) derived from the measurements of the height of the ocean surface relative to a 3-yr mean field to which all the standard environmental corrections are applied. The TOPEX/POSEIDON altimeter has a 10-day repeat cycle. The spacing between individual ground tracks of the satellite is about 250 km and the along-track grid points have been sub-sampled at 25-km resolution. In this study, two separate ground tracks that cross the Arabian Sea upwelling area are chosen to form an upwelling box (Fig. 1), and the SLA measurements on the two tracks are used to calculate the geostrophic transport anomalies through the upwelling zone. At each end of the sub-tracks (U1, U2 and U3; Fig. 1), the mean SLA is an average of three consecutive data points. With the SLA pointwise error about 4 cm, the error of the average SLA is only 1.3 cm. In addition, if the SLAs were averaged over the whole SW Monsoon (approximately 9 T/P crossings), the error of the average SLAs would be about 0.5 cm.

Fig. 1. Charts showing the standard JGOFS Stations and the boundaries that have been chosen to encompass the upwelling region in this study: (A) Standard JGOFS station positions; and (B) Stations P1 and P2 are pseudostations chosen for extraction of SST time series in the region of persistent, strong upwelling; Sections U1, U2, U3 and U4 that define the upwelling region. Various geographical features referred to in the text and the 1000-m isobath are also displayed in this figure.
In order to obtain a qualitative description of the geostrophic flow field, the SLAs are also objectively interpolated to a 0.25 resolution grid using the method described by Hendricks et al. (1996). The objective analysis scheme dictates that the greater the number of along-track points that sample an ocean feature the more realistic its spatial representation in the gridded data. If a feature varies temporally in intensity or moves about a mean position in space at time scales less than 10 days (T/P repeat period), it may appear dilated in the objectively analyzed maps. Thus, while the analysis may not preserve fine-scale structure in SLA, mesoscale features O (200 km) in size with time scales of O (10 days) are qualitatively described. When used with these inherent limitations in mind and in conjunction with the other data sets, these data can provide valuable insight into the dynamics of the region.

It must be remembered that as we do not have an adequate geoid at T/P scales, we are dealing with sea-level anomalies (SLAs) and not absolute sea level. Therefore, we are actually resolving the time varying portion of the geostrophic velocities and transports perpendicular to sections U1, U2 and U3. If the seasonal variability of the northern Indian Ocean is so great that the mean flow, if any, is comparatively negligible, then the utilization of SLA data is particularly advantageous. In fact if the mean can be neglected relative to the instantaneously oscillating field associated with the monsoonal forcing, the residual geostrophic velocity associated with the sea-level deformation field is representative of the total geostrophic flow. Shi et al. (1999) show that this assumption is a reasonable one in the northern Arabian Sea. They argue that, when using the altimetry data to compute the variable portion of the geostrophic velocity and transport and the modeled wind data to compute Ekman transports, the Ekman transport (as expected when the winds reverse) and the geostrophic transports both change sign almost “instantaneously”, suggesting that the mean flow is negligible. Therefore it appears that in this region, the T/P SLAs are representative of the absolute flow field to within a small (approaching zero) arbitrary mean. This implies that to a first approximation we can assume that the geostrophic velocities and transports estimated using T/P SLA are representative of the total geostrophic velocity and transports. This assumption will be used to estimate transport in the upwelling area. Keep in mind in our characterization of the upwelling that even if this assumption is not true, the geostrophic velocities and transport anomalies should still be representative of the seasonally varying coastal current and upwelling, as these flows are driven by the seasonally varying portion of the winds.

Sections U1, U2 and U3 enclose the main upwelling region (Fig. 1). The average T/P SLAs (see above) are used to estimate the sea-surface slope and thus the velocities and transports in the upwelling zone. Section U4, from the coast at 18.25°N, 56.765°E to an offshore location at 16°N, 59.0°E, is used to quantify the offshore extent of the upwelling. (Sections U1, U2 and U4 are along T/P ground tracks.)

2.2. Wind field data

Modeled wind-stress data for the Arabian Sea for 1993 through 1995 from the Navy Operational Global Atmosphere Prediction System (NOGAPS) are used in the study (John Kindle, personal communication). NOGAPS is a global spectral model with
a resolution of 79 triangular truncation, corresponding to a 1.15° transform grid. NOGAPS has proven itself to be competitive with any of the large forecast models run by the large operational forecast centers around the world (Rosmond, 1992). The NOGAPS wind fields agree well with measurements made by a mooring array in the central Arabian Sea during 1995. Weller et al. (1998) estimate that the NOGAPS wind stress was overvalued by 24% in the winter and undervalued by 23% in the summer. In this study, in order to be consistent with the repeat cycle of the TOPEX data, the wind stress is averaged over a 10-day period. After applying the error corrections on the wind stress for the SW Monsoon and NE Monsoon separately, the wind-stress data have an accuracy of approximately 5%.

2.3. Satellite sea-surface temperature data

Satellite SST data for 1995 are derived from infrared data collected by the Advanced Very High Resolution Radiometer (AVHRR) sensor onboard the NOAA-14 polar-orbiting environmental satellite (POES). The AVHRR data have a spatial resolution of approximately 1.1 km at nadir. The data were collected by an HRPT satellite receiving station in Muscat, Oman, that was operated by the Oman Ministry of Agriculture and Fisheries as part of the ASPS (Morrison and Brown, 1997). A nonlinear, 2-channel algorithm was applied to the infrared data to derive the SST (McMillin and Crosby, 1984) with an accuracy of 0.5–1.0°C. In this study, satellite-derived SSTs are used to describe the seasonal cycle of SST in the upwelling area. As the daytime images at times show strong surface warming, only nighttime passes are used in the SST analysis. Time series of SST are constructed for each ASPS standard station position (see Fig. 1) by computing a mean SST for a 5-km² square centered at each station position. Also, to quantify the SST variation in the upwelling zone, locations within the two bays where the strongest upwelling is observed are chosen as pseudostations (Fig. 1) at 18.603°N, 57.020°E for pseudostation P1 and 17.649°N, 55.728°E for pseudostation P2.

At most mid-latitude locations SST varies seasonally, reaching a maximum in mid-to-late summer and a minimum in mid-to-late winter. The SST in the northern Arabian Sea is a notable exception to this trend. Here, there is noticeable temporal and spatial variability in the SST associated with atypical summer cooling of the sea surface (‘Arabian Sea Cooling’) where convective overturning due to strong winds, advection of cool water from the upwelling zone along the coast of Oman, and/or open ocean upwelling, results in a cooling of the water and a reduction in the steric height (Düning and Leetmaa, 1980). One method to sort out which of these processes is responsible for the reduction in SST is Principal Component Analysis (PCA). PCA is well suited for resolving complex variance patterns of physical fields into simple pieces (Kelly, 1986; Miller, 1993; Rasmussen et al., 1981). Principal component analysis of a physical field produces a set of eigenvectors. These are the “empirical orthogonal functions” (EOFs): “empirical” because they arise from data, and “orthogonal” because they are uncorrelated over space. From the EOFs, one can construct the principal components (or amplitudes) of the data set. The principal components have the important property of temporal uncorrelatedness, and they may carry information
about the variance of the data set along the directions of the eigenvectors. Empirical orthogonal functions do not describe the dynamics in a system but are useful to find the statistically significant patterns of the variations. The first few EOFs, which contain most of the variance in the data, provide excellent information on the spatial and temporal variation of the physical field. An EOF analysis of the time-series of SST extracted from the satellite imagery at each of the ASPS standard station positions helps us understand the SST variations.

2.4. Hydrographic data

Between September 1994 and December 1995, the ASPS collected extensive, high-quality hydrographic data (temperature, salinity, dissolved oxygen and nutrients) during all seasons in the northern Arabian Sea. An analysis of this unique data suite is presented in Morrison (1997) and Morrison et al. (1997, 1998). All of the hydrographic data are available from the US JGOFS database maintained at the Woods Hole Oceanographic Institution (http://www1.whoi.edu). “Readme” files accompany the data from each of the R/V Thompson legs and give details on data processing and collection procedures, as well as on data quality. In this paper, the mean geostrophic shear (Shi et al., 1999) derived from the CTD data for cruise TN049 in the summer of 1995 (07/18/95–08/13/95) and for cruise TN054 in the winter of 1995 (11/30/95–12/26/95) are used in the computation of the upper layer transport in the upwelling zone.

3. Results

3.1. Sea surface temperature

The mean SST of the northwestern Arabian Sea for the summer of 1995 (Fig. 2) shows that the lowest SST (\(< 23°C\)) is observed in the Oman upwelling zone. The mean SST gradually increases with the offshore distance, reaching 25–26°C in the central basin. At the mouth of the Gulf of Oman, a strong front is observed off Ras al Hadd. This is the offshore extension of the Arabian Coastal Current (Böhm et al., 1999). This jet separates relatively cool (25–26°C) waters found in the central Arabian Sea and waters \(> 30°C\) found within the Gulf of Oman and northern Arabian Sea. In the Gulf of Aden, a strong thermal front also is observed with the temperature difference of 4–5°C, separating the warm surface waters of the Gulf of Aden from the cooler surface temperatures found in the Arabian Sea. Far off the Arabian coast, the mean SST is approximately 25–26°C. While considerably warmer than SST in the upwelling region, temperatures in this region are still much lower than observed here before the onset of SW Monsoon when temperatures are \(> 30°C\).

Time-series information on the seasonal evolution of SST over the entire US JGOFS study area are computed by extracting a mean SST at each standard ASPS station position. The results of performing a PCA on these time-series for 1995–1996 are presented in Fig. 3. The amplitude of EOF Mode 1 is displayed at the top of the
Fig. 2. Satellite-derived mean SST (°C) during SW Monsoon of 1995.

The figure and the spatial distribution of Mode 1 is displayed at the bottom of the figure. (Remember, to retrieve the actual SST variability at a particular station, it is necessary to multiply the amplitude by the magnitude of the spatial distribution at a particular station.) Rather than the typical seasonal variability expected at mid-latitudes, the amplitude of the SST shows a biannual distribution because of the effects of 'Arabian Sea Cooling'. The heavy dashed line on this figure is a sketch of what might be thought of as the “normal” seasonal variability in SST variation where the SST would reach a maximum during the summer and a minimum in the winter. The amplitude of the cooling signal in the SW Monsoon is on the same order as the seasonal heating of
the sea surface that occurs prior to the onset of the monsoon. The spatial distribution of Mode 1 shows that in general the magnitude of the cooling signal decreases the further the station is from the upwelling area along the coast of Oman (the spurious peaks in the spatial distribution are associated with mesoscale eddies that lie along the US JGOFS cruise track). While most of the signal in the spatial distribution is within 600 km of the coast, the signal is still seen as far out in the basin as the US JGOFS data were taken. At station S13, the signal is still obvious, but the magnitude has been reduced by 50%. Also note that in the plot of spatial distribution, a decrease is seen at
the stations closest to the coast (N1 and S1; Fig. 1). This appears to be associated with a narrow southward flowing surface current that is carrying water of Persian Gulf origin southward (Morrison et al., 1998). Finally, the spatial distribution shows that the cool upwelling waters penetrate much further into the basin along the southern line than the northern line. This is especially true if one ignores the “peak” at station N5, which is associated with the presence of a mesoscale eddy (Manghnani et al., 1998; Böhm et al., 1999).

Fig. 4 shows the mean SST during the SW Monsoon of 1995 at Section U4. The mean SST increases gradually from approximately 22.5°C at the coast to about 26.5°C approximately 120 km off the coast, defining the offshore extent of the mean upwelling zone. Further from the coast, in the central basin, SST is fairly constant. It is interesting that the SST only shows a surface expression of the upwelling to about 120 km from the coast, while the subsurface temperature field showed effects of upwelling to approximately 260 km.

The standard deviation (SDV) about the mean SST during the SW Monsoon of 1995 (not shown) gives a qualitative estimate of the strength of the upwelling. The lowest SDV (< 3°C) is observed between the two capes along the coast of Oman, indicating strong and persistent upwelling in that area during the entire SW Monsoon. Along the southern coast of Oman in Salalah Bay, the SDV increases to > 3°C, indicating a more variable SST and therefore variable upwelling in this region.

Fig. 5 shows the anomaly of sea-surface geostrophic flow computed from the 0.25°, objectively mapped T/P SLAs and the relevant mean SST distribution for the period when the SW Monsoon was spinning up (mid-June 1995) and a period when the SW

![Graph: Mean SST in the Summer of 1995 along U4]

Fig. 4. Satellite-derived mean SST (°C) along section U4 during SW Monsoon of 1995.
Monsoon was at its peak (early August 1995). During mid-June in 1995, an alongshore current flows eastward between 54°E and 56°E, then around the cool-core Socotra Eddy at 56°E, turns onshore, and then flows northeastward along the Oman coast. A large anti-cyclonic eddy is observed at 17.5°N, 58.5°E off Ras Marbat. Manghmani et al. (1998) discuss the reasons for the presence of this eddy and hypothesize that it is responsible for “pulling” a filament of freshly upwelled water from the main upwelling region into the central basin (Brink et al., 1998). The alongshore current seen near the coast moves offshore at Ras Madraka. The 25°C SST isotherm for mid-June suggests a similar circulation pattern as seen in the surface currents.

During early August (Fig. 5), the circulation pattern shifted southward. The Socotra Eddy has shifted south of 15°N, and the eddy seen off Ras Marbat is now centered at about 16°N. The northeastward flow along the coast of Oman has greatly weakened. In addition, looking at the 25°C isotherm, the cold filament is no longer observed off Ras Madraka.

As shown above, the SST distribution can be used to “define” the position of the coastal upwelling front. Fig. 6 shows the offshore distance of the upwelling front at Section U4 during the SW Monsoon in 1995. In the mid-June, the upwelling front is near the coast, moving farther offshore until it reaches 120 km offshore in early
August as the monsoon progresses. With the collapse of SW Monsoon and the weakening of the upwelling, the front becomes indistinct in late August.

In both of the geostrophic flow fields during mid-June and early August (Fig. 5), several eddies are located off the Arabian coast. Brink et al. (1998) suggest that it is difficult for the wind regime to produce an organized upwelling circulation offshore and that any coherent offshore upwelling northwest of the Findlater Jet quickly becomes unstable and collapses to an eddy field.

Time-series of the SST and wind stress for locations within the upwelling zone are presented in Fig. 7. At stations S2 and S3, along the ASPS southern line (see Fig. 1) there is a dramatic and almost instantaneous decrease in SST with the onset of the SW Monsoon. At station S2, the SST decreases 5°C from above 28.5 to 23.5°C within a period of 20 days. Temperatures continue to decrease until a minimum (< 21°C) is reached in late August to early September, while at both stations, the strongest wind stress of 0.25 Pa (i.e., the peak of the SW Monsoon) is reached in mid-July. Therefore, as the SW Monsoon progresses and relatively warmer surface waters have been advected out the region, they have been replaced with upwelled waters from deeper in the water column. Waters with temperatures of approximately 21°C are upwelled from approximately 100–150 m (Fig. 8). Fig. 8 is a vertical section of temperature along the southern US JGOFS track during TN049 in August 1995. The upwelling and associated alongshore geostrophic current is obvious with the ascending isotherms inshore of station S2. In addition, this section crosses the upwelling filament at approximately 500 km (station S4) off the coast. The coldest upwelled waters at stations S2 and S3 come from approximately 100–150 m.

The mean SST for the SW Monsoon (Fig. 2) shows that the most persistent upwelling is confined to the continental shelf in the bays along the coast of Oman. The
time-series of SST constructed for pseudostations located in 2 of these bays are presented in Fig. 7. The SST variations at pseudostations P1 and P2 are considerably different from the observed at stations S2 and S3. Within a short period after the onset of SW Monsoon in mid-May, SST decreases to below 20°C and remains there throughout the entire SW Monsoon. The period of lowest temperature lags about one month behind the period of maximum winds at stations S2 and S3, but rather than briefly reaching a minimum temperature, the minimum temperatures remain relatively constant for a period of greater than 3 months. Another important difference is that
at stations S2 and S3, SST begins to increase dramatically with the collapse of SW Monsoon to temperatures > 27°C in late September, while at stations P1 and P2, the SST does not begin to increase until approximately one month after the collapse of the SW Monsoon.

3.2. Sea surface height and transport

The mean SLA for the upwelling region for 1993–1995 is presented in Fig. 9. The mean SLA in the upwelling area shows a periodical change, reaching a maximum at the onset of the SW Monsoon in late May to early June in all three years. The mean SLA then decreases dramatically, up to 30 cm, during the SW Monsoon, reaching a minimum of approximately −10 cm during late August. This decrease in steric height is atypical; normally steric height increases as the summer heating season progresses and does not begin to decrease until the fall. Here we see a significant decrease in steric height associated with “Arabian Sea Cooling” (Düing and Leetmaa, 1980). After the SW Monsoon, the SLAs begin to increase, reaching values of 10 cm to −5 cm during the height of the NE Monsoon. Morrison et al. (1998) show that, on average, SST has its warmest values during the intermonsoon periods. There is then a slight decrease in SLA with the collapse of the NE Monsoon. This might be attributed to the winter cooling in the upwelling zone which consequently decreases the sea temperature and thus leads to the decrease of the sea-surface steric height.

Due to geostrophic adjustment caused by the decrease in sea level along the coast associated with the upwelling during the SW Monsoon, a northeastward alongshore current (Arabian Coastal Current) should occur along the coast or Oman. Since the
Fig. 9. Time series of spatially averaged SLA (cm) within the upwelling region for 1993–1995. The mean wind stress for the upwelling region is presented to allow the reader to more easily reference the changes in geostrophic and Ekman transports to changes in the monsoonal winds.

Alongshore current along the Oman coast is relatively weak and narrow, with eddies dominating the area offshore of the main upwelling area, it is difficult to infer a significant alongshore current in the SST fields, although such currents can be observed in other western boundary current regimes such as the Gulf Stream. We do see the result of this flow in the jet that leaves the coast at Ras al Hadd (Böhm et al., 1999). The T/P data provide us a useful tool to quantify the flow in the upwelling area.

The mean SLA variations along Section U3 for 1993, 1994 and 1995 are shown in Fig. 10. Since the SLAs are averaged over the whole summer (9 T/P repeats), the error associated with each individual point along the T/P track is decreased to 0.4 cm. For the summer of 1993 and 1994 and 1995, the SLAs gradually increase seaward. This sea-surface slope drives a mean northeastward, alongshore current during the SW Monsoon. For the summers of 1993 and 1994, the northward alongshore currents are confined within an offshore distance of 150 km, while during the summer of 1995, the alongshore current extends over 250 km offshore, but the flow is weaker (SLA slope is not as steep as in the other two years).

The time-series of mean flow through the upwelling region for 1993 through 1995 is shown in Fig. 11. The transport has been computed every 10 days (T/P repeat period). It is important to note that when the winds change sign, the curves for geostrophic and Ekman transports cross at close to the zero transport line. This systems that the actual mean that has been removed from the geostrophic velocity is approximately
zero and that our assumption that the anomalies of surface geostrophic velocities computed from the altimeter data are actually quite close to the actual geostrophic surface velocity. Therefore, these plots are in the first approximation representative of the actual geostrophic transport. In addition, it is assumed that the total flow is represented by the sum of the geostrophic flow and the Ekman flow (see similar argument in Shi et al., 1999). An SLA of only 1.5 cm would result in an error in computing the geostrophic transport of $0.3 \times 10^6$ m$^3$ s$^{-1}$.

The net geostrophic transport perpendicular to the upwelling box is displayed in Fig. 11. The total transport per unit width thus becomes a sum of the geostrophic and Ekman transports. Northeastward geostrophic flow along the coast through sections U1 (Fig. 12a) and U3 (Fig. 12c) occurs during the SW Monsoon (Arabian Coastal Current). This current is associated with the geostrophic adjustment caused by the offshore pressure gradient established by the offshore Ekman flow. The Ekman transport through sections U1 and U3 is negligible, which is to be expected since the winds are parallel to the coast. Across section U1 during the SW Monsoon, the geostrophic transport reaches $11 \times 10^6$ m$^3$ s$^{-1}$ in 1993, $7 \times 10^6$ m$^3$ s$^{-1}$ in 1994, and $4 \times 10^6$ m$^3$ s$^{-1}$ in 1995. The flow through section U1 exits the region by monsoonal forcing in the upwelling zone and is presumably one of the main sources of water for the Ras al Hadd Jet (Böhm et al., 1999).

The net transport through section U2 parallel to the coast (2 in Fig. 11) represents the net flow of upwelled water from the coastal region into the central basin. The observed flow through this section is quite variable, most likely because offshore flow from the upwelling zone occurs in squirts and jets (Brink et al., 1998). The Ekman
Fig. 11. The net transport through each section within upper 100 m: (1) northeastward out of the upwelling box across section U1; (2) onshore into the upwelling box across section U2; (3) northeastward into the upwelling box across section U3; and (4) into the upwelling box across U1, U2 and U3. The mean wind stress for the upwelling region is presented in (5) to allow the reader to more easily reference the changes in geostrophic and Ekman transports to changes in the monsoonal winds.

transport across section U2 does show a periodical change with the seasonal monsoon: strong advection offshore during the SW Monsoons and weak advection onshore during the NE Monsoons. During the SW Monsoon, the Ekman transport out of the upwelling box has a maximum of $6 \times 10^6$ m$^3$ s$^{-1}$, while the Ekman
Fig. 12. The mean wind stress (arrows, Pa) and wind stress curl (green contours, $10^{-8}$ dyn cm$^{-3}$) during the summers of: (a) 1993, (b) 1994 and (c) 1995.
transport into the upwelling box has a maximum of $2 \times 10^6 \text{ m}^3 \text{ s}^{-1}$ during the NE Monsoon. Even though the Ekman transport has a smaller magnitude than the geostrophic transport across this section, the stability of the Ekman flow direction makes it important in determining the volume budget across the upwelling box.

At section U3, the Ekman transport is negligible, with the total transport being dominated by the geostrophic transport (3 in Fig. 11). Across section U3 during the SW Monsoon, the geostrophic transport reaches $4 \times 10^6 \text{ m}^3 \text{ s}^{-1}$ in 1993, $7 \times 10^6 \text{ m}^3 \text{ s}^{-1}$ in 1994, and $10 \times 10^6 \text{ m}^3 \text{ s}^{-1}$ in 1995. This is interesting because it shows that at the “entrance” to the upwelling box (U3) the interannual variability of the transport in the East Arabian Current (EAC) shows an increase from 1993 to 1995, while at the “exit” from the upwelling box the interannual variability in the EAC shows a decrease.

At section U4, the mean transport shows weak outflow from the upwelling box into the central basin during the SW Monsoon, reaching approximately $3 \times 10^6 \text{ m}^3 \text{ s}^{-1}$, while there is weak inflow into the upwelling box during the NE Monsoon, reaching approximately $3 \times 10^6 \text{ m}^3 \text{ s}^{-1}$. The total transport into the upwelling region reflects the volume exchanges across the upwelling region. In this figure, the geostrophic adjustment process is shown clearly in the three years. The Ekman transport out of the upwelling box always exceeds the geostrophic transport into the box, indicating that the water is upwelled from the deeper layer to compensate the loss.

We also can use the SLAs to try to estimate the amount of water that is upwelled within the upwelling box and the upwelling velocity. Here, rather than using the mean
Table 1
Computed upwelling parameters

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<th>1993</th>
<th>1994</th>
<th>1995</th>
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<tr>
<td>Mean upwelling volume transport</td>
<td>$2.2 \times 10^6 \text{ m}^3 \text{ s}^{-1}$</td>
<td>$1.4 \times 10^6 \text{ m}^3 \text{ s}^{-1}$</td>
<td>$0.6 \times 10^6 \text{ m}^3 \text{ s}^{-1}$</td>
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<tr>
<td>Maximum upwelling volume transport</td>
<td>$3.1 \times 10^6 \text{ m}^3 \text{ s}^{-1}$</td>
<td>$3.5 \times 10^6 \text{ m}^3 \text{ s}^{-1}$</td>
<td>$1.7 \times 10^6 \text{ m}^3 \text{ s}^{-1}$</td>
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<tr>
<td>Mean upwelling velocity at 100 m</td>
<td>$1.3 \times 10^{-5} \text{ m s}^{-1}$</td>
<td>$0.9 \times 10^{-5} \text{ m s}^{-1}$</td>
<td>$0.4 \times 10^{-5} \text{ m s}^{-1}$</td>
</tr>
<tr>
<td>Maximum upwelling velocity at 100 m</td>
<td>$2.0 \times 10^{-5} \text{ m s}^{-1}$</td>
<td>$2.3 \times 10^{-5} \text{ m s}^{-1}$</td>
<td>$1.1 \times 10^{-5} \text{ m s}^{-1}$</td>
</tr>
<tr>
<td>Mean wind stress curl</td>
<td>$2.0 \times 10^{-8} \text{ dyn cm}^{-3}$</td>
<td>$1.9 \times 10^{-8} \text{ dyn cm}^{-3}$</td>
<td>$1.4 \times 10^{-8} \text{ dyn cm}^{-3}$</td>
</tr>
</tbody>
</table>

slope of the sea surface along the boundaries of the upwelling box, we use the SLAs to compute the surface geostrophic velocity. As was done by Shi et al. (1999), we use the “mean” geostrophic shear within the upwelling box computed from the ASPS hydrographic data. The surface geostrophic velocities computed from T/P data are integrated over the upper 100 m using the mean geostrophic shear to estimate the geostrophic transport within the surface layer. The 100 m depth was chosen to include the major portion of the water column where the isotherms are observed to rise in the upwelling region (Fig. 8). We are able to compute time-series of mean geostrophic transports within the upper 100 m for sections U1, U2, U3 and U4 (not shown here as the resulting time-series display similar features to those shown in Fig. 11, with smaller magnitudes). The flow though U4 would reflect the volume exchanges across the upper 100 m of the upwelling box. Using this information, various parameters associated with the upwelling in the SW Monsoon can be computed (see Table 1). In addition to the upwelling parameters, the mean wind-stress curl over the upwelling box also was computed (see Table 1). Upwelling velocities are O ($1\text{ to }2 \times 10^{-5} \text{ m}^3 \text{ s}^{-1}$) while the maximum upwelling transport is O ($0.5\text{ to }2 \times 10^6 \text{ m}^3 \text{ s}^{-1}$) (Table 1). In 1995, the upwelling is weaker than the other two years, with an average upwelling volume at the depth of 100 m of only $0.55 \times 10^6 \text{ m}^3 \text{ s}^{-1}$, while in the other two years the magnitudes are 3 to 4 times larger. The same trend is observed in the magnitude of the wind stress curl over the upwelling box, where it decreases from $2.0 \times 10^{-8} \text{ dyn cm}^{-3}$ in the summer of 1993 to $1.9 \times 10^{-8} \text{ dyn cm}^{-3}$ in the summer of 1994 to $1.4 \times 10^{-8} \text{ dyn cm}^{-3}$ in the summer of 1995. Fig. 12 shows the mean wind stress and curl for the three years for the eastern Arabian Sea. The vortex field along the Arabian coast is weaker in 1995 than in the other two years.

4. Summary and discussion

The following upwelling characteristics are identified for the Oman upwelling zone:

1. Strong and persistent upwelling is confined to the continental shelf within the two bays between Ras Madraka and Ras Marbat. The mean SST decreases to 23°C
within the two bays in 1995, indicating that the water has upwelled from approximately 100–150 m.

2. The cold upwelling water persists in the coastal bays for approximately one month after the collapse of upwelling-favorable SW winds, with cool SST observed until late October. Offshore of the 1000 m isobath, SST increases rapidly with the collapse of the SW Monsoon.

3. The upwelling front, defined by SST, gradually moves offshore with the development of the SW Monsoon, reaching a maximum distance of approximately 100 km. The SST pattern near the coast is not only determined by the upwelling but also by the flow distribution in the upwelling region.

4. The SST variation in the US JGOFS study area is determined by the first EOF mode that captures 67% of the total variance. Rather than the typical seasonal variability expected at the mid-latitude, the time-series variation of the first mode shows a biannual distribution because of the effects of “Arabian Sea Cooling”.

5. The seasonally reversing wind fields associated with the monsoon result in upwelling and a compensating, northeastward-flowing East Arabian Current during the SW Monsoon and in downwelling and a compensating, southwestward-flowing East Arabian Current during the NE Monsoon.

6. The mean sea level drops significantly in the upwelling area (up to 30 cm) with the onset of the SW Monsoon. This signal is representative of the phenomena known as “Arabian Sea Cooling”.


8. The mean upwelling volume transports at the depth of 100 m within an area encompassing the region 260 km offshore and 600 km alongshore are estimated to be $2.2 \times 10^6$, $1.4 \times 10^6$ and $0.55 \times 10^6$ m$^3$ s$^{-1}$ for the summers of 1993, 1994 and 1993. The mean upwelling velocities through the 100 m level within the upwelling zone are estimated to be $2.0 \times 10^{-5}$, $2.3 \times 10^{-5}$ and $1.1 \times 10^{-5}$ m s$^{-1}$ for the summers of 1993, 1994 and 1995. The weaker upwelling in the summer of 1995 is attributed to the weaker wind stress curl along the Oman coast.

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**References**


